Numerical Study of an Observed Orographic Mesoscale Convective System.  
Part 2: Analysis of Governing Dynamics

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ABSTRACT

A detailed analysis of the dynamics and thermodynamics responsible for the structure, growth and propagation of an orogenic mesoscale convective system simulated in two dimensions is made. The process of scale interaction is addressed through Fourier analysis and Reynolds averaging analysis of representative predicted variables, diabatic forcing and momentum acceleration terms. Additional dynamical analysis is accomplished through sensitivity experiments in which Coriolis, diabatic heating and ambient airflow are varied.

The general conclusion is that the simulated orographic development is a geostrophic adjustment process to convective heating which is itself modulated and maintained by orographically induced flow systems. The heating scales range over a nearly continuous spectrum ranging from 10-250 km. The heating occurs in response to primary advective gravity modes. The larger-scale gravity-wave disturbances modulate the smaller scales by organizing mean upward vertical motion patterns. The largest gravity-wave modes are modulated by constraints of the slope flow circulation, namely a phasing of an advective mode with a localized break in the plains inversion.

The simulated growth to meso-α-scale proportions occurs from the horizontal expansion of the disturbance through interaction with the mountain–plains scale slope flow circulation. Similar to upscale two-dimensional turbulence cascade, the mountain plains solenoid deforms thermal patterns, increasing their scale. As the scale reaches meso-α-scale proportions, geostrophic adjustment frequencies are sufficient to allow the thermal fields to persist. Implications to the problem of cumulus parameterization and limitations of the two-dimensional framework of this numerical study are discussed.

1. Introduction

The governing dynamics of organized convection encompass interactions among a multitude of atmospheric scales of motion. The genesis process, in particular, has been a subject of considerable attention over the past few decades since its understanding is critical to the predictability of such systems. Because the origins of organized convection can be with small-scale transient motion (i.e., gravity-wave dominated dynamics), direct observation of initial atmospheric heating response is difficult. Moreover, because development is often largely upscale, attempts to model it with nonlinear meso-α-scale hydrostatic models are biased through assumptions concerning relationships between the smaller unresolved scales and those that are modeled explicitly. There is good reason to believe that this relationship, which is the crux of cumulus parameterization theory, may not be as well understood as many have been led to believe.

Much of our understanding of the growth process has come from simple linear models that depict growth of various spectral modes and thus are capable of selecting scales of development. Even these models, however, are strongly limited by their lack of predicted interaction among scales beyond the assumptions that are also necessary to the nonlinear meso-α-scale models.

In Tripoli and Cotton (1988; hereafter referred to as Part 1), we described the numerical simulation of the genesis of an observed orogenic1 MCC. An observational study of the early stages of mountain convection for this case of 4 August 1977 was presented in Cotton et al. (1983) and of the mature MCC stage by Wetzel et al. (1983) and McAnelly and Cotton (1986). In order to avoid the pitfall of using a cumulus parameterization to represent convection, the simulation was performed in two-dimensions with explicitly simulated convection. The early stages of the observed system were of linear geometry oriented parallel to a nearly linear mountain barrier. Despite the shortcoming of two-di-

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1 We have used orographic to define a convective weather system that is excited by orographic flow systems but becomes self-sustaining.
mensional simulations, this case study provided an acceptable situation for using a two-dimensional approach to simulating cumulus-scale–mesoscale interactions.

The observational study of Cotton et al. (1983) of this case attached only minor significance to the mildly disturbed synoptic situation toward influencing development of the MCS. Instead, it was hypothesized that the MCS was orogenic or organized under the influence of the local mountain-valley wind system. To test this hypothesis, an idealization of this case study was possible where the influence of synoptic-scale forcing could be neglected and the conditions for growth could be generated locally within the numerical simulation.

We demonstrated in Part I that the local flow systems did, in fact, act to organize convection and create ultimately a mesoα-scale signature. Despite its idealization, the simulated growth and propagation compared remarkably well to observations and supported the hypothesis that the MCC was excited primarily by local wind systems. It was demonstrated that there were significant mesoα-, mesoβ-, and mesoγ-scale (see Part I for explanation) signatures to the system. In addition, there were important environmental features that evolved in response to the system, such as the plains inversion, which played a crucial role in development.

The goal of this paper is to increase our understanding of the controlling dynamics simulated by the numerical model through additional analysis and sensitivity experiments. We begin in section 2 by reviewing some pertinent aspects of geostrophic adjustment and gravity-wave dynamics, as we expect them to apply to our analysis. In the next section, we analyze the orogenic MCS simulated mature structure and some dynamical aspects of its evolution. In sections 4 and 5, we describe the analysis of moisture and momentum budgets to increase our understanding of forcing of the MCS. In section 6, we look at the sensitivity of the system dynamics to changes in the model’s diabatic heating scheme. This will help in the overall interpretation of the results of the earlier sections. In section 7, we will look at the model’s sensitivity to changes in the initial wind speed. This will especially help in the understanding of the system movement. In section 8 we discuss the implications of this study to the cumulus parameterization problem. In section 9 we discuss the three-dimensional aspects of the governing dynamics.

2. Atmospheric response to convective heating

It is well known that when a local heat (or mass) source is introduced in the atmosphere, the two primary forces governing the response are gravitational and inertial. If the gravitational response is stronger, the result is the formation of a gravity wave, which propagates the mass anomaly away, leaving the wind field largely unaffected. On the other hand, if inertial forces dominate, the result is geostrophic adjustment of the wind field to the perturbation, leaving the local mass field intact. Hence, longevity is associated with the geostrophic modes while we call the gravitational modes “transient.” We know that the type of response the atmosphere will make is dependent on the scale of the disturbance.

The scale at which the gravitational response approximately equals the inertial response is the Rossby radius of deformation defined by Rossby (1938) to be

$$ L_{Rh} = \frac{c_{gs}}{f} $$

where $f$ is the Coriolis parameter and $c_{gs}$ is a gravity-wave phase speed for the $k$th gravity-wave mode. It can be shown that

$$ c_{gs} \propto N H_k $$

where $N$ is the Brunt–Väisälä frequency, and $H_k$ is the scale depth of the mode $k$. Various types of vertical structure functions can be assumed, depending on the upper boundary condition.

It should be noted that the Brunt–Väisälä frequency is reduced in a saturated, cloudy atmosphere (Fraser et al. 1973; Lalas and Einaudi 1974; Durran and Klemp 1982). Thus, as an MCS forms a large-areal stratiform anvil cloud, $L_{Rh}$ reduces in magnitude as a consequence of a reduction in the magnitude of $N$.

Mesoscale convective systems, such as the one studied here, are particularly interesting because their various components exist at scales typically both larger and smaller than the Rossby radius. Hence, we expect to find transient disturbances on the smaller scales and long-lived geostrophic modes on the larger scales. The Rossby radius itself will vary with the depth of the disturbance, latitude, and the horizontal extent of the anvil cloud.

This poses the interesting prospect that a shallow dry mountain–plains solenoidal circulation may be geostrophic in nature while a deepened version of the same circulation may be gravitational. In this study, we are concerned primarily with the deep tropospheric circulations, forced by cumulonimbus convection. We have determined from our simulation results that gravity-waves associated with those modes move consistently at phase speeds slightly less than 30 m s$^{-1}$. Therefore, at 39°N, the Rossby radius for those deep modes is about 300 km.

We must also be aware of the existence of other vertical modes that are perhaps shallower. For instance, consider heating associated with the long-wave radiative cooling of the anvil, or increased heating near the tropopause where convective updrafts rapidly decelerate, transforming kinetic to potential and thermal energy. These shallower modes will likely become geostrophic at smaller horizontal scales and therefore be-
come less transient. Hence, the shallow temperature anomaly deposited near the tropopause by convection can persist, while deep subsidence forced laterally will become a transient circulation.

Other studies of convective disturbances have shown, usually in the tropical atmosphere, that the true Rossby radius must be calculated by taking into account local angular momentum (Schubert et al. 1980). This effectively modifies the denominator of (1), so that it is a function of absolute angular momentum rather than just \( f \). We then find in regions of higher absolute vorticity, a smaller Rossby radius and hence smaller geostrophic modes. As a result, an anticyclonic environment, such as that in the upper levels of an MCS, will expand the Rossby radius, while the reverse is true at low levels. Moreover, as an MCS develops, \( L_R \) can change locally. We have chosen to neglect this effect in our analysis, since we are neglecting any ambient vertical component of relative vorticity in our initial conditions. In addition, vorticity formed by geostrophic adjustment during the MCS genesis is small compared to \( f \).

Forced transient modes consist of two types. In a wave-CISK study, Raymond (1983, 1984) labels them to be either “advective” or “propagating” modes. The advective mode does not move relative to the wind and typifies a growing disturbance characterized by an updraft and downdraft couplet. Simply put, this circulation will continue to grow as long as the updraft is warmer than the downdraft at a given vertical level. In a stable atmosphere this will last for a finite time, and when the condition is no longer met, the advective mode breaks into propagating modes that expand radially from the parent advective mode as gravity waves.

When several horizontal wavelengths coexist, nonlinear dynamics exchange energy across horizontal modes. The point of breakdown for a given advective mode may then be altered or prevented by this interaction. The nonlinear exchanges can themselves be altered by inertial forces, which inhibit gravitational modes near \( L_R \). We view MCS genesis as a geostrophic adjustment process where these complex dynamics result in the final formation of a long-lived semigeostrophic disturbance. The following analysis is performed with the intention of supporting this concept.

3. Mature system structure and structural evolution

At 16 hours into the numerical simulation described in Part 1, the local time was 1200 MST and the first convection was beginning to take place in the higher mountains. By 2000 MST, the system approached the Colorado–Kansas border and reached its overall greatest strength, although the mesoscale circulation increased somewhat afterward. In this section, we will consider the system structure at the time of peak intensity and then look at its evolution.

a. Mature structure in physical space

The larger mesoß- and mesoα-scale fields at 2000 MST (produced with an 86 km running average of predicted fields), given in Fig. 1, depict the organization of the MCS. The system core, coincident with the primary rainshaft at 103°W, was typified by intense mesoscale uplifting, reaching values in excess of 0.5 m s\(^{-1}\) over horizontal averaging scales of 86 km. Peak mesoscale upward motion occurred near 8 km or 35 kPa pressure. [Note that this is similar to Lin’s (1986) composite analysis of MCCs.] This was a result of horizontally convergent flow over the entire layer below 8 km, even though flow reversal was only found in the lowest 2 km AGL. The reversal from easterly flow preceding the system to westerly flow behind the system at the surface was observed. As was pointed out earlier, the 86 km averaged westerly low-level flow behind the system was considerably slower than the system propagation. If the gust front is to be an important mechanism forcing system propagation in this simulation, it must occur on the mesoγ-scale. Divergent flow characterized the \( u \)-component of motion above the peak vertical velocity in the upper troposphere and lower stratosphere.

The upward branch of the circulation cell formed what appeared to be a “classic” MCC signature (Maddox 1981; Lin 1986). A warm \( \theta \) core has formed at 8 km MSL (35 kPa) with relative cooling below (from precipitation melting and evaporation) and above (from overshooting updrafts and long-wave radiational cloud top cooling). The vertically integrated effect of the core temperature regime produced a 0.08 kPa high pressure center at 10 km MSL and a 0.09 kPa low pressure center at 3.5 km MSL. At this time, the low-level cooling was weak enough so that relative low mesoscale pressure was still found at the surface. The surface low was rather temporary and was more an exception than a rule throughout the integration. Most often, a meso-high was found at the surface below the upbranch.

Geostrophic adjustment to the divergent flow aloft and convergent flow below led to anticyclonic shear of the \( v \)-component aloft and cyclonic shear below. At low-levels, the \( v \)-component accelerated geostrophically to a southerly direction east of the system and a northerly direction west of the system. Because of the slow geostrophic adjustment time and because the system was moving faster than the low-level flow, the formation of mesoscale northerly flow lagged the system by 50–100 km. This structure was also found in the composites of Maddox (1981). Maximum meridional flow aloft was separated by 600 km, which is on the order of the reference state Rossby radius of deformation.

At this time, the simulated mesoγ-scale vertical component of relative vorticity was about \(-23 \times 10^{-5} \text{ s}^{-1}\) at 200 mb and \(+0.6 \times 10^{-5} \text{ s}^{-1}\) at 700 mb. For a
mature MCC, Maddox (1981) found values averaging near \(-5 \times 10^{-5} \text{ s}^{-1}\) at 200 mb and \(1.5 \times 10^{-5} \text{ s}^{-1}\) at 700 mb. Lin's (1986) composite analysis of MCCs depicted relative vorticity of \(0.5 \times 10^{-5}\) at 700 mb and \(-2 \times 10^{-5}\) at 200 mb during the growing stage. Precise comparison of such features with a composite is not feasible, but these rough comparisons demonstrate that the simulation is producing a balanced response of the right order of magnitude.

The \(\theta'\) field featured a warm system core with the maximum temperature anomaly of 2–3 K at about 8 km MSL (approximately 35 kPa) coincident with the meso\(\beta\)-scale upward motion core. To the east of the system, the warm anomaly extended downward to 4 km MSL, and extended laterally at the 8 km MSL level through the anvil. Examination of heating effects, presented later, show that the eastward extension was principally caused by longwave radiation. An asymmetric distribution of deep subsidence warming about the core resulted from enhanced precipitation-induced cooling to the east. Precipitation cooling was enhanced to the east as a direct result of upper-level advection of ice crystals downwind from the up-branch. It was hypothesized in Part 1 that this virga played an important role in the systematic erosion of the plains inversion as the MCS advanced eastward.

Because a meso\(\beta\)-scale response is less transient than the convective-scale motions, it is these features that are likely to appear in a composite of an MCC based on a large number of independent observations. Figure 2 displays the MCC composite structure found by Maddox (1981). The similarity to the predicted MCS structure in Fig. 1 is striking. The composited temperature anomaly aloft of 2–3 K at 35 kPa, the \(\alpha\) outflow pattern and the accompanying high-pressure pattern at 25 kPa are all reproduced well in the simulation.

The simulated low pressure between 50 and 70 kPa was not found to be as strong by Maddox. More recent composites of a large set of observations of MCCs reported by Lin (1986) do reveal its existence.

Relatively deep, moist air west of the up-branch is another feature that exists both in the composite and the simulation. This results from the prior existence of convection moistening at midlevels west of the system. The strong moist anomaly of Maddox's composite near the surface is not reproduced in the simulation. This is apparently a feature of the southerly low-level jet (LLJ), not modeled because of the initial assumption
of horizontal homogeneity. The overall scale of the flow is nearly the same in the simulation and the composite observation, perhaps because it is controlled largely by the Rossby radius which varies little during the summer.

It was suggested in Part 1 that the maintenance of mesoβ-scale vertical motion east of the up-branch of the mountain–plains solenoid after dark may be a result of enhanced wave trapping. Fig. 3 clearly depicts the emitted internal gravity waves in the w-field and shows the collapse of the core upward motion regime at 2100 and the resurgence occurring by 2200 MST. Note the existence at 2200 MST of upward motion as far east as central Kansas and as far west as Denver that lingers from the intense stage at 2000 MST. A second minor intense stage around 2130 MST was collapsing at 2200 MST, and regions of upward motion were beginning to separate and move apart.

The mesoγ-scale structure of the core, displayed in Fig. 14 of Part 1, depicted the existence of a varying number of deep gravity-wave-like oscillations of wavelengths about 15–20 km and periods of 15–30 minutes. These were the cumulonimbus clouds themselves. The mesoβ-scale waves seemed to be emitted from this core region from a continuing vertical oscillation with a time period of 2 hours. The mesoβ-scale core behaved as a forced oscillation gaining energy from the release of convective available potential energy. The variations in intensity of the cumulonimbus to this oscillation suggested that their strength is closely tied to the mesoβ-scale support.

Trajectory analysis, shown in Part 1, suggested that parcels were routinely exchanged across the mesoβ-scale flow. This effectively produced an eddy transport of high moist static energy upward and low moist static energy downward across the mesoβ-scale easterly storm-relative flow. This was, in essence, the process one would try to represent with a cumulus parameterization.

These depictions suggest that an ongoing scale interaction whereby cumulonimbus, parasitic upon a mesoβ-scale circulation, transfer heat into that circu-
lation. The mesoβ-scale circulation was found to grow to a critical point, drifting with the mean wind, and then eventually break down, giving rise to oppositely propagating gravity waves. We believe that this critical point signals the end of an advective mode and the beginning of propagating modes analogous to those depicted in the wave-CISK model of Raymond (1983).

The critical point is reached when the forces accelerating the mesoβ-scale up-branch and down-branch reverse. This occurs when the relative buoyancy (including effects of condensate) within the core becomes less than the surrounding environment. Processes such as evaporational cooling, precipitation loading and environmental subsidence warming all contribute to this process.

A classic example of the growth/breakdown process is the Byers and Braham “airmass” thunderstorm model. The mature phase is the point of breakdown and it occurs when waterloading and entrainment initiate a downdraft. The downdraft grows and eventually consumes the entire cloud.

In the case of the simulated mesoβ-scale circulation, breakdown is triggered more from the thermal effects of subsidence and midlevel cooling. The massive breakdown in the suppression zone (on the west High Plains) may have been triggered when surface convergence was interrupted slightly by the increased relative easterly flow over the lower amplitude slope of the plains topography. This forces the up-branch to entrain slightly more air from the midlevel dry zone, which then cools the updraft as evaporation occurs. This decreases the surface pressure gradient, which weakens the low-level inflow, increases midlevel entrainment and so on, leading to the total collapse of the core.

Subsequent collapse of the core seems to occur in a periodic fashion. Sensitivity tests, performed later in this paper, suggest breakdowns are at least partially a consequence of inertial stability. The Earth’s Coriolis acceleration restricts the scale of divergent response to cumulus heating. This eventually leads to excessive warming of the immediate environment and entrainment of midlevel dry air into the updraft leading to
the breakdown. We will show that the breakdown process is reduced and unrestricted growth occurs when the Coriolis parameter is set to zero.

The breakdown of the advective phase is the birth of a propagating phase. In general, the simulation demonstrated that most of the gravity-wave amplitude in the troposphere is lost within one wavelength of horizontal propagation distance to vertical propagation. As Lin and Smith (1986) predict, it is apparent that the propagating mesoβ-scale mode is ineffective in initiating new convective cells. After sunset, this situation is shown to change as the mesoβ-scale propagating mode apparently becomes "trapped" within a wave duct. This will be addressed later in this paper.

The mesoβ-scale growth and breakdown process was quite similar to that ongoing on the mesoγ-scales (see Part 1); except in the latter case, the wavelengths were reduced one order of magnitude and the period of oscillation was closer to 30 minutes, similar to the time scale of the "airmass" thunderstorm. Detailed examination of the mesoγ-scale simulated structure reveals that there tended to be a single dominant cumulonimbus cell within the storm core. Surrounding that cell were several updrafts that decreased in strength with horizontal distance from the core updraft. As shown in Part 1, the surrounding updrafts appeared to be propagating away from the central dominant cell.

Figure 14 in Part 1 depicts the existence of one to eight mesoγ-scale wave nodes. When a smaller contour interval was chosen (not shown), more lower-amplitude wave nodes appeared farther away (horizontally) from the central updraft and phase lines, suggesting that some propagation into the stratosphere was visible. The fact that the mesoγ-scale waves seem to travel more wavelengths in the troposphere than do the mesoβ-scale waves might be simply a result of their shorter wavelength, which would result in a decreased vertical group velocity. It is also possible that the mesoγ-scale waves may be partially trapped by the thermal structure on that scale.

We find that only the central one to three updraft cells actually contained cloud water (not shown), while most of the propagating gravity waves were cloud free, suggesting they lack sufficient amplitude to lead to latent heat release. It also confirms that the propagating updrafts are actually a wave rather than a convective updraft. Examination of the concurrent equivalent potential temperature (θ_e) field (also not shown) reveal that only the central updraft cell actually taps the high θ_e air at the surface. The updraft cells to the east were shielded from the surface by the plains inversion while those to the west were cut off by the central updraft.

The question naturally arises: How was the core mesoγ-scale updraft cell produced in the first place? Examination of a sequence of plots of θ_e suggests that the dominant cumulonimbus updraft was initiated periodically when a parent core cell produced condensate that was advected downshear, fell or mixed downward into the low θ_e, midlevel air, produced a cool and water loaded downdraft and, finally, cut off the inflow to the
parent cell, causing it to break down. At the same
time, surface convergence-generated downshear of the
downdraft initiated a new "daughter" cell.

It appears that gravity waves were emitted as the
parent cell broke down and either passed through or
preceded the initiation of the daughter cells to the east.
Hence, consistent with many observational studies, the
simulation suggested that cumulonimbus cells were
generally initiated by low-level convergence produced by
a downdraft.

This is generally consistent with the observations that
showed no evidence of strong convection cells west of
another such cell. Caution must be taken, however,
accepting the dynamics of these two-dimensional
clouds as realistic. It is likely that the actual clouds
experienced a somewhat different life cycle in the three-
dimensional environment, although they did appear
to be multicellular. Further study is needed to learn
what differences there would be in the scale interaction
between the meso-y-scale and larger scales when a fully
three-dimensional framework is considered. What
seems to be representative about this simulation is that
the emission of gravity waves from the breakdown of
the convective cells was not likely responsible for ini-
tiating new convection, and a dominant cell formed at
the western extremity of the plains inversion by pre-
cipitation-induced circulations.

b. Structure in wave space

In order to better understand this scale-interaction
process, we examine the Fourier power spectra of sev-
eral of the modeled variables. In doing so we gain in-
sight into the scales of relative importance. We must,
however, keep in mind that we are dealing with neither
a periodic nor an infinite series of data points, and so
some of what we find is simply error in applying a
Fourier transform to a limited-area problem. Also, we
are working with a limited time and space resolution,
so we should expect a bias toward resolvable scales.
We believe that, in spite of these shortcomings, analysis
in wave space adds to our overall interpretation of the
results.

For the purpose of this analysis, we have determined
it most instructive to perform a one-dimensional Fou-
rier transform in the horizontal only. This is because
the overall wave structure changes rather dramatically
when moving from the troposphere to the high stability
regions of the stratosphere.

We employ the transform given by

\[ A = \sum_{l=0}^{NX/2} \left[ \hat{a}_l e^{i2\pi x/L_x} + \hat{b}_l e^{-i2\pi x/L_x} \right] \]  \hspace{1cm} (3)

where \( \hat{a}_l \) and \( \hat{b}_l \) are the Fourier coefficients, \( L_x \) and \( L_z \)
are the lengths of the domain in the \( x \) and \( z \) direction,
\( l \) is the zonal wavenumber ranging from 0 to \( NX/2 \)
where \( NX \) are the number of grid points in the \( x \)-di-
rection. The power in each wavenumber is then given by

\[ P = |\hat{A}_l|^2 = |\hat{a}_l|^2 + |\hat{b}_l|^2 \]  \hspace{1cm} (4)

By plotting the power spectrum \( \langle P \rangle \) against \( l \) on a
log-log scale, we can optimally view the contribution of
each scale to the total variance. In order to eliminate
noise, common to any Fourier transform of a limited
discrete series, we also apply a \((\frac{1}{4}-\frac{1}{2}-\frac{1}{4})\) Hanning filter
to the spectrum four times.

We select three time-dependent variables to describe
wave activity. First, anelastic vertical motion \( w \) is taken
to represent the activity of all dynamic meteorological
circulations, including turbulence, propagating gravity
modes and advective gravity modes. The elastic fluc-
tuations of \( w \) are filtered by computing the meridional
component of momentum vorticity and relaxing the
field to obtain the rotational \( u \)- and \( w \)-components.
The \( v \), or meridional component of the wind, is selected
to represent the intensity of the geostrophic circulation
resulting from the orogenic disturbance. The forcing of
a secondary vertical ageostrophic circulation by the
geostrophic fields will not be properly simulated in the
\( w \)-field since the effects of meridional temperature and
zonal wind gradients are not represented in the two-
dimensional framework. A meso-o-scale vertical cir-
culation is produced by the model, however, in direct
response to the heating on those scales. This thermally
forced circulation, in a three-dimensional framework,
would build gradients in the geostrophic mass field
normal to our two-dimensional plane, which could
perpetuate the vertical circulation as active thermal
forcing dissipates. Finally, the \( \theta_v \), or virtual tem-
perature, is used to represent the structure of the simu-
lated mass field. This field will represent mass variations
on all simulated scales. As with the \( v \)-component, we ex-
pect our partially represented geostrophic modes to
show up here.

Figure 4 depicts the power spectra of \( w \) and \( \theta_v \) at the
midtropospheric level of 6.3 km MSL. This level tends
to be near the level of maximum vertical motion as-
associated with deep convection and with the meso-o-
scale overturning of the mountain–plains solenoid.
We present the spectra as a function of time in order to
gain a historical perspective of the significance of the
spectral peaks.

Discussing first the \( w \)-spectra, we note that variance at
2000 MST is centered between wavelengths of 10
and 200 km, with peaks at 33 and 17 km as seen on a
more detailed plot not shown. The fact that these peaks
differ by a factor of 2 suggests that the 33 km peak
may be an aliased response to the same oscillation as
the 17 km peak within the limited analysis domain.
Our examination of the meso-o-scale fields in Fig. 15
of Part 1 suggest that these peaks reflect the presence
of individual cumulus clouds.
Fig. 4. Power spectrum [$\log\left(\log[P]ight)$] of $w$ and $\theta_v$ at 6.3 km MSL and $v$ at 10 km MSL, contoured as a function of time (ordinate) and log wavenumber (abscissa). Contour intervals 0.5. Abscissa runs from wavenumber 1 (500 km wavelength) to wavenumber 480 (2.2 km wavelength) and are labeled in wavelength for convenience.
These spectral peaks are also part of a general maximum centered between wavelengths of 10 and 200 km, but reflected in other wavelengths as well. From an evolutionary perspective, we note that this general maximum is the second major maximum of strong $w$-variance, with the first occurring at about 1500 MST. It was shown in Part 1 that this first growth regime was cut off by the suppression zone reached between 1600 and 1700 MST. It can be seen that the first growth regime lasted longer and had peak variances centered at the somewhat longer wavelengths of 30–100 km.

The two-hour period is of less amplitude but can also be identified. As the system underwent breakdown after 1600 MST, an interesting $w$-variance maxima was found at 1800 MST and at the large meso-$\beta$-scale of 150–250 km, and to a lesser extent in the smaller meso-$\beta$-scale range. This suggests that we were correct in Part 1 in inferring that the 1800 period of regrowth was led by the apparent large meso-$\beta$-scale oscillation. It is noteworthy that the meso-$\gamma$-scale motions did not respond strongly to the first resurgence. When it did not respond, the primary growth was shifted more strongly toward the large meso-$\beta$-scale than at other times when the meso-$\gamma$-scale response was strong. A large meso-$\beta$-scale resurgence was also apparent at 2130–2200 MST and 2300–2330 MST. This suggests that the oscillation period was slightly less than 2 hours after sunset. Nevertheless, the 2 hour period seems to be led in the $w$-variance by the large meso-$\beta$-scale oscillation even at the later times.

On smaller scales, the $w$-variance decreases strongly with wavenumber. This occurs as a result of the increasingly strong model filters (the eddy viscosity treatment and the fourth order horizontal smoother) applied to the fields, maximizing at the $2 \Delta \lambda$ wavelength.

Now we look at the $\theta_v$-variance. Unlike the $w$-variance, the $\theta_v$-variance is centered in the shortest meso-$\omega$- and longest meso-$\beta$-scale ranges of 150–500 km wavelength. The dual maxima are apparent again; however, they lag the $w$-maxima by about 1 hour at the 333 km wavelength and are coincident with the $w$-maxima by wavelengths in the meso-$\gamma$-scale range. This strongly suggests that the $\theta_v$-variance is working its way up the scale, within the two-dimensional flow beginning from the convective scale, and then accumulating at the large meso-$\beta$- and small meso-$\omega$-scales.

A stronger $\theta_v$ maxima at the meso-$\omega$-scale wavelengths after 2000 MST is evident. We attribute it to at least three processes. First, 2000 MST is just after sunset and, as we will show in the next section, this marks a time of increased radiation-induced cooling at the anvil top. Second, it will also be shown that resulting destabilization of the upper anvil traps the meso-$\beta$-scale waves beneath the anvil. This results in increased ice content within the anvil and a secondary temperature maxima and minima associated with the wave. Hence, the wave-trapping may be causing an increased consolidation of the temperature field on the anvil scale.

The 2-hour period is somewhat less visible in the $\theta_v$ fields. Nevertheless, the 1600 and 2100 MST maxima are traceable to upscale traveling maxima originating with the meso-$\gamma$- and meso-$\beta$-scales.

For the $v$-variance, we have chosen the 10 km level because it is the region of anticyclonic outflow and the region where we found the two-dimensional analog of a jet streak to be produced. The $v$-variance spectra, also shown in Fig. 4, shows a dominance on scales of 333–500 km or greater than the Rossby radius. The dual maxima in time and the 2 hour oscillation period is not evident in this field. The $v$-variance seems to grow more steadily in time until 2100 MST and then weaken little thereafter. The $v$-scales seem to parallel the $\theta_v$-scales of variance accumulation, suggesting the tendency for a geostrophic balance between $v$ and the mass field represented by $\theta_v$ in the zonal direction. As we noted, the lack of a corresponding $w$-variance accumulation may simply reflect the limitations of a two-dimensional framework.

Because we have suggested a major role by the Coriolis accelerations in determining these scales, it is interesting to examine the consequences of simply setting the Coriolis force to zero. We did just that by repeating the control experiment exactly as in Part 1 but then removing Coriolis accelerations after 1200 MST. The system core movement was identical to the control; however, its overall intensity was considerably more steady. Its peak updrafts reached similar values, but the minimum peaks were less. In fact, as the system moved over the suppression zone between 1600 and 1700 MST, the large meso-$\beta$-scale advective mode did not break down immediately into propagating modes. Instead, the breakdown process was delayed until 1800 MST. When it did break down, the results were far less severe. Unlike the control case, meso-$\gamma$-scale convection continued through the breakup phase.

The $w$- and $\theta_v$-spectra are displayed in Fig. 5, for the no-Coriolis case. The $w$-spectra is dramatically changed. Most important is the fact that the temporal continuity and average intensity is far greater. There is far less evidence of a two-hour fluctuation. The broad spectral band of variance is centered around wavelength 50–100 km and seems to have less variance than the control at wavelengths less than 20 km. This suggests a lesser role of the meso-$\gamma$-scale when Coriolis accelerations are excluded. The $\theta_v$-spectra also has changed markedly, showing considerably more growth in the meso-$\omega$-scale range.

The circulation in real space at 2000 MST is displayed in Fig. 6. Comparing this with Fig. 1, it is readily apparent that a much stronger, deep, mountain–plains solenoid has formed in the case of infinite Rossby radius (i.e., $f = 0$). The scale has obviously exceeded that of the simulation domain, increasing the mean vertical motion. We can quantify this by viewing the
net average vertical motion across the domain produced by both experiments. Shown in Fig. 7, is the domain-scale mean vertical motion within the midtroposphere. In the control case there is a surge of such motion to over 0.03 m s\(^{-1}\) at 1600 MST, prior to the first breakdown of the 150–200 km meso\(\beta\)-scale circulation. This growth regime represents a net buoyant and uncompensated (by surrounding subsidence) upward acceleration of mass. The majority of the upbranch was compensated by surrounding subsidence, with the associated warming helping lead to the first breakdown. As the system underwent the meso\(\beta\)-scale breakdown phase over the suppression zone, the mean vertical motion weakened and grew slowly thereafter. The lack of strong growth in mean \(w\) at later times is particularly interesting since the maximum strength of the meso\(\beta\)-scale updraft was not achieved until 2000 MST. This suggests that all later surges are dynamically compensated and may even be dynamically forced by the gravitational oscillation.

In comparison, when the Coriolis parameter is turned off, the surge continues through the suppression zone, until 1800 when the growth weakens for about 2 hours and then returns as the sun sets and rich plains moisture is incorporated into the circulation. The delay in the onset of the initial surge beginning at 1400 MST of the control is attributed to the slower growth of slope flow within the no-Coriolis case. This results from the lack of super-geostrophic acceleration of slope flow in conjunction with residual nocturnally generated northerly surface flow on the eastern slopes. It is worth noting that the surge growth continued almost unbounded until 1800 MST, when it finally broke. It recovered quickly and resumed exponential growth. This can occur because the lack of any inertial stability allows the growth of a divergence field of unlimited horizontal scale. This supports the hypothesis that the meso\(\beta\) wave scale is related to the Rossby radius.

We now return to the control experiment to view the spectrum of variances in the lower stratosphere,
which receives energy primarily from lower levels. We have theorized that the gravity wave energy created in the troposphere as a result of convective overturning is projected into larger scales in the troposphere by the two-dimensional mountain–plains solenoid or is dissipated into the stratosphere. We have also made the observation, of the simulation, that perhaps some of the gravity-wave energy is trapped in the troposphere at times. A comparison of the tropospheric spectra to the stratospheric spectra may shed light on these subjects.

Figure 8 depicts the spectra of $w$ and $\theta_0$ at 14 km MSL. In the case of $w$-variance, we find a strong reflection of tropospheric variance especially in wave-lengths between 50 and 250 km before 1900 MST. After that time, the correlation lessens, suggesting a decoupling of the two layers after sunset. The $\theta_0$-variances show much more power in the stratosphere, from before sunset and increasing after dark. This is apparently due to the preferred heating at those levels by radiative processes at anvil top and the cooling from overshooting tops of cumulonimbis.

4. Diabatic forcing

Our scale analysis discussed in the previous section has posed some interesting questions concerning the system forcing. We saw a surprisingly “white” $w$-spectrum and stronger intensity on longer wavelengths in the $\theta_0$-spectrum. We ask, At what horizontal scales is the diabatic heating taking place, and what scales are being forced? We then must consider how this forcing evolves over time and correlates with the evolution described in the previous section.

If we neglect the heat storage and its effect on air heat capacity, consider only ice-phase thermodynamics and neglect precipitation fluxes, we can express the change in moist enthalpy as

$$dh = \rho_o c_p T d \ln \theta + \rho_o L_{iv} r_v + Q_r$$  \hspace{1cm} (5)

where $\rho_o$ is the mean density, $c_p$ the specific heat at constant pressure, $T$ the air temperature, $\theta$ the potential temperature, $L_{iv}$ the latent heat of sublimation, $r_v$ the vapor mixing ratio, and $Q_r$ is the diabatic heating by radiation. We will use the subscript $o$ to mean a horizontal domain meso-$\alpha$-scale average. Because of the importance of the ice phase in this study, latent heat of sublimation more appropriately depicts latent energy than simply latent heat of vaporization. We can decompose the moist static energy (or any spatially dependent variable) into perturbation scales by

$$h = \langle h \rangle + h^* = h_o + \langle h' \rangle + h'^*$$  \hspace{1cm} (6)

where the brackets are a meso-$\beta$-scale averaging operator (80 grid points in this application), $h'$ is the deviation of the meso-$\beta$-scale average from $h_o$ and $h'^*$ is the deviation of a local $h$ from $h_o + h'$.

Now, for anelastic motions, we can write

$$\frac{dh}{dt} = \frac{\partial h}{\partial t} + [wh]_z + H = 0,$$  \hspace{1cm} (7)

where $H$ represents horizontal and turbulent transport terms. We can then decompose Eq. (7) and express the local enthalpy tendency as

$$\frac{\partial \langle h \rangle}{\partial t} = -\langle \langle w' h'^* \rangle \rangle_z - \langle \langle w \rangle \langle h \rangle \rangle_z$$

$$- \langle \langle w \rangle h_o \rangle_z + \langle H \rangle + \langle Q \rangle.$$  \hspace{1cm} (8)

We can now use (6) to rewrite (8) as

$$\frac{\partial \langle h \rangle}{\partial t} = - \left[ \rho_o c_p \frac{T_o}{\theta_o} \langle w' \theta'^* \rangle + \rho_o L_{iv} \langle w' r_v' \rangle \right]_z$$

$$- \left[ \rho_o c_p \frac{T_o}{\theta_o} \langle w \rangle \langle \theta \rangle + \rho_o L_{iv} \langle w \rangle \langle r_v \rangle \right]_z$$

$$+ \langle Q_r \rangle - \left[ \rho_o c_p \frac{T_o}{\theta_o} \langle w \rangle + \rho_o L_{iv} \langle w \rangle r_v \right]_z + \langle H \rangle$$

$$= \langle Q_r \rangle + \langle Q_o \rangle + \langle Q \rangle + \langle A \rangle + \langle H \rangle.$$  \hspace{1cm} (9)

The first term on the right-hand side of (9), $\langle Q_r \rangle$, represents the transport of moist static enthalpy by small meso-$\beta$- and meso-$\gamma$-scale motions (which for these purposes we will refer to as convective scales,
which are in the form of perturbations relative to an 80 point or 1° longitude running average). The second term, $\langle Q_b \rangle$, represents a net transport by the 80 km scale motion itself, which we refer to as the convective cluster scale. Because the domain mean is removed, this term represents the correlations on the large meso-$\beta$ scale, resulting in a net heating of the domain scale. In effect, for those executing meso-$\beta$-scale models with 80–100 km scale grids, we can view the cluster scale as resolvable mesoscale heating and those of the convective scale as unresolved scale, requiring parameterization.

The third term on the right-hand side of (9), $\langle Q_r \rangle$, represents heating by radiation on the cluster scale. The fourth term, $\langle A \rangle$, represents a local uncorrelated change in enthalpy that does not provide any forcing to wave motion over the domain. The last term takes into account horizontal transport and diffusion. In fact, this is a rather important term that performs the majority of the actual scale interaction. Its action may be to spread a disturbance anchored relative to a flow horizontally, increasing its power in successively larger wavelengths. We, therefore, can view the first three terms as diabatic forcing terms to the disturbance, while nonlinear wave–wave interactions move the basic perturbation through the spectrum of waves. It will be instructive to examine the behavior of the first two terms and the radiation term, in order to understand the nature of the scales of diabatic forcing.

The spatial distribution of the diabatic $Q$ terms are displayed in Fig. 9. The $\langle Q_r \rangle$ heating rates up to 497 K day$^{-1}$ are calculated. The $\langle Q_b \rangle$ heating, on the other hand, is somewhat less (but still on the same order) at this time, reaching peak magnitudes of less than 100 K day$^{-1}$. Virtually all of the heating is concentrated in the vicinity of the system core, at vertical levels just below the tropopause.

The radiative-induced heating averages considerably less, reaching only 16–20 K day$^{-1}$. The scale is considerably larger, reaching meso-$\alpha$-scale proportions. Note that the down shear stratiform region is warmed from below and strongly cooled from above, primarily
from longwave divergences. This has an important destabilizing impact within the downstream anvil or stratiform layer.

This effect is depicted in Fig. 10, which shows the Brunt–Väisälä frequency at 2000 MST, the time of sunset. Note the region of nearly neutral stability stretching from the mesoscale system core eastward to the eastern model boundary. This can be compared with $\langle Q_r \rangle$ and Brunt–Väisälä frequency at earlier times before the sunset (not shown), which show a far more stable stratiform layer. The upper layers of the anvil are cooled strongly ($16 \text{ K day}^{-1}$) at night and weakly ($<4 \text{ K day}^{-1}$) during the day. The difference arises from the absorption of shortwave radiation during sunlight hours. We hypothesized that this upper tropospheric unstable layer creates a gravity-wave duct (Lindzen and Tung 1976), which may help trap tropospheric gravity-wave energy. This allows the kinetic energy released in the convective core to spread laterally. The vertical motion analysis mentioned in the previous section quite clearly shows this effect.

Reviewing Fig. 3, we see that the gravity wave fluctuations to the east of the core are stacked vertically and weaken dramatically above the anvil. To the west, as the fluctuations pass into the airflow without the radiatively induced wave duct, the waves tilt vertically upward toward the west and propagate to the model top.

Since the 2000 MST analysis time represents a period of rather unusual strength, it is desirable to find a longer term average convective heating rate. This can be obtained by time-averaging several of such computations over a period of varying system intensity. Figure 11 displays the vertical profiles of $\langle Q_r \rangle$ and $\langle Q_o \rangle$ for the system core. These depictions represent an average of five points in time, spaced by 15 minutes and centered at the 2000 MST analysis time. We find that the $\langle Q_o \rangle$ heating rate averages near the instantaneous value shown in Fig. 9, about 70 K day$^{-1}$. The $\langle Q_r \rangle$ is found to be much more transient in nature, as we would expect over the period of one hour, and to average about 100 K day$^{-1}$. This suggests that the growth scenario is characterized by the rise of a mesoscale motion regime (period of fluctuation of two hours) followed by the more rapid growth of a mesoscale component that grows to exceed the intensity of the supporting mesoscale component.

It is notable that the peak $\langle Q_r \rangle$ is near 7.5 km MSL, while the peak $\langle Q_o \rangle$ is found some 2 km higher. This is suggestive of the scale interaction process we envision whereby divergence of the flow is spreading the heat source in the anvil, emphasizing larger horizontal scales as altitude increases. This will be shown later also to occur with momentum transport.

We can also view these heating terms in Fourier wave space, as we did for several predicted variables in the previous section. This analysis will aid us in understanding which scales the diabatic heating most strongly forces directly through vertical transport. Since the
spectral analysis is in itself a scale analysis, we will not average the equations on the mesoB-scale, and instead, we will look at the co-spectrum of $w$ and $h$ given by

$$Q_i = \langle w'h' \rangle_z.$$  

Fig. 12 depicts the spectra of $Q_i$ at the upper tropospheric 8 and 10 km levels where we found local maximum of large mesoB- and mesosY-scale heating as a function of time. This presentation enables us to view the temporal evolution of the heating in addition to its structure at 2000 MST. Looking first at the 2000 MST period, we find an intense band of heat flux. The temporal fluctuation of the flux suggests that nearly all wavelengths greater than 5 km have a significant heat flux response, although the power is centered in the band of 5–125 km wavelength.

A notable feature of the spectrum at 2000 MST is that the flux at the 10 km level appears slightly more concentrated at the smaller mesoB-scales, having a lesser contribution at both the mesoA- and mesoY-scales than is found at the 8 km level.

The temporal variability of the heating strongly depicts the 2 hour oscillation of system intensity at the 10 km level, but less apparently at the 8 km level. We have identified the 2 hour oscillation to result from the periodic breakdown of the large mesoB-scale advective mode. We believe its stronger influence at higher levels reflects the fact that the depth of penetration of convection is seriously modulated by the degree of mesoB- and mesosA-scale support.

The 2 hour oscillation seems to have greater amplitude for shorter wavelengths earlier but to favor the large mesoB-scale wavelengths after dark. This results because of the dominance of mesoB-scale heating after dark and may partially reflect the passage of mesoB-scale updraft nodes through the domain boundary. Because there are only two to four nodes of the wave in the simulation domain at any given time, the passage of each node (at the mesoB-scale period of 2 hours)—through the lateral boundaries—could affect the forcing spectrum, which is linked to the upward motion of the wave. Before dark, we found these large mesoB-scale nodes to dissipate in the troposphere before reaching the boundaries.

As was suggested earlier, the 2 hour oscillation does seem to begin with the massive breakdown phase at 1630 MST. There is also some evidence that there may have been an earlier minor breakdown phase at 1400 MST, which we could not detect by viewing the predicted fields.

At both levels, the early growth phase at 1500 MST and the later phase at 2000 MST stand out. It is significant that the earlier growth phase forced features at the 10 km level far stronger than was the case at the 8 km level. This can be at least partially attributed to the fact that the system existed over somewhat higher topography at the earlier time. There may also have been a preference for deeper heating resulting from the existence of an unmodified sounding with a higher level of neutral buoyancy for a rising parcel.

5. Momentum budget

The processes directly responsible for the acceleration of air motion are now investigated. Acceleration of horizontal and vertical momentum occurs by pressure, Coriolis and transport processes. Our primary concern here will be to decipher the relative roles of these processes on the overall momentum budget.

The accelerations by vertical momentum transport, pressure gradient and Coriolis are displayed in Figs. 13 and 14 for $u$ and $v$, respectively, at the 2000 MST analysis time. The mesoY- and mesoB-scale accelerations are separated as in the previous section for moist static energy. In addition, local transport of the mean shear flow is displayed. From this analysis, it is strongly evident that the acceleration of zonal momentum is dominated locally by pressure acceleration that reaches peak magnitudes larger than transport or Coriolis by nearly a factor of 4.

The acceleration by vertical transport of mesoB- and mesoY-scale terms are strongest in the region of the system core. Because, as was found with the heat budget, the 2000 MST peak intensity period is anomalously strong, the 1 hour mean vertical transport of zonal momentum profile is displayed in Fig. 15. Note
that peak accelerations remain near 0.2 cm s\(^{-2}\). The maximum westerly acceleration is between 8 and 11 km MSL, which is the lower half of the upper-level westerly jet. Easterly acceleration is produced below. The vertical profiles of vertical zonal momentum flux (not shown) show that the accelerations are produced by a downward-directed, down-gradient momentum flux between 4 and 7 km MSL and an upward-directed, up-gradient momentum flux between 7 and 11 km MSL. The behavior of these fluxes indicate that the flow is tilted upshear with height below 7 km and downshear with height above 7 to 11 km MSL.

A comparison of the meso\(\beta\)- and meso\(\gamma\)-scale components reveal the transport is of comparable magnitude for both scales. As with the vertical heat transport, the meso\(\beta\)-scale component peak fluxes are shifted upward by approximately 1 km. This again seems to indicate a shift in energy from small to larger horizontal wavelengths in the generally stably stratified upper levels of the storm.

The effect of the Coriolis acceleration, which is nothing more than an inertial oscillation of the ageostrophic wind, is to oppose the pressure-induced acceleration aloft and near the surface. On the scale of the MCS core, the pressure gradient clearly dominates and a geostrophic balance is not evident. Beyond 200–300 km outside the convective core, the pressure gradient and Coriolis forces reach comparable magnitudes and appear to be somewhat compensatory. This is consistent with the scale of the Rossby radius of deformation, which is about 300 km for this case.

Terms affecting the acceleration of \(v\) are given in Fig. 14. In this case, local pressure gradient acceleration is neglected and Coriolis clearly dominates the acceleration on the domain scale. Locally, vertical transport is important, however. The primary impact of vertical transport seems to be the transport of relative southerly momentum from below into the upper storm levels. This implies a meso\(\beta\)- and meso\(\gamma\)-scale updraft sloped upward toward the north.
Vertical motion is affected primarily by local accelerations produced by an imbalance between buoyant motions associated with a local density anomaly and the local vertical gradient of perturbation pressure. The magnitude of the mesoβ-scale $w$-acceleration by hydrostatic imbalance is given in Fig. 16. It can be seen
that the $w$-acceleration is primarily at the base of the updraft, while the deceleration is aloft within the updraft.

This is typical of the growing phase of the MCS. During a weakening phase (not shown) the tendency is for deceleration at all levels with upward $w$-acceleration displaced laterally from the updraft where compensating subsidence was centered. This begins as the temperature of the subsidence regions increases to exceed that of the up-branch.

6. Sensitivity to diabatic forcing

In order to gain a full understanding of the relationship between forcing and growth of the MCS, we can study the sensitivity of the simulated structure to specific variations in forcing. In this section, we will do so by looking at development 1) in the absence of cloud, 2) in the absence of precipitation and 3) in the absence of cloud-induced radiative effects. If the reader is interested in the details of each experiment, their full description is included in Tripoli (1986). In this section we will summarize the salient features of each experiment.

a. Dry simulation

This simulation was not truly dry in the sense that water vapor and its virtual temperature effects were included. We simply disallowed any condensation or freezing to take place. In all other respects, the setup of this simulation was identical to the control.

The result of this experiment was that the meso-scale mountain–plains solenoid developed over the shallow layer 5 km above the ground but never deepened to the tropopause as in the control case. The mountain-wave convergence zone 60 km east of the Continental Divide formed as in the control case but, in the absence of convection, never propagated eastward. The dry solenoid simply accelerated its circulation as the daytime heating cycle reached its maximum and then decelerated afterward, in a stationary position. The flow above 5 km was largely unaffected.

The dry simulation, in some respects, resembles the mountain boundary-layer simulations performed by Banta (1986). In both this case and that of Banta (1986), the effect of latent heat is neglected. These results, however, are not consistent with Banta's findings that the mountain surface convergence line propagated eastward with time. There are several reasons why this simulation differs. The first is that the "dry" days which Banta studied were characterized by a deep, strong westerly flow extending to lower elevations than present in this case study day. That, in fact, is why the days were dry. As a result the low-level mean flow had a mean westerly component that even this analysis predicts should move.

Second, Banta concentrated on the boundary-layer evolution and used much finer horizontal and vertical grid resolution in addition to looking at smaller-scale topography. As a result, Banta studied local shallow slope flows that could be within a single vertical grid interval for this case. In fact, the regime Banta was studying would likely exist entirely between the ridge and the simulated mean convergence at 105.5°W in this case. The so-called valley breeze, or slope circulation on the scale of the entire mountain barrier, was not studied by Banta. The Banta-type solenoid may yet be another scale of circulation that occurs early in the morning. It is likely, however, that as a circulation of Banta's type reaches the valley breeze convergence line it will move no further and dissipate with the changing scale of governing dynamics.

The results certainly verify the notion that the cumulus convection directly leads to growth of the deepened mountain–plains solenoid. Moreover, it alerts us to the fact that the eastward movement of the up-branch is a direct result of deep moist convection. The following series of tests is designed to help us clarify what physical processes are necessary for that propagation.

b. No precipitation experiment

We ask here, of what importance is the precipitation process (as opposed to the latent heating process) to
the growth of the circulation? This is an important consideration toward our understanding of the integrated process of organization. Precipitation has the effect of not only unloading the weight of water from an updraft, but also of moving water mass into regions of relative low-valued $\theta$, air and evaporatively cooling those regions. This can create a greater stabilization and increase horizontal baroclinicity, which in turn can feed development. Without precipitation, the deep convection will be nearly wet adiabatic, rather than pseudoadiabatic. In effect, the deep convection will more nearly resemble deep dry thermals.

The experiment was run identical to the control, except that precipitation was not allowed to form. The only condensate permitted was cloud water, diagnosed to maintain 0% supersaturation. An additional effect of this restriction is that the latent heat of fusion is never realized, which further suppresses the virtual temperature warming within the updraft over the control.

The result of this sensitivity test was that the developing convection prior to 1600 MST was similar, except that the mesoscale thermal response was somewhat weaker aloft, attributable to reduced latent heating. Perhaps more important, stabilization west of the system near the surface was decreased, obviously from lack of precipitation. This had little effect early as the system continued to drift eastward at 10 m s$^{-1}$.

The major impact was identified at 1630 MST, as the suppression zone was reached. As with the control, the initial advective mode broke into oppositely propagating gravity modes. The propagating modes were again the 150–200 km gravity waves, which move at 30 m s$^{-1}$ relative to the flow. This time, however, the primary core never fully recovered from the breakdown phase. We attribute this to three factors. 1) The propagating mode moving westward encountered less (than the control) stabilized air and initiated new competing convection at the mountain-wave convergence zone. This helped prevent the vulnerable system core from reestablishing its link to surface convergence. 2) Whereas the control experiment was found to reestablish its link to surface convergence following breakdown via a precipitation-induced density current, the lack of any precipitation prevented that here. 3) The absence of precipitation prevents erosion of the inversion by virga.

All subsequent convection never attained sufficient strength to control the surface convergence field. Hence, the convection remained rather disorganized and never penetrated much farther eastward than the suppression zone.

c. The effect of cloud-induced longwave radiation heating

This section explores the sensitivity of the MCS growth, propagation and strength to longwave radiation effects induced by absorption and emission by condensate. To do so, an experiment was run where
the attenuation of longwave radiation by condensate was removed. This experiment is denoted NLWR. It was speculated, based on the analysis of the radiation-produced temperature tendency, that an important effect of radiation is to form an unstable layer near anvil top that acts to trap internal gravity waves emitted by the storm core. In addition, it may be possible that destabilization of the stratiform region by longwave divergence could lead to overall intensification of the core vertical motion as Chen (1987) found for another MCC case. These effects will be of particular interest in this section.

The simulations showed close similarity prior to sunset. The most noticeable differences were in the slightly weakened amplitude of the meso-β-scale tropospheric response in the NLWR case. This seemed especially true in the pressure field at 1600 MST whose amplitude decreased (from the control) by about 50% in the troposphere but increased similarly in the stratosphere. This suggests increased leakage of kinetic energy into the stratosphere. Perhaps, as a result, the breakdown within the suppression zone was somewhat less severe in the NLWR case.

The greatest impact was found after sunset. The explosive growth simulated in the control experiment near sunset (about 2000 MST) was less strong in the NLWR case. Perhaps this resulted, as with the first breakdown phase, because of less energy overall in the meso-β-scale advective and propagating modes. The pattern of a weaker 2 hour cycle became much more apparent after dark. Whereas the meso-β-scale propagating modes of the control case diffused the system core into several weaker up-branches after dark, the NLWR disturbance remained concentrated into a single core.

This effect is apparent in the w-field at 2200 MST displayed in Fig. 17, which can be directly compared to Fig. 3. Note the reduced extent to which downwind trapped waves are visible in the NLWR experiment. The hypothesis presented in the previous section that the nearly neutral layer within the stratiform anvil was formed by radiative effects is supported by Fig. 18, which depicts the Brunt-Väisälä frequency at 2000 MST for the NLWR case. Note that there is no evidence of this neutral layer. From these clear differences, we are confident that the increased tropospheric meso-β-scale vertical motion away from the system core is indeed a response to longwave radiation-induced trapping. It is important to point out, however, that the simulated extent of the anvil was somewhat large (see Part 1) and so the simulated lateral scale of trapping may be exaggerated.

We also point out that because of the prolonged consolidation of the system in the NLWR case, it becomes overall more intense than the control at 2200 MST, as measured by the amplitude of its spectral response.

It has been observed by the authors and others, during projects AIMCS and PRE-STORM, that a developing nocturnal MCS undergoes a transformation 1–
3 hours after sunset, wherein the core convection weakens and a more general anvil system evolves. These experiments may very well be demonstrating the mechanism causing such a transformation.

7. Sensitivity to ambient airflow

Thus far we have described the core of the MCS disturbance as moving with a mean tropospheric wind of about 10 m s$^{-1}$. One way to verify that this is indeed the case is to rerun the experiment with a lesser (or greater) wind and find if the storm movement adjusts as expected. On the other hand, our large meso-$\beta$-scale modes should continue to propagate relative to the flow at 30 m s$^{-1}$. Here we report on two experiments where we assumed first 50% and then 0% of the observed flow.

a. The 50% wind case

This case was run exactly as the control except that the wind was reduced by 50%. The primary difference was that the mountain-wave convergence zone was oriented about halfway between its control position and ridge top. Western slope advection of moisture was less, and as a result of both factors, only a single cloud formation zone formed. This was at the mountain-wave convergence zone, and it was delayed about 1 hour from the control.

As we expected, the system moved about half as fast eastward as the control, verifying our hypothesis that the storm movement is with the mean wind. The slower movement caused the storm to reach the suppression zone after 1800 MST, at which point it broke down similar to the control.

An interesting feature of this 50% wind experiment is an enhanced ability of some convection to leap ahead of the system core and form over the inversion. A plausible explanation might be that because the anvil does not extend quite as far ahead of the system, surface warming to the east is enhanced and increased boundary-layer development weakens the inversion. Also, the weaker mean wind at midelevels lessened the advection of the warmed mountain boundary-layer air over the plains and weakened the mountain-wave-induced subsidence, thus leading to a weaker plains inversion. As a result of an overall weaker inversion, internal gravity-wave motion induced by the system core has a greater chance of initiating some convection to the east of the MCS.

Because the anvil decreased in size as a result of lower winds aloft, the trapping of internal gravity waves remains closer to the core. In concert with this fact, it is found that propagating internal gravity-wave-induced vertical motion radiates strongly upward east of the anvil edge and does not affect regions as far east of the core as it did in the control simulation. This is additional evidence suggesting the importance of the trapped-wave effect.

It is perhaps of some significance that the MCS generated within the weaker airflow possessed somewhat less intensity measured by the number of vertical flow streamlines. This would likely be a result of a weaker supply of CAPE caused by a slower movement relative to the surface slope flow. It is significant that the system did not move faster relative to the mean wind. This suggests that the movement of this system was not a function of CAPE supply, but the intensity was.

b. No mean wind case

The case of no mean wind is interesting because there will be no external advection of the flow. Instead, the role of the internally generated forces dominate. Without this test, these forces may still exist but may be masked by the dynamic effects of the environment.

The experiment was initiated and integrated exactly as the control except that the initial wind was set to zero. The convection again formed initially at ridge top where the east and west slope flows collide. As the system grew, several breakdown phases occurred, forming the familiar 150–200 km propagating modes moving east and west from ridge top. Precipitation cooling beneath the updraft eventually forced twin surface convergence zones and up-branches on the east and west slope, which were nearly stationary by 1700 MST and separated by 200 km. Figure 19 depicts the flow system at 1900 MST, near its maximum strength.

An interesting feature of this case was the weakened nature of the plains inversion, apparently a result of the weaker convection and no mountain-wave-induced subsidence. Its effect was that the large meso-$\beta$-scale transients emitted by the system core were more apt to initiate weak convection to the east. Competition with the stationary core convection, however, still suppressed major development of this propagating mode.

8. Cumulus parameterization

Throughout our analysis we have found the predominance of the large meso-$\beta$-scale gravity wave, created from the combined heating effects of a multitude
of scales. Its growth is a CISK (conditional instability of the second kind) process, whereby heating on smaller scales forces the growth of the larger-scale disturbance. The unstable growth only occurred during the advective mode of the disturbance, and as propagating modes were created, the convective heating effectively decoupled. Growth was restricted to the narrow region at the western edge of the plains inversion. Hence there was no wave-CISK of the propagating mode as described by Raymond (1976).

The transients of all scales closely interact with each other, such that the vertical motion fields of the larger transients control the life cycle and intensity of the smaller ones. Therefore, the largest gravitational disturbance controls the overall action of all of those smaller. Our spectral analysis revealed no distinct spectral gaps in horizontal scale among the transients. In fact, dynamic growth was simultaneous among scales, while thermal variance worked upscale.

In order to simulate this process in three-dimensions in an economical way, it will be necessary to parameterize these scale interactions. Traditionally, we have assumed a distinct scale separation between the transients and the simulated motions. However, as our grid resolutions increase, we begin resolving the transients, which we find approach scales of 200 km. A typical numerical model will reasonably resolve the transient scale of 150 km, when the grid spacing is about 25–40 km. Then the convection is partially resolved and not decoupled temporally or spatially. What will happen if the scale separation assumptions are made anyway?

In an attempt to answer this question, we reran the control simulation with a grid spacing of 14 km, or 1/6° longitude spacing. This makes the minimum resolvable wavelength about 56–84 km. We then added a Fritsch and Chappell-type (1980a,b) cumulus parameterization to predict heating and moistening tendencies and eliminated our resolvable-scale microphysics prediction. It will be beyond the scope of this paper to present the modifications we made to the parameterization scheme; however, we note that the general concept remained the same. In particular, boundary-layer vertical motion fields trigger convection which then acts to stabilize a grid column over a period of time prescribed as a function of the wind speed.

After a great amount of experimentation, we could not reproduce the observed movement of the system as we did with the control experiment. Instead, we obtained the solution illustrated in Fig. 20. We display the time history of the vertical motion pattern and cloud outline. The growth of a pair of oppositely propagating large mesoβ-scale modes can clearly be seen. They are born from an advective mode that developed in the mountain-wave convergence zone, which then broke down into oppositely propagating modes. They moved at 30 m s⁻¹ relative to the wind, as they did in the control simulation. As they moved they grew stronger, essentially mimicking the propagating mode of a wave-CISK process.

In effect, the parameterization improperly coupled convection with the large mesoβ-scale wave. Considerable effort on our part to make the parameterization more sensitive to the inversion failed to resolve this problem. This was because, at that resolution, the systematic erosion of the inversion by smaller convective elements was not possible and a gradual eastward movement of the inversion-break could not be produced. The problem was probably exaggerated by the two-dimensional framework, which maintains a greater intensity of what would be a radially expanding mesoβ-scale gravity wave in three dimensions.

We also attempted this simulation on the same coarse grid, but with resolvable scale microphysics and without any parameterization. Development was delayed until very late, but when it happened, the disturbance remained decoupled from the propagating large mesoβ-scale gravity-wave modes. The timing of the system was poor, but the movement was good. Also, many aspects of the internal structure were not well simulated because of the course resolution. Forcing all modes to the resolvable scale produced unrealistic cloud dynamics. For instance, one overshooting downdraft actually warmed the surface 30°C.

FIG. 19. As in Fig. 1a, except for no wind experiment and at 1900 MST.

FIG. 20. Depiction of vertical motion and cloud boundary evolution for the case of parameterized convection. Light contour represents the 30 cm s⁻¹ upward vertical motion, while the heavy contour is the cloud outline at times indicated. The plot background is as in Fig. 1.
This exercise was important because it demonstrated a number of critical points. Most important is the fact that the simulated process of scale interaction is difficult to reproduce with a cumulus parameterization in the traditional framework. We are not suggesting that cumulus parameterizations absolutely cannot work, but we do believe there is reason for concern. At the same time, we feel that explicit microphysics on a large grid does poorly also, but in a different way. The correct representation of convection on a meso-$\beta$-scale grid may have to involve a higher order turbulence approach or perhaps a parameterization more responsive to the resolvable modes.

Another point is that this exercise may, in fact, explain why the wave-CISK hypothesis so often overpredicts the actual phase speed of convective disturbances such as squall lines. Nehrkorn (1986) demonstrated that the wave-CISK hypothesis predicted midlatitude squall lines to move about two to three times too fast. This was also the case with our fully nonlinear model when we introduced a cumulus parameterization. This suggests that a wave-CISK solution can be obtained erroneously by improper coupling of convection with propagating modes. In essence, the wave-CISK phase speeds may actually be reasonable if such coupling occurs. In our study, the coupling was simply prevented by the presence of a strong plains inversion and the action of other scales of motion, namely the meso-$\alpha$-scale mountain–plains circulation along with the environmental modification of earlier activity.

9. Conclusions

We have employed a nonhydrostatic primitive equation model in two dimensions to simulate the development of an orogonic mesoscale convective system. Our analysis of its structure revealed that it is characterized both by a spectrum of transient internal gravity-wave disturbances and a geostrophically balanced part at large scales.

The internal wave disturbances range from the scale of individual clouds to horizontal scales in excess of 200 km. The internal waves are of two types, namely, advective and propagating modes. The advective modes, similar in many respects to those described by Raymond (1983), drift with the wind and grow as long as the updraft remains warmer than the downdraft. They break into propagating modes, which expand outward and upward relative to the wind at internal gravity-wave phase speeds. The propagating modes can also be maintained or grow through continued latent heat release. Raymond (1983, 1984) has labeled the unstable growth of both propagating modes and advective modes as a wave-CISK process.

The growth upscale was helped by the convectively deepened two-dimensional mountain–plains solenoid, which restricts the region of convection to its up-branch and then spreads the resulting thermal fields to the east. There is considerable wave interaction among all scales of transient gravity-wave disturbances as well. The larger-scale transients modulate smaller ones so that, ultimately, the convective activity is modulated by the largest transient. That happens to be a large meso-$\beta$-scale of 150–250 km, which are the largest scales smaller than the Rossby radius of deformation. It reoccurs as an advective mode at a 2 hour period within the up-branch of the meso-$\alpha$-scale mountain–plains circulation, located at the western edge of the plains inversion.

Generally, the emitted, large meso-$\beta$-scale propagating mode does not excite convection itself due to unfavorable atmospheric conditions away from the meso-$\alpha$-scale up-branch and its great loss of energy to vertical propagation. After sunset, however, radiative destabilization of the downwind anvil partially ducts this wave energy beneath the anvil, increasing the amplitude of the propagating mode away from the primary up-branch. We speculate that this may increase the opportunity for growth of the propagating, large meso-$\beta$-scale mode after sunset in the region of the low-level jet. Indeed, there is evidence of this in the observations that depicted numerous meso-$\beta$-scale convection centers developing ahead of the main up-branch just after sunset (McAnelly and Cotton 1986).

Whether all of the dynamics simulated in the two-dimensional framework are representative of the observed development will remain a point of debate. Many of the observed facets of this system were properly reproduced by the model, but some were not. Nevertheless, this numerical investigation has provided new insight into the highly nonlinear scale interaction processes that occur during orographic MCS development.

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