Water and Energy Budgets of a Florida Mesoscale Convective System: A Combined Observational and Modeling Study

J. Halverson, M. Garstang,* J. Scala,† and W.-K. Tao*†

Universities Space Research Association Laboratory for Atmospheres, NASA/Goddard Space Flight Center, Greenbelt, Maryland

(Manuscript received 8 May 1995, in final form 13 November 1995)

ABSTRACT

Mesoscale water and energy budgets are diagnosed for a squall line during the Convection and Precipitation Electrification Experiment and combined with the results of the two-dimensional Goddard Cumulus Ensemble Model. The fine temporal and spatial resolution of cloud-scale processes contained in the model is used to reduce uncertainty in the diagnosed water budget residual and, thus, to arrive at a good estimate of storm-total rainfall. Profiles of cumulus heating (Qc) and drying (Qd) inferred from the sounding observations are in turn compared with the cloud-scale energy budget terms calculated from the model. This comparison reveals near-agreement in the magnitude and vertical distribution of the peak Qc and Qd, and also the relative size of the heating and drying at different levels in the column.

When the size of the mesoscale convective disturbance is approximately the same as the sounding observation network, it may be wrong to assume that the diagnosed vertical eddy heat transport accounts for most of the total eddy transport of moist static energy, F. The cloud model is used to resolve the relative contribution of the horizontal and vertical eddy flux convergence of heat and moisture, and thus it serves as a guide to interpreting the sounding-diagnosed total flux. The model results suggest that although the mean column vertical flux convergence is significantly larger than the column-mean horizontal flux convergence, the horizontal flux divergence does play a significant role in midlevels of the convective region. This flux convergence may be associated with a strong front-to-rear inflow that develops during the mature stage of the squall line.

This study suggests that when combined with the independent results of a mesoscale cloud model, the sounding diagnostics can provide a sensitivity test for the Tropical Rainfall Measuring Mission measurements of rainfall and diabatic heating over the life cycle of an entire mesoscale convective system.

1. Introduction

Networks of mesoscale and synoptic-scale rawinsonde observations traditionally have been used to determine the vertical distribution of cumulus heating, drying, and vertical energy fluxes in convective clouds (Yanai et al. 1973; Johnson and Young 1983; Frank and McBride 1989; Chong and Hauser 1990; Gallus and Johnson 1991; Johnson and Bresch 1991; Greco et al. 1994). These networks also have been used to estimate total storm rainfall (Kuo and Anthes 1984; Fuehrberg et al. 1986). Because cumulus convection occurs on scales smaller than can be resolved by conventional observation networks, its effects on larger scales are not directly measured and must be inferred or "diagnosed" as the residual term in budget calculations.

However, inadequate time and space resolution and the inherent inaccuracies associated with the operational networks lead to considerable uncertainty in the residual itself and, ultimately, in the interpretation of the budget results. For example, the storage of cloud water and ice is often neglected in mesoscale water budgets. Cloud storage, however, is known to be important when convective clouds undergo rapid evolution and expansion over time (McNab and Betts 1978; Houze 1982). Surface rainfall is also derived from cloud storage during the storm dissipative stage when the precipitation efficiency exceeds 100% (Cotton et al. 1989).

Furthermore, mesoscale energy budget studies often assume that the vertical eddy heat flux contributes most to the apparent source of moist static energy. The extent to which this assumption remains valid recently has been called into question by Gallus and Johnson (1991) and Greco et al. (1994). When the sounding network is the same size or smaller than the mesoscale convective system (MCS) being sampled, the horizontal cloud-scale flux convergence of heat and moisture across the sides of the budget volume may not be
negligible. If the horizontal flux convergence is neglected, then the actual vertical energy transport diagnosed by the sounding network may be overestimated. An accurate description of the heat balance in the Tropics and export of energy toward higher latitudes (Riehl and Malkus 1958; Riehl and Simpson 1979; Greco et al. 1994) necessitates an accurate specification of the vertical energy fluxes accomplished by the deep convection.

The detailed microphysical and subcloud-scale interactions contained in numerical cloud models make it possible to explicitly calculate energy and water budgets on the scale of the cumulus convection itself (Soong and Tao 1980; Soong and Ogura 1980; Tao and Soong 1986; LaFond et al. 1988; Tao and Simpson 1989; Tao et al. 1991, 1993; Simpson and Tao 1993; Caniaux et al. 1994). This allows the various components of the cloud water budget such as evaporation and cloud water storage to be precisely quantified. The relative contribution of the cloud-scale horizontal and vertical energy flux convergences also can be resolved. This information can be used to help remove some of the aforementioned uncertainties contained in the diagnostic sounding budget calculations.

In the case of the 27 July 1991 squall line over central Florida, the combination of a cloud model with the sounding diagnostics provides a reliable estimate of rainfall when evaluated over the lifecycle and total area of the squall line. The observations for this study are taken from the Convection and Precipitation Electrification Experiment (CaPE), and the cloud model calculations are provided by the Goddard Cumulus Ensemble Model (GCE). The combined techniques can provide an important sensitivity test for the Tropical Rainfall Measuring Mission (TRMM) estimates of rainfall over the large space scales and timescales of an MCS event. Furthermore, this study will show that the GCE model-derived vertical profiles of cumulus heating and drying compare favorably with budget profiles derived from the sounding network. Such comparisons provide an important test of the cloud model performance and are necessary to evaluate the increasingly complex set of physical processes being incorporated into the models.

Section 2 presents a description of the CaPE dataset, the methods used to diagnose the energy and water budgets from the sounding network, and a description of the GCE. Section 3 discusses the synoptic setting and evolution of the 27 July squall line. Section 4 compares the general kinematic and reflectivity structure of the simulated squall line against the CaPE radar and surface Portable Automated Mesonet (PAM) observations. A good level of agreement is obtained and suggests that the cloud model calculations may be used confidently as a tool to refine the sounding-diagnosed budgets. Section 5 diagnoses the water budget residual term from the sounding observations, and the results are used to estimate the total volume of cloud-generated rainfall. Although the initial comparison of the residual with the observed rainfall shows disagreement, this estimate is substantially refined using explicit calculations of cloud precipitation efficiency provided by the GCE. Section 6 presents the sounding-diagnosed vertical profiles of sensible heat source \( Q_s \) and apparent moisture sink \( Q_a \) and compares them against the \( Q_s \) and \( Q_a \) derived from the GCE. The GCE also is used to examine special circumstances in which the horizontal eddy flux convergence of energy is large relative to the vertical flux convergence. Section 7 summarizes the results and suggests problems directed at future research.

2. Methodology

a. Description of the CaPE dataset

This study utilizes data collected from the CaPE conducted in central Florida from 8 July through 18 August 1991. Figure 1 depicts a schematic of the CaPE observational network. The observations used in this study include 1) a mesoscale network of four National Center for Atmospheric Research (NCAR) Cross-chain Loran Atmospheric Sounding System (CLASS) stations; 2) a surface network of 47 PAM stations located within the CLASS network; 3) radar coverage, including both regional-scale surveillance composites of low-level reflectivity and Doppler-derived velocity and reflectivity fields; and 4) half-hourly GOES (Geostationary Operational Environmental Satellite) images.

The CLASS mesoscale sounding network (Fig. 1) is used to calculate volume water and energy budgets for the squall line observed on 27 July 1991. The total area of the sounding quadrilateral is \( 2.4 \times 10^4 \text{ km}^2 \) with an average along-meridional length of 160 km. The mesoscale network samples a large fraction of the squall line as it evolves from the genesis stage (1700 UTC) through early dissipation (2300 UTC). Vector winds are derived from loran navigation tracking of balloon position. Loran methods provide a more sensitive measurement of wind speed and direction than conventional radio-tracking techniques. Values of wind speed and direction are accurate to within 0.5 m s\(^{-1}\) and \( \pm 5^\circ \), respectively (Passi and Morrel 1987). The improved accuracy reduces error in the budget horizontal and vertical advection terms, and thus the error accumulated in the budget residual calculation. The four soundings were released simultaneously at 3-h intervals between 1100 and 2300 UTC, inclusive, on 27 July.

Forty-seven NCAR PAM stations were arranged in a mesonetwork within the sounding array. The average station spacing in the dense east-central region of this array is approximately 10 km. These stations sampled ambient temperature, moisture, pressure, rainfall, and vector winds at 10-m height and stored these measurements as 1-min averages. Additional parameters derived from these variables include the dewpoint, mixing ratio, and equivalent potential temperature.
Regional composites of PPI (plan position indicator) reflectivity collected at 15-min intervals and synthesized from multiple National Weather Service WSR-57 radars operating in the Florida region were obtained through NCAR. The time sequence of images provides information about the squall-line dimensions, intensity, propagation, and evolution of convective and stratiform rain regions at low beam angle and a pixel resolution of about 1 km². The NCAR CP-2 research Doppler radar was located north of Cape Canaveral (Fig. 1). This radar was used to derive PPI sector scans and RHI (range-height indicator) cross sections of the squall line in the S-band (10 cm) and X-band (3 cm) wavelengths. The Doppler-derived horizontal velocity and reflectivity structure are used in this study to validate the results of the two-dimensional cloud model simulation.

b. Calculation of diagnostic energy and water budgets

The vertical profiles of systemwide cumulus heating $Q_1$ and drying $Q_2$ can be diagnosed from ensemble means of the CLASS sounding observations. The diagnostic budget equations are similar to those specified in Johnson and Young (1993) and Johnson (1984):

\begin{align}
Q_1 &= \frac{\partial \tilde{s}}{\partial t} + \mathbf{V} \cdot \nabla \tilde{s} + \tilde{\omega} \frac{\partial \tilde{s}}{\partial p} \\
&= -\frac{\partial}{\partial p} \left[ \tilde{\omega} \tilde{s}' \right] - \nabla \cdot \tilde{s}' \nabla \tilde{r} + L(c - e) + \tilde{Q}_s \\
Q_2 &= -L \left[ \frac{\partial \tilde{q}}{\partial t} + \mathbf{V} \cdot \nabla \tilde{q} + \tilde{\omega} \frac{\partial \tilde{q}}{\partial p} \right] \\
&= -L \frac{\partial}{\partial p} \left[ \tilde{\omega} \tilde{w}' \right] - L \nabla \cdot \tilde{q}' \nabla \tilde{r}' - L(c - e),
\end{align}

where $s = c_pT + gz$ is the dry static energy, $q$ is the water vapor mixing ratio, $\mathbf{V}$ is the horizontal wind velocity, $\tilde{\omega}$ is the sounding network-mean vertical pressure velocity, $c$ is the rate of condensation, $e$ is the rate of evaporation, and $L$ is the latent heat of condensation. In the first line of terms of (1) and (2), ensemble mean quantities calculated from the CLASS rawinsonde network ($\tilde{s}$, $\tilde{q}$) are denoted by a short overbar. The ensemble means are calculated by averaging the individual sounding station profiles of $s$ and $q$ at each analysis time (1100, 1400, 1700, 2000, and 2300 UTC). The terms in the first line are derived from the mesoscale observations and include the local time change ($\partial \tilde{q} / \partial t$), horizontal advection ($\mathbf{V} \cdot \nabla \tilde{q}$), and vertical advection
\((\omega \partial / \partial p)\). The set of primed terms on the second line of (1) and (2) are deviations from the mesoscale observables and are assumed to be associated with unresolvable cloud-scale processes. The individual cloud-scale processes include water phase changes \((c - e)\), radiation effects \((Q_e)\), and the horizontal and vertical divergence of eddy fluxes. In this study, the set of prime terms in (1) and (2) will be calculated explicitly from the high temporal and spatial resolution of cloud-scale processes in the GCE. Profiles of \(Q_1\) and \(Q_2\), derived in this manner are then compared with the profiles inferred from the mesoscale sounding observations.

Numerical methods used to solve (1) and (2) include a leapfrog (centered in time) technique for the time derivative and a finite-centered difference approximation for the horizontal and vertical advection terms. Vertical pressure velocity is computed from the kinematic method. This technique employs a line integral calculation of velocity divergence, which is then adjusted for mass balance according to O’Brien’s (1970) scheme. Budgets are calculated at 25-mb intervals in the vertical and every 3 h between 1100 and 2300 UTC 27 July. This period of time spans much of the lifecycle of the squall line.

The diagnosed total eddy flux of total heat, \(F\), is found by taking the difference between the sounding calculations of \(Q_1\), \(Q_2\), and \(Q_e\) at each pressure level and then integrating the result vertically through the column:

\[
F = \frac{1}{g} \int_{p_1}^{p_2} \left( Q_1 - Q_2 - Q_e \right) dp
\]

\[
= - \frac{1}{g} \omega \bar{h} + \frac{1}{g} \int_{p_1}^{p_2} \nabla \cdot \bar{h} \nabla \cdot dp
\]

where \(h = z + Lq\) is the moist static energy, \(p_1\) is pressure at tropopause level, and \(p\) is pressure at some intermediate level. The two prime terms on the second line of (3) cannot be resolved by the mesoscale sounding network. These terms are the horizontal eddy flux divergence of total heat (term B) and the vertical transport of total heat (term A). In (3), \(Q_e\) is the net radiative heating due to both longwave and shortwave radiative processes. The calculation of \(Q_e\) follows from the one-dimensional radiative transfer calculations described in Greco et al. (1994). Separate equations are used for the calculation of radiative fluxes in clear air and in the presence of clouds. Fractional cloud cover \(\sigma\), at each pressure level is determined from the sounding network-averaged profile of relative humidity. A linear relationship is used between cloud cover and relative humidity. It is assumed that clouds form at a relative humidity of 90% and that the sounding polygon is totally covered by clouds at a relative humidity of 100%. Results of the radiation calculations (Fig. 2) indicate a slight net radiative cooling \((-1\) to \(-2\) K day\(^{-1}\)) during the morning and early afternoon hours in the absence of widespread convective-induced cloud cover. However, late in the afternoon an expanding region of stratiform cloud is generated with the formation of the squall line. This leads to a net cloud-base warming rate of 10 K day\(^{-1}\) and a net cloud-top cooling of \(-15\) K day\(^{-1}\) in the midtroposphere.

An expression for the conservation of water vapor is used to estimate the volume of rainwater generated by the squall line over its life cycle and follows from the methods described in Fuehrberg et al. (1986):

\[
R = P + C - E = - \frac{1}{g} \int_{p_1}^{p_2} \left( \frac{\partial \bar{q}}{\partial t} \right) dp
\]

\[
+ \frac{1}{g} \int_{p_1}^{p_2} \nabla \cdot \left( \bar{q} \nabla \right) dp - \frac{1}{g} \int_{p_1}^{p_2} \frac{\partial}{\partial p} \left( \bar{q} \bar{\omega} \right) dp
\]

where \(\bar{q}\) is the water vapor mixing ratio, \(\bar{V}\) is the horizontal velocity, \(\bar{\omega}\) is the mean vertical pressure velocity, and \(p_1\) and \(p_2\) are the pressure levels bounding the successive 25-mb layers over which the integrations are carried out. In (4), terms on the right-hand side are calculated from ensemble averages of sounding quantities (represented by the short overbars). These terms include the storage of water vapor in the sounding network volume (term A), the horizontal divergence of water vapor flux (term B), and vertical divergence of water vapor flux (term C). A leapfrog differencing technique was used to solve the time derivative. The horizontal flux divergence term (B) was evaluated by taking the line integral of water vapor flux normal to the four sides of the sounding polygon. A finite-centered difference scheme was used in the vertical to evaluate term C. The term on the left side is the budget residual, \(R\), which accounts for unresolved cloud-scale sources and sinks of water vapor within the budget volume. When summed over all vertical layers in the column, the total residual term includes contributions from precipitated water \((P)\), liquid and solid water stored as cloud \((C)\), and evaporation/sublimation of cloud water and ice \((E)\). The relative fraction of the residual term that accounts for surface precipitation, however, is not easily determined from the sounding observations. This poses a major uncertainty in the volume rainwater estimate. The explicit microphysical interactions and cloud-scale conservation equations contained in the GCE will be used in section 5 to derive the squall line precipitation efficiency, thus refining the estimate of total rainwater derived from the sounding budget residual.

\[c.\] Description of the Goddard Cumulus Ensemble Model

The two-dimensional version of the GCE is used in this study. The full formulations, improvements, and
results of recent sensitivity studies are presented in Tao and Simpson (1993). The GCE is nonhydrostatic, and cloud-scale motion is calculated anelastically. The model utilizes a Lin et al. (1983) three-category ice phase scheme to augment a Kessler-type two-category liquid water scheme. Subgrid-scale turbulence processes are parameterized using a turbulent kinetic energy formulation. A stretched vertical coordinate is used to increase resolution in the lowest levels. The model top is about 24 km and uses 33 grid points. The horizontal domain contains 804 central grid points with a 500-m resolution. Outside of this region, the grid is stretched resulting in a total domain that is about 674 km wide. An open lateral boundary condition allows the normal component of velocity to exit through the sides of the domain with minimum reflection. A 5-km-deep Rayleigh absorbing layer is employed at the model top to dampen vertically propagating gravity waves. The version of the model used to simulate the 27 July squall line does not include surface fluxes or radiation processes.

An 8-h simulation of the 27 July squall line was run on the NASA/Goddard Space Flight Center Cray-C98. Initial conditions were specified by CLASS sounding observations taken from Daytona Beach prior to squall-line genesis (Fig. 3). A prescribed vertical profile of mesoscale lifting was based on the sounding calculations of mean divergence over central Florida at the onset of deep convection (1700 UTC). The main effect of the mesoscale ascent was to increase the total cloud rainfall by about 20%. Several sensitivity tests were required to improve the general level of agreement between the observed and simulated squall line. Modifications to the near-surface initial state \((T, \, q)\) were based on the time evolution of surface fields sampled by the dense PAM network. Adjustments to the model base state were not made to explicitly match the observed distribution of energy and water budget processes in time and space. Rather, the aim was to improve the level of correspondence between the simulated cloud and the observed radar structure of the squall line in terms of its fundamental kinematic, thermodynamic, and microphysical properties (discussed in section 4). When satisfactory agreement was obtained, the specific \(Q_1, \, Q_2\) and water budget processes were evaluated and used to compare against and refine the sounding budget diagnostics.

3. Synoptic setting and evolution of the 27 July 1991 squall line

The prestorm vertical temperature and moisture profiles taken from the Daytona Beach sounding at 1700 UTC 27 July 1991 are shown in Fig. 3. The sounding features a moderately unstable air mass [convective available potential energy (CAPE) of 1664 J kg\(^{-1}\)] and a weakly sheared environment characterized by northwesterly flow. Calculations of horizontal divergence made from the mesoscale sounding network (not shown) indicate that strong sea-breeze convergence be-
came established over central Florida through the entire lower troposphere by 1700 UTC (1300 LST). Divergence also was indicated in the midtroposphere and may have been generated by the approach of a shortwave trough in the westerlies. After 1700 UTC, a mean upward velocity exceeding $-4 \mu \text{b s}^{-1}$ was diagnosed through most of the troposphere and sustained through the period of squall-line genesis and evolution.

The evolution of the 27 July squall line is shown by the sequence of composite PPI radar images in Figs. 4a–c. In these images, three levels of reflectivity (and hence inferred rainfall rate) are delineated as level 1 (threshold) — 10 dBZ; level 3 (stippled) — 30 dBZ; and level 5 (solid black) — 50 dBZ. The outline of the mesoscale CLASS network is also shown. At 1800 UTC (Fig. 4a), the first clusters of organized deep convection formed inland of the western Florida coast and propagated eastward along the west coast sea-breeze front. By 2100 UTC (Fig. 4b), the east coast sea breeze collided with the combined gust front—west coast sea-breeze impulse over east-central Florida. The region of convergence between the east coast and west coast sea-breeze circulations is implied from Fig. 5a, which shows an analysis of the PAM surface winds at 2000 UTC. The collision triggered the formation of a second, more intense line of convective activity, which then expanded over the eastern half of the CLASS network. The deep convection within this line is clearly delineated by the embedded level 5 cores (indicative of surface rainfall rates approaching several centimeters per hour) and large horizontal reflectivity gradient. During the squall-line mature stage (2100 UTC), the system propagated toward the east at approximately 6–7 m s$^{-1}$. The early formation of some patchy stratiform rain (level 1 and 3 echoes) can be identified just to the west of the leading convective line.

By 2330 UTC (Fig. 4c), a broad stratiform rain region expanded to the rear of the convective line. The PAM surface analyses of vector winds, equivalent potential temperature $\theta_e$, and pressure are shown at 2000, 2100, 2200, and 2300 UTC in Figs. 5a–d. During this 4-h time period, convective downdrafts and suppressed conditions became progressively better established beneath the stratiform rain shield. This is suggested by 1) formation of an elongated mesohigh accompanied by
an increase in surface pressure of about 0.5 mb; 2) 
-12-K decrease in $\theta_e$; (c) development of an axis 
of divergence (heavy solid line in Figs. 5c,d); and (d) a 
pronounced increase in the westerly outflow to the east 
of the divergent axis. The formation of the mesohigh-- 
wake low couplet in Figs. 5c and 5d is a defining charac-
teristic of a squall-line mesosystem (Fujita 1963). 
After 2300 UTC, the squall line is thus categorized as 
entering its dissipative phase within the sounding net-
work. At this time, precipitation within the CLASS net-
work is mainly stratiform, but the rainfall also contains 
some contribution from an area of deep convection 
located in the far southeastern corner of the sounding 
network (Fig. 4c).

The squall line that formed on 27 July has charac-
teristics of a mesoscale convective system with distinct 
regions of convective and stratiform rainfall (Gamache 
and Houze 1982; Houze 1989). Based on an analysis 
of the CLASS kinematic fields and radar observations, 
the volume water and energy budgets calculated from 
the soundings are classified as undisturbed (1100 and 
1400 UTC); genesis of organized deep convection 
(1700 UTC); squall-line maturity (2000 UTC); and 
dissipation (2300 UTC).

4. Comparison between the cloud model and 
observations on 27 July 1991

In this section, the results of the GCE simulation of 
the 27 July squall line are compared with the CaPE 
radar and surface observations. The purpose of this de-
tailed comparison is to evince confidence in the model 
results and thus use the GCE as a tool to help refine 
our interpretation of the diagnostic sounding budgets. 
Table 1 shows a comparison between the observed and 
modeled squall line dimensions, propagation velocity, 
and $\theta_e$ decrease within the cool pool. These compar-
isons utilize the NCAR composite radar data analysis at 
2330 UTC, surface PAM data, and the model output at 
300 min after initialization. Table 1 reveals that the 
level of agreement obtained between the general attri-
butes of the modeled and observed squall line is very 
favorable. The maximum cloud height, widths of the 
total cloud, and convective and anvil regions compare

---

Fig. 4. (a) Composite plan position indicator (PPI) images of developing convection at 1800 UTC 27 July 1991. Rainfall intensity contours correspond to 10 dBZ (threshold), 30 dBZ (hatched), and 50 dBZ (solid black). (b) As in (a) except for 2100 UTC. (c) As in (a) except for 2200 UTC.
well. The GCE produced a slightly smaller decrease in cool pool $\theta_e$ and a slightly faster storm propagation velocity. The maximum convective updraft speed simulated by the GCE was 16.5 m s$^{-1}$. Although direct observations of updraft velocity could not be obtained for the 27 July system, in situ aircraft penetration of a squall line on the previous day revealed maximum updrafts of 15 m s$^{-1}$.

Figure 6 shows the GCE field of equivalent potential temperature ($K$) (departure from the horizontally averaged mean of all grid points at each vertical level) at 210 min. The total width of the model horizontal domain in this figure is 200 km. The squall line is propagating toward the east, which is located along the right side of the domain. Vertically erect updraft cores and a deep outflow clearly distinguish the multicellular convective region. Buoyant cores penetrate to a maximum height of 13 km. Negative values of $\theta_e$ to $-8K$ are found near the surface in the cool pool, compared with the $-10K$ measured by the PAM network.

Figure 7a shows the two-dimensional GCE field of radar reflectivity at 210 min after initialization. Because the modeled and observed squall lines do not necessarily evolve at the same rate, the best level of correspondence obtained between the modeled and observed squall line during the mature stage motivated the choice of the simulation time used for the comparison. A narrow, deep convective region is located between about 70 and 90 km and is labeled "CON." Both of the two intense convective rain cores contain a maximum reflectivity of 50 dBZ. The leading edge core is in the developing stage and extends to a level of 7 km, and the second deeper, more-mature core extends to nearly 14 km. The deep convection is separated from a narrow anvil region (identified as "ANV") by a distinct and precipitation-free transition zone (Braun and Houze 1994). The width of both the convective and anvil regions at the surface is comparable (both about 25 km wide).

Figure 7b shows the corresponding two-dimensional reflectivity structure obtained from the NCGR CP-2 RHI scan at 2130 UTC. The CP-2 radar site was located just north of KFSC (Fig. 1). The scan was made at an azimuth of 240$^\circ$, which is approximately perpendicular to the long axis of the squall line. The location of this cross-sectional scan is denoted by line $AB$ on the PPI radar image in Fig. 4b. The horizontal range from the radar site and the vertical distance coordinate are delineated by tick marks spaced every 10 km. The total horizontal width of the squall line is about 50 km. Features in the CP-2 analysis appear mirror-imaged with respect to the cloud model reflectivity field in Fig. 7b, such that the direction toward east lies along the left side of the radar figure. The leading edge of the squall line thus lies near the left side of the figure. Separate convective and anvil regions are identified by the CON and ANV labels, respectively. The rainfall reflectivity is based on an S-band (10 cm) wavelength and is contoured every 10 dBZ. Similar to the model results in Fig. 7a, the CP-2 reveals the presence of multiple deep, convective cores that increase in depth toward the rear of the system. The deepest core achieves a height of 14 km and maximum reflectivity of 50 dBZ. These findings agree well with the GCE. As in the GCE, the width of the respective convective and anvil regions are nearly the same size, and the system as a whole displays an upshear tilt through most of its depth. In both the GCE and the CP-2 observations, a region of 30-dBZ rainfall falls from beneath about the 5-km level in the anvil region. In both cases, some of the 30-dBZ rain remains aloft and some reaches the surface. The 5-km level corresponds closely with the height of the melting level as revealed by the CLASS soundings.

The GCE field of horizontal velocity is shown in Fig. 8a. In this figure, front-to-rear flow through the squall line is directed from right to left across the panel, and its magnitude is given by dashed contours analyzed every 5 m s$^{-1}$. A pronounced rear-to-front flow (Smull and Houze 1987a,b; Houze et al. 1990) is oriented from left to right across the figure and is shown in solid contours. The horizontal velocity field in this figure is defined in a ground-relative reference frame. In the model anvil region (labeled ANV), front-to-rear flow is observed through the deep convective region to a height of 9 km. The peak velocity in this layer is $-10$ m s$^{-1}$ and is centered near 4 km. A small area of positive velocity ($5$ m s$^{-1})$ is directed toward the east and away from the upper-level convective region in association with a forward anvil overhang. A second maximum in the front-to-rear inflow branch is found at higher levels in the anvil region and contains a peak magnitude of $-10$ m s$^{-1}$. The model rear inflow layer is confined to below 7 km and contains a peak velocity of 10 m s$^{-1}$. This flow extends toward the surface from below 2.5 km and advances outward toward the leading edge of the convective region.

Figure 8b shows the CP-2 radial velocity normal to the squall line. In this figure, the radial velocity is contoured at intervals of 5 m s$^{-1}$. Positive values are outbound along the range radial and indicate front-to-rear flow through the system. Negative values are inbound and indicate a rear-to-front flow. As in Fig. 8a, all the flows are defined in a ground-relative framework. A broad front-to-rear inflow extends across the width of the system. As in the GCE, a local maximum of 10 m s$^{-1}$ occurs in the deep convective region, but this region is located higher in the troposphere at the 7-km level. Secondary horizontal velocity maxima with a magnitude of 10 m s$^{-1}$ also are found in the anvil region at about 12-km height. Weak rear inflow extends below the 7-km level toward the leading edge of the convective region with a peak velocity of $-10$ m s$^{-1}$ in the surface layer.

Overall, the GCE simulation and the set of CP-2 observations compare well. The important similarities include 1) a multicellular reflectivity structure in the con-
vective region; 2) comparable height and magnitude of the deepest reflectivity core in the convective region; 3) a 1:1 ratio that characterizes the width of the convective to anvil regions; and 4) distinct horizontal front-to-rear and rear-to-front branches of flow. The vertical structure and magnitude of these flow branches compare well and are key identifying features of meso-scale squall lines (Houze 1989, 1990).

5. Budget calculation of rainfall volume

In this section, the diagnostic budget expression for water vapor [Eq. (4)] in section 2 is used to calculate the volume of rainfall as a residual. Figure 9 presents a time–vertical cross-sectional analysis of the distribution of the budget residual term. In this figure, the residual has units of $10^{-3} \text{ kg m}^{-3} \text{ s}^{-1}$. Net positive values of the residual imply net removal of water vapor (through condensation, deposition, and/or precipitation processes), and negative values imply accumulation of vapor (i.e., from evaporation/sublimation). Hatched regions in the figure indicate locations where the residual exceeds $-1.0 \times 10^{-3} \text{ kg m}^{-3} \text{ s}^{-1}$.

The residual term is initially small and negative in the lower and mid troposphere during the morning hours. The implied evaporation may result from lingering cloud debris deposited by the previous day’s deep convection. Negative values also may result from cloud-top evaporation of shallow cumulus. The pronounced negative residual, seen during 1700 UTC at the 700-mb level, suggests that strong evaporation is occurring at the tops of growing cumulus congestus. At 1700 UTC, the residual term becomes positive between 800 and 900 mb. The positive residual grows upward and increases in magni-
itude by 2000 UTC. The vertical expansion and intensification of the residual coincide with the rapid vertical development and organization of the squall line. The largest magnitude of the positive residual is $+4 \times 10^{-3}$ kg m$^{-3}$ s$^{-1}$ in the 700–800-mb layer. This is similar to the height obtained for a Great Plains squall line (Fuehrberg et al. 1986). Figure 9 also shows an increase in evaporation in the boundary layer after 2000 UTC as rainfall accumulates at the surface.

Values of the residual calculated for each 25-mb layer are summed vertically through the entire column at each analysis time. The vertical summation yields total residuals that are positive at 1700, 2000, and 2300 UTC and suggest that the column is being dominated by condensation and precipitation processes at these times. The column totals are summed between 1700 and 2300 UTC (the period of deep convective growth and organization within the CLASS network) in order to derive a storm-total residual and then converted into a mass of water. We infer that this mass of water is generated by cloud processes contained within the sounding envelope. The calculation yields a water mass of $4.6 \times 10^{11}$ kg. The density of liquid water ($10^3$ kg m$^{-3}$) is used to convert this mass into an equivalent volume of liquid water equaling $4.6 \times 10^8$ m$^3$. If we assume that all of the water volume falls out as surface precipitation,

<table>
<thead>
<tr>
<th>Table 1. Comparison of the 27 July squall-line model simulation with observations based on radar and surface PAM data.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Observed squall line</td>
</tr>
<tr>
<td>----------------------</td>
</tr>
<tr>
<td>Height of cloud top</td>
</tr>
<tr>
<td>Storm propagation velocity</td>
</tr>
<tr>
<td>Total system width</td>
</tr>
<tr>
<td>Width of convective region</td>
</tr>
<tr>
<td>Width of stratiform region</td>
</tr>
<tr>
<td>$\theta_e$ decrease (cool pool)</td>
</tr>
</tbody>
</table>
conversion of this volume to an area-averaged depth of surface rainfall yields 1.9 cm.

A network of 72 recording rain gauges is used to provide an independent comparison with the diagnosed rainfall depth. Approximately one-half of these sites are widely distributed throughout the central Florida peninsula and operated by the National Oceanographic and Atmospheric Administration (NOAA) National Climate Data Center (NCDC). The NOAA network is augmented by a dense grid of 36 CaPE PAM rain gauges in east central Florida. The PAM gauge network provided fine spatial sampling in the region where the squall line evolved. Individual gauge rainfall totals accumulated over a 24-h period on 27 July ranged from a trace to 6 cm. The totals were summed and divided by the total number of gauges to obtain a mean areal rainfall depth of 1.2 cm.

The discrepancy between the sounding-diagnosed rainfall estimate (1.9 cm), and the observed average depth of 1.2 cm is about 37%. This is unacceptably large and suggests that other processes such as cloud hydrometeor storage, evaporation, and sublimation of rainfall and cloud water must be accounted for in the budget residual. These results contrast with those of Fuelberg et al. (1986), who found a close level of agreement between their calculated budget residual and observed surface precipitation. They reasoned that cloud storage effects and other unresolved processes were not significant players in the budget residual. As their study pointed out, convective clouds formed both locally in the analysis network and also translated across the boundaries of the region. This contrasts with the 27 July study in which most of the deep convection formed in situ and remained within the confines of the sounding envelope (except at the end of the analysis period between 2300 and 2330 UTC when the leading edge convection moved just outside the eastern leg of the sounding network). Thus, with little of the cloud mass advecting through the sides of the budget volume, the effects of cloud storage may be more critical in the case of the 27 July squall-line study.

As discussed earlier, the storm-total residual term contains contributions from rainfall, cloud water storage, and evaporation/sublimation. Both evaporation and sublimation can occur within the convective downdrafts and also along the cloud edges. The concept of precipitation efficiency (PE), defined here as the ratio of total surface rain to total condensed water vapor (Doswell 1985), provides a measure of the relative fraction of total condensed water vapor neither stored nor evaporated aloft. Numerous studies have determined the average precipitation efficiency of mesoscale convective systems. These findings include 50% for a Colorado hailstorm (Foote and Fankhauser 1973); 49% for a tropical Atlantic squall line (Gamache and Houze 1983); 25%–40% for a midlatitude squall line (Heymsfield and Schotz 1985), and 45%–54% for a west African squall line (Chong and Hauser 1989). From the studies cited, a PE of about 50% would ap-

**Fig. 6.** Two-dimensional GCE model field of equivalent potential temperature (K), departure from the mean at 210 min. Total width of the horizontal domain is 200 km.
Fig. 7. Comparison of (a) the GCE model reflectivity field at 210 minutes, and (b) the NCAR CP-2 range-height indicator (RHI) S-band reflectivity analysis at 2130 UTC. Horizontal and vertical distance marks are spaced every 10 km. A contour interval of 10 dBZ is used in both figures. \textquoteleft ANV\textquoteright refers to the width of the anvil region and \textquoteleft CON\textquoteright refers to the width of the convective region. In (a) the squall line is moving toward the right side. (b) Mirror image of (a) with the squall line moving toward the left.

pear to characterize MCS in general. Additional work by Ulanski and Garstang (1978) suggests that the storm-averaged PE may be as large as 72\% for a long-lived and widespread MCS in Florida. Explicit calculations of cloud-scale water budget processes in the GCE simulation on 27 July imply an average precipitation efficiency of 45\% over the cloud life cycle. Because the level of agreement between the simulated and observed radar squall-line structure is demonstrably good (section 4), the GCE estimate of PE is used here to adjust the diagnosed sounding residual in order to ascertain the fraction of total water mass that contributes to surface rainfall. The model-adjusted, network mean rainfall depth is 0.9 cm, which compares better
with the 1.2-cm depth measured by the rain gauge network (to within 75%).

6. Energy budgets of the 27 July squall line

This section will describe the energetics of the 27 July squall line. The vertical distribution of $Q_1$ and $Q_2$ inferred from the diagnostic budgets will be compared with the $Q_1$ and $Q_2$ profiles calculated explicitly from the GCE model. The total eddy transport of moist static energy is then inferred from the diagnostic sounding calculation in (3). Calculations of the horizontal and vertical flux convergences of energy by the GCE are then used to gain insight into whether a
pronounced drying at 850 mb and its vertical separation from the $Q_1$ peak both indicate deep convective processes (Johnson 1984). As the deep convection develops into a squall line, the region of pronounced heating expands throughout the depth of the troposphere and increases in magnitude. During maturation of the squall line between 2000 and 2100 UTC, the heating peak shifts upward to 550 mb. The low-level drying peak also intensifies and moves upward to 800 mb.

At 2000 UTC, a second peak in the drying curve appears between 400 and 600 mb (Fig. 10d). A double peak such as this is frequently noted in the systemwide $Q_2$ profile due to widely separated drying contributions from cumulus clouds and stratiform anvil (Johnson 1984). By 2300 UTC, the upper-level drying peak intensifies and the lower-level peak weakens. The intensification may be related to the fact that the mesoscale sounding network sampled a broad region of stratiform rain between 2300 and 0000 UTC.

The PAM analyses of surface winds, pressure, and equivalent potential temperature (Figs. 5a–d) reveal that convective downdrafts develop and increase in aerial coverage and intensity after 2000 UTC. After 1700 UTC, a progressive increase in the evaporative cooling and moistening occurs in the lowest 1 km of the $Q_2$ profile and is probably an indication of these downdraft processes. The pronounced upper-tropospheric cooling observed above 200 mb between 2000 and 2300 UTC (Fig. 10c) may be caused by longwave radiative flux divergence and ice sublimation in the stratiform cloud top. This contention is partly supported by the net radiation calculations (Fig. 2), which show that a local region of net cooling (up to a maximum of $-5$ K day$^{-1}$) is generated in the 250–150 mb region between 2000 and 2300 UTC. Results of the GCE $Q_1$ calculations also reveal that a net cooling in the $Q_1$ profile near the top of the troposphere (150 mb and above) can be caused by the vertical divergence of the eddy heat flux, $\delta P/\delta z (\rho w' \theta')$.

The vertical profiles of the system-total cumulus heating and drying inferred from the sounding budgets [sum of terms on the first line of (1) and (2)] are now compared with profiles calculated directly from the subcloud-scale processes contained in the GCE model [sum of terms on the second line of (1) and (2)]. These results are shown in Fig. 11. In this figure, results of the sounding budget calculations during the mature phase between 2000 and 2100 UTC (Fig. 11a) are compared with the model budget results averaged over both the first four hours of the squall-line simulation and the total number of model grid points (Fig. 11b). The first 4 h of the model simulation contain much of the period of squall-line maturity. Both the cloud model and diagnostic budgets have been normalized according to the surface rainfall rate (cloud model is 1.5 cm day$^{-1}$, observed system is 1.2 cm day$^{-1}$), and the same

---

**Fig. 9.** Vertical–time cross-sectional analysis of the diagnostic water budget residual term analyzed at 3-h intervals between 1100 and 2300 UTC. Values of the residual are expressed in units of $10^{-3}$ kg m$^{-3}$ s$^{-1}$. Positive contours denote regions where the residual exceeds zero, indicative of water vapor removal through condensation and precipitation. Hatched regions correspond to negative values less than $-1 \times 10^{-3}$ kg m$^{-3}$ s$^{-1}$.

---

**a. Cumulus heating and drying profiles**

Figures 10a–d show the time evolution of diagnosed vertical profiles of the sensible heat source $Q_1$ and apparent moisture sink $Q_2$ at different times during the storm life cycle. Heating and cooling rates for both sets of profiles are expressed in terms of kelvins per day.

The prestorm $Q_1$ and $Q_2$ signals calculated at 1100 and 1400 UTC (Figs. 10a and 10b) generally are very small throughout the troposphere, with the exception of strong surface heating in the $Q_1$ profile at 1400 UTC. With the onset of organized convection in the budget network at 1700 UTC, the heating peak shifts upward to 700 mb (Fig. 10c). The simultaneous appearance of complete neglect of the horizontal flux convergence term [term B in (3)] in the sounding-diagnosed total of $F$ is justified (or, in other words, whether the diagnosed total can be interpreted uniquely as a vertical eddy transport).
heating—cooling rate scale [K day$^{-1}$ (cm day$^{-1}$)$^{-1}$] is utilized in both sets of profiles.

Close agreement is obtained between the overall shape of both sets of heating and drying profiles and also the vertical location of the profile maxima. The model places the vertical location of the heating peak at 550 mb and simulates the shallow layer of cooling near the surface. Near-exact correspondence is obtained with regard to the location of the lower tropospheric drying maximum at the 800 mb level. The abrupt transition to an isolated, slight moistening signature at 750 mb is also simulated. The cloud model tends to split the second, upper-level drying maximum into two smaller peaks, but nonetheless predicts a deep layer of net drying between 700 and 300 mb. The cloud model also simulates the relative difference in magnitude of the respective $Q_1$ and $Q_2$ profiles (e.g., $Q_2 > Q_1$ in the lower troposphere, $Q_1 > Q_2$ in the upper troposphere).

The general magnitude of the modeled $Q_1$ and $Q_2$, however, is considerably smaller than the corresponding magnitude of the sounding-diagnosed $Q_1$ and $Q_2$. 
Some of this difference may have occurred because the cloud model budgets are averaged with respect to a 4-h period, whereas the diagnosed budgets are calculated over a shorter 1-h interval (e.g., between 2000 and 2100 UTC) and thus represent a more instantaneous "snapshot." Furthermore, the cloud model does not capture the slight boundary layer moistening observed in the diagnostic $Q_2$ profile. This may have occurred because the version of the cloud model used in this simulation does not contain a surface moisture flux parameterization.

b. Eddy fluxes of heat and moisture

When energy budgets are diagnosed from a large-scale network of soundings, it is often assumed that the contribution of the horizontal eddy flux convergence [term B in (3)] to the diagnosed total transport is small. Under this assumption, the diagnostic total becomes a reliable estimate of the vertical energy transport accomplished by the convection contained within the sounding network. This assumption probably is valid when the area of active convective cloud is quite small relative to the total area of the large-scale sounding network (Greco et al. 1994). However, when a mesoscale network of soundings and the MCS being sampled are of similar size, portions of the MCS circulations may extend across the lateral boundaries of the sounding volume. Horizontal fluxes of heat and moisture associated with the convective-scale and mesoscale flows transect the sides of the sounding volume. The radar image in Fig. 4c indicates that the squall-line convective region translates through the eastern leg of the mesoscale sounding network between 2300 and 2330 UTC. Appreciable horizontal transfers of heat and moisture may occur perpendicular to this leg in the form of front-to-rear and rear-to-front flows through the active convective region (Gallus and Johnson 1991; Tao et al. 1992).

To gain qualitative insight into whether the horizontal flux in the convective region may be significant compared with the vertical eddy flux, the GCE is used to calculate the heat and moisture flux convergence inside the mature convective region of the simulated squall line. Figure 12a shows the vertical profile of the vertical and horizontal eddy heat flux convergences ($Q_{1w}$ and $Q_{1h}$, respectively), and Fig. 12b shows the two components of the eddy moisture flux convergence ($Q_{2w}$ and $Q_{2h}$). The flux convergences in both figures are shown in terms of their absolute magnitude and are expressed in units of kelvins per hour. It is apparent that the vertical heat flux convergence (Fig. 12a, dashed curve) is about an order of magnitude larger than the horizontal heat flux convergence, except in the lowest 100 mb where the difference is indistinct and in the vicinity of 500 mb. The vertical flux convergence of moisture (Fig. 12b) appears to dominate much of the upper troposphere but becomes quite small relative to the horizontal flux convergence in the 600–700-mb layer. The relatively large horizontal eddy flux convergence of energy in the midtroposphere may be gener-

---

**Fig. 11.** Vertical profiles of the squall-line total sensible heat source $Q_1$ and apparent moisture sink $Q_2$ obtained from (a) diagnostic sounding energy budgets at 2000 UTC and (b) two-dimensional cloud model averaged over the 4-h simulation of the squall-line mature stage. Both sets of profiles have been normalized according to rain rate (cm d$^{-1}$). The model profiles at each level have been averaged with respect to the total number of horizontal grid points contained in the simulated convective region. Heating and cooling rates are expressed in units of kelvins per day per centimeter per day.
ated by a significant front-to-rear flow directed through the convective region in the 500–800-mb layer (Fig. 8a). Small inflections in the base-state moisture profile (Fig. 3) also may be related to sharp features in the eddy flux divergence profile (Dudhia and Moncrieff 1987). The exact cloud-internal mechanisms that contribute to the relative size of the horizontal and vertical flux convergences will be investigated in future GCE simulations.

Averaging the horizontal and vertical eddy flux convergence terms through the column reveals that the net column vertical heat and moisture flux convergences are about 77% and 72% larger than the net horizontal flux convergences of heat and moisture, respectively. However, because appreciable horizontal transfer of latent and sensible heats are generated in the middle portions of the column, caution still must be used when the sounding-diagnosed total F is interpreted as an unambiguous measure of the vertical convective energy transport.

Figure 13 displays the vertical profiles of the total eddy flux of moist static energy, $F$, at 1400, 1700, 2000, and 2300 UTC. These profiles area obtained by vertically integrating the difference between the sounding-diagnosed $Q_1$, $Q_2$, and $Q_6$ profiles. The pre-storm vertical profile of total eddy flux (1400 UTC) is generally small throughout the column, but an exception occurs in the surface layer. Here strong sensible heat fluxes associated with shallow cumulus convection may be mixing heated surface air upward. With the onset of organized deep convection at 1700 UTC, the total eddy flux increases by nearly 300% and the peak shifts upward to 750 mb. As the deep convection organizes into a squall line at 2000 UTC, however, the peak flux appears to decrease. The decrease is most marked in the lower troposphere and continues until 2300 UTC, at which time the total eddy flux becomes negative. At 2300 UTC, the profile of $F$ is negative through most of the column but most apparently below the 550-mb level.

In the calculation of $F$ [Eq. (3)], it was found that the $Q_6$ term is small relative to the difference between $Q_1$ and $Q_2$. This is true even at 2300 UTC, when cloud-base warming and cloud-top cooling are fairly substantial. This is because the differential heating and cooling rates occur over a fairly shallow layer. When the rates are integrated vertically in (3), their effect on $F$ tends to cancel. Thus, negative values of $F$ arise chiefly because drying effects associated with the $Q_2$ progressively increase and heating effects associated with the $Q_1$ progressively decrease over time. The exact reasons for this remain unclear. In terms of the vertical eddy flux, both cold downdrafts ($h' < 0, \omega' > 0$, where $h'$ is the departure from the mean profile of moist static energy) and warm updrafts ($h' > 0, \omega' < 0$) can potentially contribute to $F > 0$. However, a warm downdraft, which is generated through mesoscale descent beneath the anvil cloud, can contribute to negative $F$ ($h' > 0, \omega' > 0$). The presence of a warm, dry mesoscale downdraft at 2300 UTC is difficult to establish from sounding sites at Ruskin and Dunnellon (not shown) in the wake of the stratiform cloud region. However, these locations may be too far removed from the cloud disturbance to reveal the effect clearly. The

![Fig. 12. Absolute magnitude of the GCE model horizontal and vertical flux convergence (K h⁻¹) of heat ($Q_{xh}$ and $Q_{yh}$, respectively), and moisture ($Q_{xm}$ and $Q_{ym}$, respectively) in the model convective region during squall-line maturity. The flux convergences have been averaged over the total number of horizontal grid points in the convective region and over the 4-h period of squall-line maturity.](image-url)
surface pressure field at 2200 and 2300 UTC (Figs. 5c and 5d) does show a weak wake low beneath the stratiform rain region, which is an indicator of warming in the presence of mesoscale descent (Johnson and Hamilton 1988). It is also possible that the horizontal eddy flux convergence term in (3) contributes to negative $F$. The questions raised in this analysis warrant further investigation using the GCE to separate the individual effects of convective and mesoscale updrafts and downdrafts.

7. Summary and conclusions

In this study, mesoscale energy and water budgets calculated from CaPE are combined with the results of the two-dimensional version of GCE model. The goal is to refine the calculation and interpretation of energy and water budget processes on mesoscale time and space scales. This is achieved by utilizing the following three elements:

1) High spatial and temporal resolution of convective-scale cloud processes are contained in a numerical cloud model. The model can provide detailed calculations of the critical microphysical processes and fluxes that generate sources and sinks of heat energy and water that cannot otherwise be resolved by a mesoscale network of soundings;

2) Sensitive sounding measurements of the horizontal vector wind, such as those provided by the NCAR CLASS are used. Because the budget residual term is calculated from a very small difference between the large-scale horizontal and vertical advective terms, a more precise specification of the horizontal wind potentially will reduce error in the residual.

3) These techniques are applied to a well-defined MCS sampled in its near-entirety in both space and time by a mesoscale network of soundings.

Comparison of the diagnostic budget results with the GCE budgets reveals three important conclusions. First, near-agreement in the vertical profiles of cumulus heating and drying is obtained. The cloud model successfully reproduces (a) the shape of the heating and drying profiles; (b) the vertical location of the individual heating and drying peaks; and (c) the relative magnitude of the heating versus drying. Because the GCE $Q_1$ profiles are being used to develop heating rate algorithms for the TRMM rainfall measurements (Simpson and Tao 1993), independent verification of the GCE $Q_1$ and $Q_2$ budgets using the sounding observations constitutes a vital component of ongoing model research and development.

Second, the cloud model can be used as a guide to identify circumstances in which the relative contribution of the horizontal eddy flux convergence to the total transport, $F$, may be significant. The total $F$ diagnosed from the sounding budgets contains an unknown combination of the horizontal flux convergence and the ver-
tical eddy flux of energy. In circumstances where the space scale of the sounding observations is comparable to the horizontal dimensions of the convective system, total neglect of the horizontal flux convergence may be unjustified and can lead to an overestimate of the vertical energy transports accomplished by the MCS. In the 27 July cloud simulation, the horizontal eddy flux convergence of both heat and moisture are shown to be nonnegligible in the midlevel tropospheric region of the deep convective line. Continued cloud modeling work is needed to elucidate the relative magnitude of the horizontal flux convergence in the convective, stratiform, and transition regions of an MCS and the exact mechanisms that control the time and space distribution of these processes. For instance, the horizontal flux convergence may increase later in the lifecycle of an organized convective system, as erupt updrafts weaken and become progressively more tilted into a front-to-rear-oriented current through the cloud. If the horizontal flux convergence can be quantified adequately, then it is possible to evaluate more effectively the efficiency of the MCS as a vertical energy pump.

Results from the sounding-diagnosed total eddy heat flux suggest that the largest fluxes of moist static energy occur very early in the development of the squall line. As the deep convection intensifies and becomes better organized, the total positive energy flux declines. The total eddy flux then becomes negative through much of the lower and midtroposphere during dissipation of the system. The reasons for this negative flux are probably very complex and require a more complete cloud model evaluation of the relative role of downdrafts on the convective scale and mesoscale, and also the possible contribution of the horizontal eddy flux convergence.

Third, precipitation efficiency is calculated using the cloud model. These results are then combined with the residual term from a sounding-diagnosed water budget calculation. This leads to a refined estimate of total rainfall for the entire squall-line system over most of its life cycle. The combined technique yields a network-averaged rainfall depth of 1.0 cm, which compares well with the 1.2 cm calculated from an independent network of 72 rain gauges.

The accurate measurement of rainfall amount over large-scale regions of the Tropics and subtropics remains a fundamental problem. The combination of cloud model and sounding diagnostics described herein produces better results than the diagnostic calculations alone. Ultimately, this technique can be used to place constraints on remotely sensed precipitation estimates, particularly on the large time and space domain sampled by the TRMM satellite.

Acknowledgments. The authors gratefully acknowledge the support provided by the NASA TRMM program. We wish to thank Claire Cosgrove, Clay Davenhport, and Ed Hutchison of the University of Virginia for their assistance with computer programming and data reduction. Dr. Jeff Caylor of Science Systems Applications, Inc., provided valuable assistance with the CP-2 radar processing. Computational support for the cloud model simulations was provided by the NASA/Goddard Space Flight Center. Finally, we wish to acknowledge several anonymous reviewers who helped to significantly improve the manuscript.

REFERENCES


