Heat and Moisture Budgets of an Intense Midlatitude Squall Line

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ABSTRACT

Rawinsonde data from OK PRE-STORM are used to calculate the heat and moisture budgets during the late mature through dissipating stages of an intense squall line system with a trailing stratiform region. Budgets have been computed at three separate times using composited data from approximately 30 rawinsondes over three-hour intervals. Data spacing is marginally adequate to resolve features within the broad (≈200 km) stratiform region, but not the narrow (≈40 km) leading convective line. Low-level radar data and surface accumulated rainfall are used to partition the system into convective line and stratiform regions, and to check the accuracy of the heat and moisture budgets.

The squall line had a pronounced descending rear inflow jet that extended toward the front of the system with time, and strong front-to-rear flow above and beneath this jet. The diagnosed vertical motion showed deep ascent in the leading convective line and upper-level descent above lower-level descent in the trailing stratiform region. The two centers of ascent were somewhat merged together, due in part to resolution limitations. The axes of ascent and descent weakened and became increasingly sloped over time as the trailing stratiform region widened and became separated from the leading convective line. The heat source (Qh) distribution and its time variations are similar to those of vertical motion with heating peaks between 300 and 400 mb in the leading convective line and trailing stratiform region. Cooling due to evaporation and melting occurred in the stratiform region and was most intense just behind the back edge of stratiform echo around 550 mb. The moisture sink (Qe) peaked in the convective line region, but at lower levels than the heat source. Moistening due to evaporation was typically strongest around 700 mb in the stratiform region.

Vertical integrations of both budgets produced rainfall rates generally close to the observed ones for averages over the entire system. Convective line rates were underestimated by 40%, due in large part to inadequate sounding resolution. In the later stages, when the stratiform region was well separated from the leading convective line and could be reasonably well resolved, diagnosed rates underestimated observed rainfall rates by as much as 2–3 mm h⁻¹ as the system decayed. Radar reflectivity data showed that the rearward transport of hydrometeors from the leading convective line could add as much as 2–4 mm h⁻¹ to the diagnosed stratiform precipitation rates. This transport, along with a possible additional contribution from the fallout of hydrometeors stored from earlier times, can well explain the deficiencies in rainfall rates diagnosed from the moisture budget.

1. Introduction

Research from several field projects during the last few decades has shown that well-developed squall line systems have detailed internal structures with very distinct circulations taking place on different horizontal scales. One important class of squall lines contains intense convective-scale updrafts, which occur within a narrow line at the leading edge of the system, and to the rear a broad mesoscale updraft at high levels and a mesoscale downdraft at lower levels within a region of widespread stratiform rain (Houze 1977; Zipser 1977; Ogura and Liou 1980; Gamache and Houze 1982). The stratiform rainfall has been shown to account for roughly 30%–50% of the total rainfall occurring with these systems in both the tropics and mid-latitudes (Cheng and Houze 1979; Johnson and Hamilton 1988), and therefore, a significant amount of the heat and moisture transport in these systems must take place in their stratiform portions. Unlike the convective line region where condensation generally occurs throughout the troposphere, the stratiform region is characterized by evaporation at low levels with condensation restricted to higher levels within the extensive trailing anvil (Gamache and Houze 1983). The differences in the circulations within the convective line and stratiform rain regions of a squall line result in pronounced differences in their vertical heating/cooling and drying/moistening distributions (Houze 1982; Johnson 1984).

The OK PRE-STORM (Oklahoma–Kansas Preliminary Regional Experiment for STORM-Central) project that took place in the south-central United States in 1985 provided an extensive dataset from which
midlatitude squall lines could be studied in great detail. Budget studies done on short space and time scales for midlatitude convection (e.g., Kuo and Anthes 1984) have not been as common as those done on the synoptic scale over longer time periods for tropical convection (e.g., Yanai et al. 1973). Other studies which have used composited data (Gammache and Houze 1982; 1983, for a tropical case; Ogura and Liou 1980 for a midlatitude case) have generally done so over time intervals of many hours (nine in the aforementioned studies), thereby not allowing the observation of the evolution of the convective systems. Betts (1973) used composited data over a small time interval (two hours) to study the growth and decay of tropical convection, but had access to data from only one rawinsonde. Budget studies done over long temporal and spatial scales can neglect hydrometeor storage effects in the budgets, because cloud storage is relatively small in those cases. Storage effects have also been generally ignored for budget studies done on smaller time and space scales, with the exception of McNab and Betts (1978). They found water storage to be important in regions of convection where cloud cover was increasing substantially in area during the developing stage of convection. The dissipating stage of convection was not considered in their study. (Their budgets were based on 14 days of rawinsonde data, however, and specific results for a single case were not presented.) Little research has been done to study storage effects during dissipating convection, although Cotton et al. (1989) show that precipitation efficiency exceeds 100% during the later stages of an MCC (mesoscale convective complex), implying that storage is important.

The main objective of this paper is to study variations in the heat and moisture budgets for an intense midlatitude squall line using rawinsonde data for the 10–11 June 1985 PRE-Storm case. This case has been the subject of numerous other studies (Augustine and Zipser 1987; Smull and Houze 1987b; Johnson and Hamilton 1988; Rutledge and MacGorman 1988; Rutledge et al. 1988; Zhang et al. 1989; Zhang and Gao 1989; Johnson et al. 1990). Composites have been done using 90-min rawinsonde data taken over 3-h intervals, a period sufficiently short to 1) allow an assumption of near steady-state conditions and 2) permit a determination of the evolution of the budgets from the mature through dissipating stages of the convective system. Kinematics will also be examined, and rainfall rates diagnosed from the budgets will be compared with observed values.

Observational studies of the heating and moistening profiles within mesoscale convective systems (MCSS) are necessary for the accurate parameterization of moist convection in numerical models. The level of peak heating and drying is particularly important. This paper will add to the observational knowledge of midlatitude squall line systems, and will examine heating and drying rate variations in the horizontal and vertical.

2. Budget equations and data analysis

a. Budget equations

Cumulus convection occurs on scales smaller than those normally resolvable from data gathered by conventional means, and therefore, its effects on the larger-scale circulation cannot be directly measured. Instead, effects are inferred indirectly from diagnostic budgets. For this study, the apparent heat source \( Q_1 \) and apparent moisture sink \( Q_2 \) (from Yanai et al. 1973), with modifications to account for ice effects, are defined by

\[
Q_1 = \frac{\partial \bar{s}}{\partial t} + \nabla \cdot \bar{s} \mathbf{V} + \frac{\partial \bar{s} \omega}{\partial p} = Q_R + L_v (\bar{c} - \bar{e}) + (L_v + L_f) (\bar{d} - \bar{s}_*) + L_f (\bar{f} - \bar{m}) - \nabla \cdot \bar{s} \mathbf{V} - \frac{\partial}{\partial p} \bar{s} \omega, \tag{1}
\]

\[
Q_2 = -L_v \left( \frac{\partial \bar{q} \omega}{\partial t} + \nabla \cdot \bar{q} \mathbf{V} + \frac{\partial \bar{q} \omega}{\partial p} \right) = L_v (\bar{c} - \bar{e}) + L_v (\bar{d} - \bar{s}_*) + L_v \nabla \cdot \bar{q} \mathbf{V} + L_v \frac{\partial}{\partial p} \bar{q} \omega, \tag{2}
\]

where \( s = c_p T + gz \) is the dry static energy; \( q \) the specific humidity; \( L_v \) and \( L_f \) the latent heats of vaporization and fusion; \( c, e, d, s_*, f, \) and \( m \) the condensation, evaporation, deposition, sublimation, freezing, and melting rates; and \( Q_R \) the average radiative heating rate. Primes denote deviations from a horizontal average (designated by overbar). From Eq. (1), it can be seen that \( Q_1 \) is a measure of radiative heating, latent heating from net condensation and ice phase changes, and the convergences of the vertical and horizontal eddy transports of sensible heat. Often for large-scale studies, the horizontal eddy flux convergences are assumed small, but they may be important for the applications considered here. For example, within a squall line system, subgrid heat and moisture transport from the convective line to the stratiform region may not be negligible. Likewise, in Eq. (2), \( Q_2 \) is a measure of the net condensation and the horizontal and vertical convergences of the eddy moisture transports. Taking the difference of (1) and (2) gives

\[
(Q_1 - Q_R) - Q_2 = L_f (\bar{d} - \bar{s}_*) + L_v (\bar{f} - \bar{m}) - \nabla \cdot \bar{h} \mathbf{V} - \frac{\partial}{\partial p} \bar{h} \omega, \tag{3}
\]

where \( h = c_p T + gz + L_v q \) is the moist static energy. Equation (3) is a measure of changes in \( h \) due to radiation, convergences of the eddy moist static energy transport, and heating or cooling due to ice effects. If ice effects are assumed small and there is no transport from convective processes, \( Q_1 - Q_R \approx Q_2 \) (Arakawa and Chen 1987). Therefore, one indicator of stratiform
rainfall is approximately equal profiles of $Q_1$ and $Q_2$
(assuming small $Q_R$; Luo and Yanai 1984). Precipitation rates can be
diagnosed from the heat and moisture budgets by integrating Eqs. (1) and
(2) from $p_T$ (cloud top pressure) to $p_s$ (surface pressure)
(Yanai et al. 1973). Integration of (1) gives
\[
\frac{1}{g} \int_{p_T}^{p_s} (Q_1 - Q_R) dp = \frac{L_v}{g} \int_{p_T}^{p_s} (\tilde{e} - \tilde{e}_0) dp + \frac{(L_v + L_f)}{g} \int_{p_T}^{p_s} (\tilde{d} - \tilde{s}_0) dp - \frac{1}{g} \int_{p_T}^{p_s} \nabla \cdot \mathbf{\overline{q}} \overline{\nu} dp + \frac{L_f}{g} \int_{p_T}^{p_s} (\mathbf{\tilde{f}} - \tilde{m}) dp + S_0,
\]
and (2) gives
\[
\frac{1}{g} \int_{p_T}^{p_s} Q_2 dp = \frac{L_v}{g} \int_{p_T}^{p_s} (\tilde{e} - \tilde{e}_0) dp + \frac{L_v}{g} \int_{p_T}^{p_s} (\tilde{d} - \tilde{s}_0) dp + \frac{L_v}{g} \int_{p_T}^{p_s} \nabla \cdot \mathbf{\overline{q}} \overline{\nu} dp - L_v E_0,
\]
where the first two terms on the right-hand side of each equation
represent water production (fallout + storage), and the third terms are the
net convergence of the horizontal eddy heat flux (4), and moisture flux
(5). The fourth term on the right-hand side of (4) should be rather small,
since some inherent cancellation would take place due to freezing aloft and melting
below. Finally, the surface sensible heat flux, $S_0$, and
surface evaporation term, $L_v E_0$, are also small for midlatitude
convection at night and can be neglected in the budget computations.

Equations (4) and (5) can be used to diagnose the precipitation budget
since $Q_1$ and $Q_2$ are measurable quantities. The observed precipitation rate at the
surface is the sum of the fallout of water generated in the column and the fallout of water transported into it or stored from an earlier time. Therefore, if the net
convergences of eddy fluxes are assumed small [along with the integrated effects of melting and freezing in (4)], discrepancies between the measured rainfall and the integrated values of $Q_1$ and $Q_2$ may be thought of as a
measure of the importance of both water storage and transport. As stated earlier, storage effects are minimal for budgets over long time periods, but for this case, on a shorter time scale, they are substantial (McNab and Betts 1978). Storage becomes considerable during dissipation convection when hydrometeor production may cease, but stored water falls out as rain. Transport is important in the stratiform region where large amounts of ice and liquid water are carried rearward from the convective line.

The vertical eddy flux of total heat, $F$, at any level can be calculated by integrating Eq. (3) downward to that level from the pressure at cloud top (Yanai et al.
1973). If the integration is done over the same limits as that done to (1) and (2), the vertical eddy flux of total heat at the surface can be determined and used as an additional check on the accuracy of the budgets. This integration of (3) yields
\[
\frac{1}{g} \int_{p_T}^{p_s} (Q_1 - Q_2 - Q_R) dp - \frac{L_f}{g} \int_{p_T}^{p_s} (\tilde{d} - \tilde{s}_0 + \mathbf{\tilde{f}} - \tilde{m}) dp + \frac{1}{g} \int_{p_T}^{p_s} \nabla \cdot \mathbf{\overline{h}} \overline{\nu} dp = F_0 = S_0 + L_v E_0.
\]

When averaged over the entire system (in which case ice effects and the horizontal eddy flux term become small), $F_0$ should approach zero since surface sensible heat flux and evaporation are also small for the conditions under which this particular squall line occurred.

In the heat and moisture budgets, finite differences are used to determine horizontal and vertical derivatives, and because the horizontal grid spacing is approximately 50 km, variables with overbars in the equations represent averages over areas of 50 km × 50 km centered at the grid points. Vertical velocities ($\omega$) are calculated at each grid point from the kinematic technique using the mass continuity equation. Adjustments are made using a procedure from O'Brien
(1970) to make $\omega$ vanish at 125 mb. This level is chosen as the tropopause because it represents the highest tropopause observed in the region, and also the highest level at which a majority of rawinsondes reported data. As an independent check on $\omega$ at high levels, vertical velocities are calculated using a rewritten form of (1) (Nitta 1977),
\[
\tilde{\omega} = \frac{[\mathbf{-} \tilde{d} / \frac{\partial}{\partial t} - \mathbf{\nabla} \cdot \mathbf{\nabla} \tilde{q} + Q_R]}{\frac{\partial \tilde{\nu}}{\partial p}}
\]
which is valid assuming latent heat effects and the convergence of eddy heat flux are small. Using reasonable values of $Q_R$, $\omega$ is found to decrease quickly to less than 1 µm s$^{-1}$ at most grid points by 125 mb. Therefore, the choice of 125 mb as the level of zero vertical motion appears reasonable.

b. Data analysis

For this budget study, dry static energy and specific humidity data from the PRE-STORM supplemental sounding network were along with the accumu-
lated rainfall measurements from the network of NCA PAM (Portable Automated Mesonet Network-II) and NSSL SAM (Surface Automated Mesonet) stations. Both of these networks, in addition to the radar, National Weather Service (NWS) rawinsonde, and wind profiler sites can be seen in Fig. 1 (from Cunning
1986). In order to decrease the station spacing and better resolve small-scale features in the region, com-
posites were made using the supplemental soundings taken over 3-h periods centered at 0300, 0600, and 0730 UTC, using an estimated average squall line motion. Since the soundings were taken at 90-min intervals, the compositing reduced the station spacing from 150 to roughly 80 km. It was believed that the best speed of motion to use in compositing was an average of the motions of the central axis, leading edge, and back edge of the radar echo over each 3-h period. This averaging was done since the stratiform region was moving at a slower speed than the convective line. From this procedure, the system appeared to move from 315° on average with a speed decreasing from 16 m s⁻¹ at 0300 UTC to 13 m s⁻¹ at 0600 UTC, and then increasing again to 14 m s⁻¹ by 0730 UTC. This agrees fairly well with the storm motion of 14 m s⁻¹ assumed during the period 0130–0530 UTC by Rutledge et al. (1988) for a Doppler radar study of this case. Compositing procedures assume a steady-state system, which is probably a reasonable approximation for the relatively short 3-h time interval used.

Compositing increased the number of stations used in the analyses to 30 or more, and the location of these sites as well as the vertical extent of the soundings can be seen in Fig. 2. In general, rawinsondes adequately sampled the atmosphere not only within the convective system, but also ahead of and behind it. An exception is within the stratiform rain region where data from the soundings often ended in the 450–650-mb layer. NWS soundings were included in the analyses only when they were taken within the 3-h time interval and were needed to help fill data-sparse areas.

Because the average station spacing of the composited data was 80 km, features within the ~40 km-wide leading convective line could not be resolved, and aliasing of data associated with this narrow phenomenon was common. At 0600 UTC, aliasing appeared to present the most obvious problems with dry air from ahead of the system being smoothed into the convective line in the objective analysis. This caused spurious features in the moisture budget. Therefore, four bogus soundings (shown by circles in Fig. 2) were included.
at that composite time. One was placed in a data-sparse area of the presquall environment using both thermodynamic and wind data averaged from two other presquall sites. The other three (humidity data only) were positioned at the leading edge of the convective line, and were given nearly saturated profiles following the profile of the Chanute 0535 UTC sounding which ascended at the leading edge of the system. The addition

Fig. 2. Locations of composited soundings used in the budget studies at (a) 0300, (b) 0600, and (c) 0730 UTC. Squares represent rawinsonde sites, circles bogus soundings. The length of the vertical line at each station is proportional to the balloon ascent distance. The composited radar echo is marked with a solid line, and dashed lines divide it into stratiform and convective precipitation regions. The area between the dashed lines in (b) and (c) is the transition zone. Dashed lines outside the squall line mark regions of scattered weak echoes. Station names are shown at 0300 UTC. Pressure scale at lower left is in millibars.
of the bogus soundings helped to diminish the aliasing of specific humidity data, but did not completely solve the problem (as will be seen in Fig. 12).

Data from the composited rawinsondes were gridded onto a 0.5° by 0.5° grid using a Barnes objective analysis scheme (Barnes 1964). The radius of influence was adjusted to allow a response of approximately 50% for wavelengths of 160 km or twice the average station spacing (Barnes 1973). This means that mesoscale features with scales at least that large should be adequately resolved in the analyses, e.g., the stratiform region, but convective-scale phenomena with scales of only a few tens of kilometers at most will not be fully resolved. Therefore, magnitudes of quantities like vertical velocity are less than those observed on the cloud scale, a common problem in diagnostic budget studies (Ogura and Liou 1980; Kuo and Anthes 1984).

Because there is a slope of the topography in the PRE-STORM region (station pressures fall from around 980 mb in the east to 930 mb in the west), a sloped lower boundary was taken into account during the computation of ω. A comparison between ω calculated with a sloped lower boundary and ω determined from a common lower boundary of 975 mb showed differences of only a few microbars per second at most. Typical descent in the mesoscale downdraft and ascent in the convective line and mesoscale updraft were on the order of tens of microbars per second, so that the differences in ω calculated by the two different methods were relatively small in light of the inherent errors in the rawinsonde winds used to compute omega. Balloon drift, important on this scale (Fankhauser 1974), was taken into account in the analyses.

Low-level radar reflectivities from the NWS WSR-57 radars located at Wichita, Kansas, and Oklahoma City, Oklahoma, were used in this study to partition the squall line into convective line, transition zone, and stratiform rain regions. Smull and Houze (1985) described these regions for a mature squall line system having a trailing stratiform precipitation region as the leading convective line, 10–50 km wide, with heavy rainfall, the transition zone with weak reflectivity and minimal surface precipitation; and the trailing stratiform region, 100–300 km wide, with light to moderate surface rainfall rates (≈1–10 mm h⁻¹). For this study the back edge of the convective line was approximated by the 25 dBZ contour, with the front edge of the stratiform rain region estimated at the location where reflectivities began to increase behind the convective line. A transition zone existed during the time period at which the budgets were conducted, and it was assumed to lie between the convective line and stratiform regions. A simple composite was made at each analysis time by taking the average position of the leading edge of the radar echo at the beginning, middle, and end time of the 3-h interval. The same procedure was used to determine the position of the back edge of the echo, and the boundaries between the different regions. The evolution of this composited echo can be seen in Fig. 2. At 0300 UTC, the transition zone was not yet wide enough to show up in the composite, although the CP-3/CP-4 Doppler radars did indicate a well-defined transition zone before this time. By 0600 UTC, a transition zone nearly as wide as the convective line was located between the two regions, and by 0730 UTC the gap between the two regions had more than doubled in area. Grid points lying within the different regions were averaged to represent the stratiform (exclusive of the transition zone), convective line, and total system (stratiform + transition + convective line) components of meteorological variables used throughout this study. A separate transition zone component was not computed since that region was narrow and not well defined by the sounding data.

3. Mesoscale flow

The squall line began before 2100 UTC on 10 June as "broken line" convection (Bluestein and Jain 1985) over southwestern Kansas and the Oklahoma Panhandle ahead of a cold front, and later grew to include a transition zone and broad stratiform rain region as it passed through the PRE-STORM network. The convective line reached peak intensity around 0300 UTC, the first composite time considered in this study. After 0300 UTC, the low-level radar reflectivities in the leading line decreased, while the stratiform region expanded in size through 0600 UTC, the second budget study time. After 0600 UTC, all portions of the system began to weaken markedly, and the final budget study was done during this time at 0730 UTC. A detailed description of the squall line history and synoptic setting can be found in Johnson and Hamilton (1988).

a. Relative winds

The system-relative airflow normal to the 10–11 June squall line consisted of a series of three well-developed mesoscale jets (Fig. 3.). The velocities shown in the figure are the along-line averages of all the grid points within 50-km wide strips at given distances away from the leading edge of the radar echo (analogous to Ogura and Liou 1980). Because the values are averages, the magnitudes are less than those obtained for the same system from Doppler radar (Smull and Houze 1987b). However, some grid point values within regions of strong flow were approximately equal to those found from Doppler data. A strong front-to-rear jet carried heat and moisture rearward from low levels ahead of the convective line to high levels through the trailing anvil. A rear inflow jet entered the stratiform region at midlevels and descended toward the convective line where it eventually reached to within 1 km of the surface. Another front-to-rear jet, the result of an overturning downdraft fed by divergence beneath the stratiform region, cut under the rear inflow and had peak intensities within 100 mb of the
surface. These jets have also been documented for this case by Augustine and Zipser (1987), Smull and Houze (1987b), Johnson and Hamilton (1988) and Rutledge et al. (1988), using profiler and Doppler radar data in addition to the rawinsonde data.

The rear inflow jet appeared to strengthen from 0300–0600 UTC, and then remained strong through 0730 UTC. The weakness of the jet at 0300 UTC may be due in part to the proximity of the rear of the system to the western boundary of the data network. At the later times, the jet descended from around 400 mb over 100 km to the rear of the back edge of the radar echo, to 900 mb within the convective line. The forward edge of the rear inflow appeared to advance toward the gust front by 0730 UTC. This may simply be due to inadequacies of the rawinsonde data. Doppler radar data shows (Smull and Houze, 1987; Rutledge et al., 1988) that the jet was already into the convective line by 0300 UTC.

A strong front-to-rear jet was present throughout the times studied in this paper. Peak velocities were located around the 200-mb level at or just behind the back

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**Fig. 3.** Component of the relative wind (in m s⁻¹), normal to the squall line for (a) 0300, (b) 0600, and (c) 0730 UTC. Positive numbers indicate rear-to-front flow (left to right). Approximate locations of convective region (darkest shading), stratiform region (lighter shading), and transition zone (no shading) are shown in the bar below each figure.
edge of the radar echo near the top of the mesoscale anvil. These velocities decreased slightly as the system weakened. Rear-to-front flow ahead of the convective line at high levels cannot be seen in this figure, probably because this system had only a very small leading anvil (Smull and Houze 1978b; Rutledge et al. 1988). A region of weak front-to-rear flow can be found instead around the 150-mb level at 0300 UTC, and the 250-mb level at 0600 and 0730 UTC. Front-to-rear flow near the surface remained strong with only a slight weakening from 0300–0730 UTC.

The along-line component of the relative winds is shown in Fig. 4. Southwesterly flow was dominant except in the vicinity of the rear inflow jet. Northeast flow there increased in areal coverage and intensity from 0300–0730 UTC, although speeds remained under 10 m s⁻¹. At high levels, the southwesterly flow was strong, with speeds exceeding 35 m s⁻¹ at 0300 UTC. This flow weakened somewhat as the system dissipated. The along-line component of the flow will be used later to offer a possible explanation for asymmetries in the precipitation patterns of the system.

b. Vertical velocities

Vertical p-velocities can be seen in Fig. 5. This figure illustrates the weakening of upward motion as the squall line decayed. Upward velocities decreased from a peak of −45 μb s⁻¹ (85 cm s⁻¹) at 400 mb at 0300 UTC to only −25 μb s⁻¹ (45 cm s⁻¹) at 450 mb by 0600 UTC. The upward motion at 0300 UTC was more intense than that found in Ogura and Liou (1980), −30 μb s⁻¹, but the 0600 UTC values were closer. The small values of upward motion at 0730 UTC are due partly to the proximity of the convective line to the edge of the grid.

Within the stratiform region, strong convergence (inferred from the ω field) was occurring at midlevels with a mesoscale updraft in the anvil above it and a downdraft below. Because of the resolution afforded by rawin sondes, some aliasing of convective motion into the stratiform region occurs in the analyses. The aliasing results in the appearance in Fig. 5 of just two main regions of differing vertical velocity (especially at 0600 and 0730 UTC): a large area of upward motion near the leading edge of the system and a similarly large area of downward motion centered near the back edge of the radar echo. The aliasing of strong upward motion into the region ahead of the line will later be shown to cause errors in diagnosed precipitation rates. Although upward motion can be found in the trailing anvil above the stratiform region, a distinct maximum as seen in Ogura and Liou (1980) is not obvious until 0730 UTC. At that time, there appears to be a well-defined peak in upward motion over the stratiform region, but unfortunately the convective line had left the data network and a direct comparison with Ogura and Liou cannot be made. The upward velocity minimum over the transition zone, detected with Doppler radar (Rutledge et al. 1988) around 0300 UTC, cannot be resolved because of its small scale (a few tens of kilometers). A small region of subsidence can be seen at the top of the anvil at 0600 UTC, near the back edge of the radar echo, and just behind it. This downward motion is discussed in more detail in Johnson et al. (1990), and has also been found in tropical convection from profiler data (Balsley et al. 1988).

The area covered by downward motion at low levels in the stratiform region expanded with time, although maximum amplitude was reached at 0600 UTC (Fig. 5b) when the stratiform precipitation was most intense (radar reflectivities were greatest). Peak descent at that time was over 23 μb s⁻¹ (30 cm s⁻¹) near 600 mb. Rutledge et al. (1988), using Doppler radar data, found peak descent of over 60 cm s⁻¹ in this system between 0300 and 0500 UTC. Locally, descent as determined from rawin sonde data approached this value. They found that the level of peak descent varied over time and space, but generally occurred from just above the melting level to roughly 1.5 km below it. The present results agree, and show that the level of peak descent rose to just above the melting level (about 625 mb) as the system weakened (Fig. 5b). By 0730 UTC, peak descent was occurring in a broad layer (~500–700 mb). A weakening of the mesoscale downdraft was observed from 0600–0730 UTC, consistent with a decrease in surface rain rates in the stratiform region during this time (shown later in Fig. 18).

Another pronounced change that took place over time was the increasing slope of the axes of maximum positive and negative vertical motion (Fig. 5). At 0300 UTC, the system was nearly vertical, with the peak updraft at high levels within 75 km of the leading edge of the radar echo. By 0730 UTC, the peak updraft had shifted to 125 km behind the leading edge. The axis of peak downward motion shows the increasing slope most dramatically. The increasing slant of the system is not necessarily an indication of an increasing tilt of the convective line elements, (e.g., Rotunno et al. 1988), but rather is a reflection of a broadening of the system with the convective line separating from the trailing stratiform cloud as in Zipser (1988).

Although the 10–11 June squall line was highly two-dimensional, there were some along-line variations in the vertical velocity patterns over the PRE-STORM domain. Figures 6 and 7 illustrate some of these variations at the approximate level of maximum downward motion in the stratiform region (700 mb; Fig. 6) and the level of maximum upward motion within the system (400 mb; Fig. 7). Although the strong upward motion toward the front of the system and the subsidence behind it lay in two prominent line-parallel bands at 700 mb, the bands did show some structure. Two downward motion centers to the rear of the line existed at all three times (Fig. 6) along the back edge of the stratiform precipitation region. The two centers
Fig. 6. Vertical velocity ($\omega$) at 700 mb in the PRE-STORM region for (a) 0300, (b) 0600 and (c) 0730 UTC. Values in $\mu$b s$^{-1}$ with 1
$\mu$b s$^{-1}$ = 1.2 cm s$^{-1}$. Composited radar echoes shown by shading. Darkest shading is the convective region, large area of light shading the
stratiform region, and a region of no shading between them in (b) and (c) the transition zone. Small areas of light shading ahead of the line
at 0300 UTC, and behind it at 0600 and 0730 UTC are regions of scattered light showers.

Fig. 7. As in Fig. 6, except at 400 mb. 1 $\mu$b s$^{-1}$ = 1.8 cm s$^{-1}$.
support the arguments presented in Johnson and Hamilton (1988) linking the two wake lows in this squall line to two suggested enhanced regions of subsidence at the back edge of the trailing stratiform region. These two features were also associated with two stratiform rainfall maxima (Fig. 18, later). At 400 mb (Fig. 7), a band of upward motion was still evident centered in the stratiform region, but the downward motion no longer occupied a line-parallel band, but rather a band perpendicular to the squall line throughout the period.

The axis of strong downward motion at high levels extended from behind the squall line into a "notch" in the radar echo at the back edge of the stratiform region (Fig. 7). A "split" in the cloud shield was also located in this region (Rutledge et al. 1988). The downward motion can be seen as early as 0300 UTC, but becomes more well defined at later times. The band was roughly 100-km wide, and descent at 400 mb reached 22 mb s\(^{-1}\) (40 cm s\(^{-1}\)) at 0730 UTC. This band was also aligned toward the direction in which the leading edge of convection had "bowed." Small and Houze (1985) found in another midlatitude squall line that the notch and bowing were aligned in the zone of maximum rear inflow at midlevels. It is possible that the current of sinking air coming into the anvil from the postsquall environment in this case contributed to evaporation of cloud droplets and the resulting notch in the echo. The bowing of the echo has been attributed to the development of new convection ahead of the system as low-level convergence is enhanced when the rear inflow jet reaches the leading edge of convection (Zhang and Gao 1989).

4. Mesoscale heat and moisture budgets

a. Heat budget

A vertical cross section of the apparent heat source, \(Q_1\), across the squall line can be seen in Fig. 8. Heating or cooling is primarily a function of the net condensational heating, and the horizontal and vertical convergences of eddy heat fluxes, although phase changes involving ice and radiative effects near cloud top may play some role. The weakening of the system is evident as the peak heating rates decreased from over 13°C h\(^{-1}\) at 0300 UTC to less than 5°C h\(^{-1}\) by 0730 UTC. The level of peak heating remains high, around 400 mb, throughout the time period of this study. Cooling due to evaporation, sublimation, and to a lesser extent, melting (the 0°C-level is near 650 mb) is evident at all levels toward the rear of the system, with peak intensity of over 6°C h\(^{-1}\) near 550 mb at the back edge of the radar echo at 0600 UTC when the stratiform rain was most intense. Leary and Houze (1979) found in a study of five GATE convective systems that cooling due to melting of ice within a 1-km layer was similar in magnitude to cooling from evaporation of raindrops below the melting layer, with both processes able to account for cooling rates as great as 6°C h\(^{-1}\) or more. Secondary cooling maxima near and below 650 mb at 0300 and 0600 UTC (Figs. 8a,b) may be associated with this melting. Weak warming found ahead of the system at 0300 UTC, and behind it at 0600 and 0730 UTC is associated with condensation taking place within weak showers. Above 200 mb just to the rear of the system some enhanced cooling can be seen. This cooling is located near the top of the extensive anvil cloud which extended beyond the back edge of the radar echo, and this cooling may be an indication of radiative cooling at the top of the deep stratiform cloud in this nighttime situation (Webster and Stephens 1980).

In Fig. 9, the apparent heat source normalized by the rain rates averaged over the entire system (a) and the convective line portion (b) for 0300 and 0600 UTC is compared with several tropical and subtropical cases. In both Figs. 9a,b, the vertical distribution of heating did not change significantly between 0300 and 0600 UTC, although the magnitude of the heating aloft decreased as the system weakened (Fig. 8). For the entire system (Fig. 9a), the normalized rates on 11 June were in reasonable agreement with the other cases, except in the lower troposphere where cooling predominated in this case. The upper-level heating peak is near 400 mb, which is close to the levels of peak heating found in the subtropics (curve J; Johnson 1976) and the tropical western Pacific (curves R and Y; Reed and Recker 1971; Yanai et al. 1973). As has been discussed elsewhere (Thompson et al. 1979), the eastern Atlantic results (curve T) show a lower level peak. The low-level cooling was due primarily to the large evaporation rates occurring within the broad mesoscale downdraft over the extensive area experiencing stratiform rainfall, although cooling also occurred at very low levels within the leading convective line (Fig. 9b). Low-level cooling was found in the midlatitude studies of Ninomiya (1971), Lewis (1975) and in the lowest 1 km of the squall line simulation by LaFore et al. (1988), and was linked to the evaporation of rain below cloud base.

In Fig. 9b the normalized heating rates within the convective line region can be seen to agree with another observational study (curve GH; Gamache and Houze 1985), but fall somewhat short of the magnitudes determined in studies by Johnson (1984, curve J) and Houze (1982, curve H). Johnson solved for the convective line heating as a residual after determining \(Q_1\) for the entire region and the stratiform area and Houze (1982) estimated heating from a simple cumulus model. The differences between curves J and H, and 0300, 0600 and GH may indicate the relative error in estimating true upward motion in the convective line due to inadequate rawinsonde coverage. Indeed it will be shown later that the convective line rainfall computed from the integrated heat budget underestimates the observed rainfall by ~40%. The magnitude of heating on 10–11 June agrees with the tropical study
by Gamache and Houze (1985), but the heating peak is at a higher level. In addition, no cooling was found in the lowest levels by Gamache and Houze, perhaps as a result of the lower cloud bases and reduced evaporation in tropical oceanic environments.

Unnormalized heating rates in the stratiform region are presented in Fig. 10. The magnitude of the heating within the trailing anvil decreased sharply as the system weakened, and the level of peak heating rose slightly from 450 mb at 0300 UTC to 400 mb at 0730 UTC. These levels are ~50–100 mb lower than the levels of peak heating found in a study of Winter MONEX anvils by Johnson and Young (1983). The difference is probably due to a higher tropopause in the tropics than in midlatitudes. At lower levels, the layer experiencing cooling associated with evaporation and to a lesser extent melting in the mesoscale downdraft deepened as the system dissipated, with cooling extending from the surface to near 500 mb by 0730 UTC. This agrees with the vertical motion field (Fig. 5) that showed downward motion averaged over the stratiform region was also occurring in a deeper layer over time. By 0730

Fig. 8. Apparent heat source $Q_h$ (in °C h$^{-1}$) at (a) 0300, (b) 0600, and (c) 0730 UTC (shading as in Fig. 3).
Fig. 9. Comparison of vertical $Q_1$ profiles normalized by rainfall rates for averages taken (a) over the entire system and (b) over the convective line region for 11 June at 0300 (curve A) and 0600 (curve B) UTC. In (a), other curves (from Johnson 1984) are from tropical or subtropical cases by Yanai et al. (1973, curve Y), Reed and Recker (1971, curve R), Thompson et al. (1979, curve T), and Johnson (1976, curve J). In (b), other curves (after Houze 1989) include an estimate from a simple cumulus model (Houze 1982, curve H), a residual inferred from rawinsonde data and stratiform region profiles (Johnson 1984, curve J), and a computation using diagnosed vertical velocities (Gamache and Houze 1985, curve GH).
UTC, peak cooling was located at the melting level, perhaps in part because the rear inflow and influx of dry air had weakened below this level (Fig. 3).

The horizontal distribution of $Q_1$ at different levels within the squall line can be seen in Fig. 11. This figure shows variations in the apparent heat source at 0600 UTC (the budget time at which the system was centered best within the network) at 850 (a), 700 (b), 500 (c) and 300 mb (d). Peak heating rates gradually shifted rearward with height with the greatest values at 500 mb (Fig. 11c). Peak cooling rates also shifted rearward, but generally were found near the back edge of the radar echo. Cooling was also greatest at 500 mb. The cooling was enhanced in two regions which corresponded to the downward motion maxima discussed earlier. Except for the highest level, 300 mb, the cooling and heating were both found in line-parallel bands. At 300 mb, however, the cooling appeared to be concentrated in two bands normal to the line. The southern band was associated with strong downward motion at high levels. This band was oriented toward the notch and bow in the radar echo and breach in the cloud canopy, and seems to indicate that enhanced evaporation and sublimation were taking place within this portion of the mesoscale anvil. It is possible that the downward motion and evaporation here were greater, and had been occurring over a longer period of time than in the band farther north resulting in a notch near the southern band but not near the northern one.

b. Moisture budget

The apparent moisture sink, $Q_2$, can be seen in Fig. 12. Because shallow dry layers were occasionally present in certain regions of the data network, the field of $Q_2$ is noisier than that of $Q_1$. Moisture varied in the vertical far more than temperature. Condensational drying can be seen with peak values near the front edge of the radar echo. A hint of a double peak structure can be seen in the figure, especially at 0300 UTC. This double peak has been found in other moisture budgets (e.g., Johnson 1976; Yanai et al. 1973) and it shows up when $Q_2$ is averaged over an entire system. Johnson (1984) has explained the double-peak profile in budget studies involving composites of many cases as a result of two different drying processes, cumulus drying at low levels in the convective line, and drying due to condensation in the anvils at high levels. LaFore et al. (1988) argue based on the numerical simulation of a squall line that the double-peak structure is a result of midtropospheric moistening by vertical convergence of the water-vapor flux in the convective line. From the computations in this study, the specific cause of the double-peak structure in this case cannot be ascertained: either mechanism or a combination of both may be operating.

Moistening due to evaporation can be found at the rear of the system (Fig. 12), in generally the same region that was experiencing cooling in Fig. 8, but the $Q_2$ field appears somewhat more variable than the $Q_1$ field. This again may be due to the highly variable pattern of humidity found across the region. Another significant difference between the $Q_1$ and $Q_2$ fields is that the axis of highest $Q_2$ is displaced toward the front of the system compared to the axis of highest $Q_1$. The displacement reaches a maximum of 50 km at 0600 UTC. The forward displacement of the low-level $Q_2$ maxima ahead of the heavy precipitation may, in part, be explained by the likely existence of growing cumulus towers and deep cumulonimbus updrafts ahead of the region of precipitation fallout (dark shaded area) and the tendency for $Q_2$ to maximize at the lowest levels. Also, since the gradient of moisture is greatest at a lower level than that of dry static energy, $Q_2$ is more susceptible to aliasing errors caused by the smoothing of $\omega$ ahead of the system at low levels.

The normalized $Q_2$ profiles for the entire system at 0300 and 0600 UTC are shown in Fig. 13. The areas under the curves at 0300 and 0600 UTC (A and B) appear comparable to those under the other curves, as they should be if the budgets and rainfall estimates are reasonably accurate (Johnson 1984). The $Q_2$ maxima in the present case (between 525 and 575 mb) are near the level of the weaker upper peaks for the cases having pronounced double-peak profiles. The other cases had primary maxima below 700 mb. A possible explanation for the difference in the $Q_2$ profiles is the existence of pronounced evaporative moistening (negative $Q_2$) occurring at midlatitudes, in contrast to the tropics, such that there is a cancellation of the low-level peak associated with drying in the leading convective line.

Horizontal variations in the $Q_2$ field can be seen in
Fig. 11. $Q_i$ (in °C h$^{-1}$) at 0600 UTC in the PRE-STORM region at (a) 850, (b) 700, (c) 500, and (d) 300 mb.
Fig. 14 at four different levels at 0600 UTC. The region of most intense drying, which was located just ahead of the leading edge of radar echo at 850 mb (Fig. 14a), shifted rearward into the convective line at 700 mb (Fig. 14b). By 500 mb (Fig. 14c), peak drying rates were found within the stratiform region. Unlike the field of $Q_2$ shown in Fig. 11, $Q_2$ becomes very small at high levels (300 mb; Fig. 14d) because specific humidity is very small there. Peak values of drying are found at 850 mb (Fig. 14a), with peak moistening at 700 mb within and to the rear of the stratiform region (Fig. 14b).

c. Total heat budget

The vertical eddy flux of total heat, $F$, for the entire system is shown in Fig. 15. Because $S_0$ and $L_v E_0$ are assumed to be small in (6), and the averaging is done for the entire system, $F$ should approach zero at the surface. At 0300 UTC, this was the case, but at 0600 UTC, the value of the vertical flux was still large (~500 W m$^{-2}$) at the surface. The anomalously high surface value of this quantity can probably be attributed to the forward displacement of $Q_2$ at low levels due to aliasing (which was largest at 0600 UTC). Peak values of the
eddy flux were found around 500 mb at 0300 UTC (2250 W m$^{-2}$) and at 700 mb at 0600 UTC (1350 W m$^{-2}$), clearly indicating the weakening of the system. The greatest contributions to $F$ came from the convective region. These values are significantly larger than the 400 W m$^{-2}$ values found at the midlevel peaks in budgets done on larger spatial and temporal scales by Yanai et al. (1973) and Esbensen et al. (1988).

d. Comparison of diagnosed and observed rainfall rates

Precipitation rates can be diagnosed by the vertical integration of the heat and moisture budgets, and compared to observed rates as a check on the accuracy of the budgets. Rainfall rates calculated from the heat and moisture budgets, respectively, for the grid points in the PRE-STORM region are shown in Figs. 16 and 17. Positive values represent a diagnosed rain rate and negative values indicate net evaporation in the vertical column with no surface rainfall. Both integrated budgets predicted a decrease in the peak rain rates as the system weakened. Rates exceeding 20 mm h$^{-1}$ at 0300 UTC fell to around 10 mm h$^{-1}$ by 0730 UTC. Both budgets yielded estimates within a few millimeters per hour of each other for the peak values found near the convective line. The axis of heaviest rainfall rates from the integrated $Q_2$ values was displaced forward from that determined from integrated $Q_1$ values. However, the rain termination line, and axis of greatest evaporation do not appear to be displaced significantly in the two budgets.

Actual rainfall rates as determined from the 5-min PAM/SAM mesonet rainfall data can be seen in Fig. 18. A subjective analysis was done at the three budget times, with rainfall rates calculated by taking the accumulated precipitation over 20-min periods centered at the times used in the compositing of the soundings. Peak rates based on the 5-min data were also taken into account in the analysis and were displaced in distance based on the speed and direction of the motion of the storm. Peak rates fell from over 50 mm h$^{-1}$ at 0300 UTC to around 35 mm h$^{-1}$ at later times, and the area within the convective line having over 10 mm h$^{-1}$ of rain (shaded in Fig. 18) decreased steadily over time. The area experiencing stratiform rainfall (generally less than 6 mm h$^{-1}$) increased from 0300–0600
Fig. 14. $Q_2$ (in °C h$^{-1}$) at 0600 UTC in the PRE-STORM region at (a) 850, (b) 700, (c) 500, and (d) 300 mb.
UTC and then remained constant through 0730 UTC.

Rawin sondes cannot resolve the small-scale structure of rain rate variations within the system, and therefore comparisons between diagnosed and predicted rates are best made using averages taken over larger regions. Some comparisons can be made, however, between the observed and diagnosed fields over the PRE-STORM region. Both budget integrations at 0300 UTC diagnosed peak rainfall rates in the southern part of the convective line (Figs. 16 and 17). This agreed with observations, although the diagnosed rates in the convective line were only 40% of the observed due to the poor ability of the rawinsonde network to resolve convective-scale processes. By 0600 UTC, both budgets diagnosed two primary regions of enhanced rainfall, separated by an intervening zone of rather light rain along the convective line. The observations also showed this pattern, but the features were about 100 km north of where they were diagnosed to be. The diagnosed zone of relatively light precipitation coincided with a band, shown earlier, of enhanced downward motion at high levels (Fig. 7), cooling (Fig. 11d), and evaporation (Fig. 14d). The 100-km displacement between the diagnosed and observed precipitation minima might be explained by the strong southwest along-line flow at high levels (Fig. 4). Cloud and precipitation particles in this flow (which exceeded 20 m s$^{-1}$ above 325 mb at 0300 and 0600 UTC) would have trajectories that could allow them to travel at least this far as they sublimated. Therefore, it would be reasonable for the observed precipitation minima and cloud breach to lie downstream of the subsidence and sublimation. By 0730 UTC, the budgets did a poor job of showing the rainfall rates within the system, probably because the squall line was leaving the data network.

At both 0300 and 0600 UTC, the budget-diagnosed back edge of rain was located a small distance (usually less than 30 km) ahead of the actual termination line. One exception is in the integrated moisture budget at 0600 UTC (Fig. 17b) where a narrow band aligned with the echo notch in Oklahoma and extending into the transition zone was diagnosed to have no rainfall. This band showed up in the integrated heat budget as a region of relatively light rain rates. The actual rainfall rates seen in Fig. 18 do show a pronounced rainfall minimum in this region with rates falling to less than 2 mm h$^{-1}$. Along the entire rear of the system where light rain was still occurring, the budgets yielded negative rainfall rates. Chen and Zipser (1982) also found negative rain rates within the stratiform region of a GATE squall line. Negative rainfall rates show that a rearward transport of hydrometeors along with other
FIG. 16. Diagnosed rainfall rates (in mm h⁻¹) from the vertically integrated heat (Q₂) budget for (a) 0300, (b) 0600 and (c) 0900 UTC.

FIG. 17. As in Fig. 16, except for the vertically integrated moisture (Q₁) budget.
forms of water storage are important in producing precipitation in the stratiform region (Rutledge and Houze 1987).

Results of a comparison of observed and predicted rainfall rates are shown in Table 1. Rainfall rates have been averaged over different regions of the MCS to reduce the problems of inadequate sounding resolution and large errors at individual grid points. Within the convective line, rates are on average underestimated by 40% at 0300 and 0600 UTC because of inadequate resolution. Within the stratiform region, resolution should not be such a problem since processes operate more on the mesoscale. At 0300 UTC, both budgets yield stratiform rain rates roughly twice those observed.

Fig. 18. Observed rainfall rates (in mm h⁻¹) for the PRE-STORM region using PAM/SAM mesonet data at (a) 0300, (b) 0600 and (c) 0730 UTC.
TABLE 1. Comparison of rainfall rate predictions (in mm h\(^{-1}\)) from the integrated heat and moisture budgets with observations averaged over different regions of the squall line. All regions are shown at 0300 and 0600 UTC with only stratiform region data shown at 0730 UTC since the convective line had left the data network.

<table>
<thead>
<tr>
<th>Region</th>
<th>Time</th>
<th>(\int Q_I) dp (mm h(^{-1}))</th>
<th>(\int Q_I) dp (mm h(^{-1}))</th>
<th>Observed (mm h(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Convective line</td>
<td>0300</td>
<td>9.8</td>
<td>13.5</td>
<td>20.5</td>
</tr>
<tr>
<td></td>
<td>0600</td>
<td>8.5</td>
<td>8.4</td>
<td>12.9</td>
</tr>
<tr>
<td>Stratiform</td>
<td>0300</td>
<td>8.1</td>
<td>5.6</td>
<td>3.3</td>
</tr>
<tr>
<td></td>
<td>0600</td>
<td>3.0</td>
<td>1.6</td>
<td>3.3</td>
</tr>
<tr>
<td></td>
<td>0730</td>
<td>0.0</td>
<td>0.0</td>
<td>2.3</td>
</tr>
<tr>
<td>Total system</td>
<td>0300</td>
<td>8.8</td>
<td>8.7</td>
<td>8.5</td>
</tr>
<tr>
<td></td>
<td>0600</td>
<td>5.6</td>
<td>4.4</td>
<td>5.8</td>
</tr>
</tbody>
</table>

At least part of this overestimation is due to aliasing of convective line vertical velocity into the stratiform region since the system was relatively narrow at this time. By 0600 UTC, both the heat and moisture budgets underestimated the stratiform rain rates, by 10% and 50% respectively. An underestimation also occurred at 0730 UTC, when the integration of both budgets resulted in zero rain rates (the subset of grid points used in averaging at this time extended to the back edge of the radar echo, but not quite to the front edge of the stratiform region). The actual rate was just over 2 mm h\(^{-1}\); however, at this time, the nearness of the system to the edge of the data network may have caused problems.

The discrepancies in the 0600 and 0730 UTC stratiform rain rates can be explained in part by the transport of ice from the convective line into the anvil (Smith and Houze 1987a). This transport has been estimated to provide over 60% of the condensate within the mesoscale anvil cloud for some systems (Gamache and Houze 1983). Precipitation content, \(M\), (kilograms of ice per cubic meter of air) can be estimated from radar reflectivity (Smith and Houze 1987a) using

\[
M = 8.0 \times 10^{-6} (4.68Z)^{0.61},
\]  

where \(Z\) is the liquid precipitation radar reflectivity. A crude estimate of the transport of this ice into the stratiform region and the amount that it could add to the rainfall rate, \(P\), can then be made using

\[
P = UMA/p_{water}S,
\]

where \(U\) is the component of the wind normal to the back edge of the convective line, \(A\) is the area in the vertical plane through which the ice is being transported, and \(S\) is the surface area over which the rain produced by the ice may fall out. Because it takes 2–3 hours from the time the ice is transported into the stratiform region to when it falls out as rain, Eq. (9) was solved using average values of wind and reflectivity, taken from the Doppler results of Smith and Houze (1987b) for the system around 0400 UTC. An average value for the mass of water per volume at the rear of the convective line appeared to correspond to an average reflectivity of 25 dBZ and a relative wind of 15 m s\(^{-1}\) was used. The vertical area through which the transport was assumed to take place was 5 km deep and 500 km long. The area of the stratiform region where the rainfall was assumed to be enhanced by this process was roughly 100 km \(\times\) 500 km.

Using these values, the transport of the ice into the stratiform region was found to add 1.8 mm h\(^{-1}\) to the average stratiform rates. This transport would account for all of the differences between the observed and diagnosed rates at 0600 UTC and most of the differences at 0730 UTC. An increase of the average reflectivity used in the calculation to 30 dBZ would increase the rainfall rates by 3.8 mm h\(^{-1}\), more than enough to account for all of the underestimations. An additional transport of ice rearward due to eddy flux within the convective cells may also increase rainfall rates (Smith and Houze 1987a). Likewise, the horizontal convergence of eddy heat and moisture fluxes, primarily in the vicinity of convective towers, may contribute to the heat and vapor budgets and hence the rainfall rates. Finally, storage of liquid water within the cloud prior to the mature stage (~0300 UTC) may result in actual rain rates exceeding those diagnosed from the budgets. Cotton et al. (1989) showed that MSCSs become very efficient rain producers during their later stages with precipitation efficiencies exceeding 100%. Dual-Doppler radar data may permit estimates of these contributions to be made, but no such analysis has been conducted for this study.

Using an average value for the calculated rearward transport of hydrometeors and the observed rainfall rate within the stratiform region, a water budget was done for the stratiform region of this case at 0600 UTC, and a comparison was made between 10–11 June and tropical oceanic studies by Leary and Houze (1980, LH) and Gamache and Houze (1983, GH) and continental studies by Chong and Hauser (1989; CH) and

<table>
<thead>
<tr>
<th>Case</th>
<th>(R_m = C_{mo} - C_{e_m} - C_{e_m} + C_s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LH-A</td>
<td>0.40 (0 \quad -0 \quad -0.04) +0.44</td>
</tr>
<tr>
<td>LH-B</td>
<td>0.40 (0 \quad -0.32 \quad -0.08) +0.80</td>
</tr>
<tr>
<td>LH-C</td>
<td>0.40 (0 \quad -0.32 \quad -0.08) +0.40</td>
</tr>
<tr>
<td>GH-I</td>
<td>0.49 (0.35 \quad -0.16 \quad -0.12) +0.42</td>
</tr>
<tr>
<td>GH-II</td>
<td>0.49 (0.26 \quad -0.16 \quad -0.12) +0.51</td>
</tr>
<tr>
<td>CH</td>
<td>0.29 (0.30 \quad -0.21 \quad -0.09) +0.26</td>
</tr>
<tr>
<td>RJ</td>
<td>0.23 (0.24 \quad -0.10 \quad -0.04) +0.14</td>
</tr>
<tr>
<td>10–11 June</td>
<td>0.30 (0.30 \quad -0.15 \quad -0.11) +0.26</td>
</tr>
</tbody>
</table>
Roux and Ju (1990, hereafter RJ) (Table 2). The numbers in the table indicate the relative sizes of the water budget parameters as a fraction of the total rainfall (as in Gamache and Houze 1983), where the water budget equation (Houze et al. 1980) is

\[ R_m = C_{mu} - E_{md} - E_{me} + C_A, \tag{10} \]

with \( R_m \) the rainfall at the surface in the stratiform region, \( C_{mu} \) the condensate produced by the mesoscale updraft, \( E_{md} \) the condensate evaporating in the mesoscale downdraft, \( E_{me} \) the condensate advected out of the cloud and evaporating in the environment, and \( C_A \) the condensate carried from the convective line into the stratiform region. The cases from LH and GH are explained in detail in Gamache and Houze (1983). In this study, \( R_m, C_{mu}, E_{md}, \) and \( C_A \) were derived from surface, radar, and sounding data with \( E_{me} \) solved for as a residual.

In general, the values of the water budget components for this case are similar to case LH-C (both mesoscale updraft and downdraft present) and the other cases. The agreement with GH is good, even though in that study, \( C_A \) was solved for as a residual, and \( E_{me} \) was derived. The best agreement occurs with the COPT (COnvection Profonde Tropicale) cases (RJ and especially CH) where all budget terms were retrieved from Doppler radar data. The fraction of rainfall occurring in the stratiform region was roughly 50% on 10–11 June (Johnson and Hamilton 1988) somewhat lower than the 49% values for the cases from GH but similar to CH and RJ. The contribution to surface rainfall from condensate produced in the mesoscale updraft was similar to that due to the transport of condensate rearward from the convective line. The value of \( C_A \) for the 10–11 June case is lower than the values from GH. This may be explained by the fact that the convective line on 10–11 June was rapidly weakening at this time, so that transport of hydrometeors rearward was decreasing. It appears that the proportion of condensate removed through evaporation \( [(E_{md} + E_{me})/(C_{mu} + C_A)] \) was greater for this case and CH than in cases GH and RJ. Precipitation efficiencies \( [R_m/(C_{mu} + C_A)] \) for CH and 11 June were only 50–55%, compared with over 60% for GH and RJ. This is due to a deeper layer of unsaturated air in these cases compared with GH and RJ (Roux and Ju, 1990). Rutledge and Houze (1987) have also done a water budget on a midlatitude squall line using a diagnostic model. They found values of \( C_{mu} \) nearly twice those found in GH with evaporation in the downdraft roughly half of \( C_{mu} \). Although \( C_{mu} \) was nearly twice \( C_A \), they showed that transport of hydrometeors into the stratiform region was necessary for substantial rainfall at the surface. Graupel and snow particles advected rearward were able to feed on the moisture available due to the mesoscale updraft, and the influx of these hydrometeors was necessary before condensate produced in the updraft could be removed and result in surface precipitation.

The predicted rainfall rates for the entire system (Table 1) were generally very close to the observed rates. Except for the moisture budget prediction at 0600 UTC, which was about 25% too low, the other diagnosed rates were within 5% of the actual rates. The especially low estimate of the moisture budget at 0600 UTC may be due to the forward shifting of the Q2 maximum ahead of the line at that time. Kuo and Anthes (1984) found that the moisture budget was a better predictor of rainfall rates in their case than the heat budget. For this study, the heat budget appears to do a bit better, but considering possible sources for error, the results for both are felt to be reasonably good. In particular, the measurement of convective rainfall on a 50-km grid, despite the 5-min data, is subject to error. The accuracy is improved somewhat by system-wide averaging, but the observed precipitation estimates are probably still only good to within 20% or so.

5. Conclusions and discussion

In this study, data from the dense PRE-STORM rawinsonde network were used to perform heat and moisture budgets on the 10–11 June squall line. Composites were made using soundings taken over three 3-h time periods in order to improve resolution of small-scale features. The evolution of the budgets during the mature through dissipating stages of the system was examined, along with spatial variations within the large squall line system, which covered roughly 10^5 km^2.

The changes have been documented within three pronounced mesoscale jets (front-to-rear flow entering the system from ahead of it at low levels, rear inflow at midlevels, and front-to-rear flow undercutting the rear inflow jet) as the system weakens, and it is shown that the rear inflow jet remains strong, even as the system becomes markedly weaker. Vertical velocities also decrease over time, and the axes of strongest upward and downward motion take on an increasing rearward tilt. Downward motion in the lower levels of the stratiform region is most intense when the stratiform rainfall is heaviest. This downward motion weakens when the rain rates decrease, indicating the importance of evaporative cooling in maintaining strong downward motion within the rear inflow jet. Horizontal analyses of vertical motion show two separate maxima of downward motion within the stratiform region, and these maxima are linked with the two weak lows found in the surface pressure field. This finding supports the argument that low-level adiabatic warming produces such pressure minima.

The heat and moisture budgets for this system showed heating and drying due primarily to condensation occurring in the front portion of the system near the convective line and cooling and moistening due to evaporation, and to a lesser extent, melting of ice, at lower levels over the stratiform region. Cooling at the
lowest levels in the convective line distinctly differed from tropical oceanic cases due to a drier lower troposphere over land and, hence, enhanced evaporation. The system-averaged drying profile showed primarily a single upper-level peak, as opposed to the double-peak structure often found within convective systems. The absence of a lower-level peak was probably due to moistening at low-levels within the stratiform region which negated the lower-level drying peak within the convective line.

Precipitation rates diagnosed from the vertical integrations of the budgets showed the importance of hydrometeor transport and storage effects on budget studies done for dissipating midlatitude convection. Underestimates due to inadequate sounding resolution occurred within the convective line where diagnosed rates were roughly 40% of the observed. The diagnosed precipitation rates within the stratiform region were progressively deficient over time, and the underestimations increased as the system dissipated more rapidly. The discrepancy exceeded 2 mm h\(^{-1}\) by 0730 UTC. The rearward transport of hydrometeors into the stratiform region from the convective line was shown to have the ability to add 2–4 mm h\(^{-1}\) to the rain rates. The complete system-averaged rates also were underestimated slightly when the system was dissipating, implying again the importance of water storage. Rainfall produced at earlier times when the system was more vigorous had been stored within the clouds and was falling out at the later times as the updrafts rapidly weakened.

Although the diagnosed rainfall rates differed from observed values, in general there was close agreement, supporting the credibility of the budgets. The location of \(Q_1\) and \(Q_2\) extrema and the predicted area covered by precipitation in the budget integrations also agreed rather well with observations. The closely-spaced rawinsondes of the PRE-STORM network were indeed helpful in allowing better documentation of small-scale space and time variations within the 10–11 June squall line system. However, even after compositing, the spacing was still 80 km, which was not small enough to accurately resolve convective-scale processes. Because of this, vertical velocities within the convective line were greatly underestimated and spread over an unreasonably large area, thus, affecting the heating and drying rates determined for that region of the system. Another problem was the early termination of soundings that ascended within the convective system. Data often ended in the 450–650 mb layer, diminishing the accuracy of analyses at higher levels. These problems should be addressed in the planning for future mesoscale field experiments such as STORM.

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