Effects of Cumulus Ensemble and Mesoscale Stratiform Clouds in Midlatitude Convective Systems

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ABSTRACT

Diagnostic and semiprognostic analyses are performed using OK PRE-STORM (Oklahoma–Kansas Preliminary Regional Experiment for STORM-Central) data to examine the cumulus–environment interaction in midlatitude convective systems. The similarities and differences of the interaction processes between midlatitude and tropical convective systems are also discussed. Analyses of PRE-STORM and GATE (GARP Atlantic Tropical Experiment) data show generally larger vertical wind shear, large-scale forcing, and moist convective instability in midlatitude MCCs (mesoscale convective complexes) and squall lines than in tropical cloud clusters. It is found that the interaction mechanism based on the cumulus-induced subsidence and detrainment is capable of explaining most of the observed heating and drying under widely different environment conditions. Convective-scale downdrafts act to cool and moisten the lower troposphere in the midlatitudes as in the tropics. The quasi-equilibrium assumption between stabilization by convection and destabilization by large-scale forcing is valid and holds better in the midlatitudes since the large-scale forcing is much stronger. Both the cumulus and stratiform cloud effects are stronger in midlatitude than in tropical convective systems.

1. Introduction

The interaction of organized cumulus convection with its large-scale environment has long been recognized as a leading problem in meteorology. Because individual cumulus clouds have dimensions that cannot be resolved by the conventional observational networks or by finite difference grids in numerical models, the collective effects of an ensemble of cumuli must be inferred indirectly from heat and moisture budgets of the large-scale circulations or "parameterized" in terms of the resolvable variables. In the last two decades our understanding of the thermodynamic interaction of organized cumulus convection with its environment has significantly increased through many observational and parameterization studies.

A cumulus ensemble modifies the temperature and moisture of its environment primarily through the subsidence of environmental air, which compensates the convective mass flux and through the detrainment of excess heat and water substance from clouds (Ooyama 1971; Yanai et al. 1973; Arakawa and Schubert 1974). In turn, the environment controls the activity of cumulus clouds through various destabilizing processes. Assuming a "quasi equilibrium" between the generation of moist convective instability by large-scale processes and the stabilizing effects of cumulus convection, Arakawa and Schubert (1974) developed a comprehensive theory of cumulus parameterization. Recently, Cheng (1989a,b) developed a cumulus ensemble model that includes dynamically modeled convective-scale downdrafts. The combined updraft-downdraft cumulus ensemble model has been incorporated into the Arakawa–Schubert cumulus parameterization (Cheng and Arakawa 1990).

Most previous studies of the cumulus–environment interaction have been made in the tropics. There have been relatively few efforts to examine the effects of cumulus convection in midlatitude convective systems. The environment of midlatitude convective systems is substantially different from the environment of tropical convective systems (e.g., Zipser and LeMone 1980; Ogura and Jiang 1985). There is stronger vertical shear of the horizontal wind, a larger horizontal temperature gradient, and less moisture in the midlatitudes than in the tropics. Many observations show that cumulus convection in midlatitudes is also strongly controlled by large-scale processes (e.g., Fankhauser 1969; Sasaki and Lewis 1970; Hudson 1971; Lewis et al. 1974; Ulanski and Garstang 1978).

The large-scale heat and moisture budget studies in midlatitude convective systems (e.g., Ninomiya 1971; Fritsch et al. 1976; Kuo and Anthes 1984; Gallus and Johnson 1991) show the presence of apparent heat sources and moisture sinks and a general agreement among the horizontal distributions of vertically integrated heating and drying, rainfall, and radar reflectivity; however, the vertical distributions of the observed

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heating and drying in midlatitudes (e.g., Kuo and Anthes 1984) differ from those in tropics (e.g., Yanai et al. 1973). The diagnostic study by Lewis (1975) showed that the observed budget residuals can be interpreted from cumulus effects estimated using a diagnostic cumulus ensemble model (Ogura and Cho 1973). He found that deep clouds are dominant in the midlatitudes, in contrast to the existence of both deep and shallow clouds in the tropics (Yanai et al. 1973; Nitta 1978). The semiprognostic tests of the Arakawa–Schubert cumulus parameterization scheme by Grell et al. (1991) show that the quasi-equilibrium assumption is valid in midlatitude convective systems.

Observations during the GARP (Global Atmospheric Research Programme) Atlantic Tropical Experiment (GATE) showed that cumulus clouds are usually organized into cloud clusters such as fast-moving squall clusters and slow-moving nonsquall clusters (e.g., Houze and Betts 1981). The occurrence of squall clusters is closely related to the appearance of the low-level easterly jet (e.g., Frank 1978; Barnes and Sieckman 1984). Midlatitude convection is also frequently organized into mesoscale convective systems such as squall line and mesoscale convective complex (MCC) (e.g., Maddox 1980; Houze and Hobbs 1982).

Mesoscale convective systems consist of a combination of convective and stratiform clouds (e.g., Houze 1977, 1989; Zipser 1977; Johnson and Hamilton 1988; Gallus and Johnson 1991). The heating and moistening profiles obtained in mesoscale anvil clouds (e.g., Houze 1982; Johnson and Young 1983) are considerably different from the previously obtained profiles (e.g., Yanai et al. 1973). There is strong warming due to condensation in the mesoscale updraft in the middle to upper troposphere, and cooling caused by melting and evaporation in the mesoscale downdraft in the lower troposphere. The heating profiles obtained in stratiform clouds of midlatitude convective systems also show warming in the upper troposphere and cooling in the lower troposphere (e.g., Houze 1989; Gallus and Johnson 1991).

Introducing an empirical parameter representing the fraction of rainfall produced by mesoscale anvil cloud, Johnson (1984) attempted to partition the heat and moisture budget residuals into cumulus and mesoscale components. Recently, Cheng and Yanai (1989) developed an objective method to isolate influences of mesoscale updrafts and downdrafts on the large-scale budgets. The contributions from mesoscale circulations and effects of detrainment on the observed heat and moisture budget residuals are estimated by eliminating the dominant effects of cumulus-induced subsidence from the residuals. Then, the mesoscale contributions are objectively isolated from the observed budget residuals with the aid of the cumulus ensemble model. The isolated mesoscale contributions clearly show warming and drying in the upper troposphere and cooling and moistening in the lower troposphere, indicating condensation within the anvil cloud and evaporation of rainwater beneath the anvil, supporting the results obtained by Houze (1982) and Johnson and Young (1983).

Although the important roles of cumulus convection and mesoscale effects on midlatitude convective systems were identified by these observation and budget studies, a more systematic examination of these effects using comprehensive and accurate datasets comparable to those obtained during GATE is desirable. The dataset obtained during OK PRE-STORM (Oklahoma–Kansas Preliminary Regional Experiment for STORM-Central) is suitable for this purpose. The PRE-STORM has an average station separation of ~150 km, which is comparable in horizontal resolution to the GATE.

The specific objectives of this paper are 1) to estimate the contributions of cumulus ensemble and mesoscale stratiform clouds to the large-scale heat and moisture budgets of midlatitude convective system, 2) to compare the vertical transports of heat and moisture between MCCs and squall lines, and 3) to compare the cumulus–environment interaction in tropical and midlatitude convective systems.

2. Budget equations and representation of cumulus effects

a. Large-scale heat and moisture budget equations

Following Yanai et al. (1973), the heat and moisture budget equations suitable for the observational data or numerical models of limited resolutions can be written as

\[
Q_1 = \frac{\partial \bar{s}}{\partial t} + \bar{v} \cdot \nabla \bar{s} + \bar{\omega} \frac{\partial \bar{s}}{\partial p} = Q_R + L(\bar{c} - \bar{c}) - \nabla \cdot \bar{s} \bar{v}' - \frac{\partial}{\partial p} \bar{s} \omega',
\]

\[
Q_2 = -L \left( \frac{\partial \bar{q}}{\partial t} + \bar{v} \cdot \nabla \bar{q} + \bar{\omega} \frac{\partial \bar{q}}{\partial p} \right) = L(\bar{c} - \bar{c}) + L \nabla \cdot \bar{q} \bar{v}' + L \frac{\partial}{\partial p} \bar{q} \omega',
\]

where an overbar (\(\bar{\cdot}\)) denotes resolvable components, that is, the running horizontal average, and a prime (\(\prime\)) expresses unresolved components, that is, the deviation from the horizontal average. \(s = c_pT + gz\) is the dry static energy, \(c_p\) the specific heat of air at constant pressure, \(T\) the temperature, \(g\) the acceleration of gravity, \(q\) the water vapor mixing ratio, \(v\) the horizontal velocity, \(\omega\) the vertical velocity, \(Q_R\) the radiative heating rate, \(L\) the latent heat of condensation, and \(c\) and \(e\) are the rates of condensation and evaporation of cloud water per unit mass of air; \(Q_1\) and \(Q_2\) are the residuals of the heat and moisture budgets of the resolvable motion, respectively, and contain real
sources and sinks as well as terms resulting from unresolved eddies. Therefore, they are called the "apparent" heat source and moisture sink.

From (1) and (2) we obtain

$$Q_1 - Q_2 - Q_R = -\nabla \cdot \vec{h} \vec{w} - \frac{\partial}{\partial p} \frac{\partial \vec{h}}{\partial \omega},$$

where $h = s + Lq$ is the moist static energy; $Q_1 - Q_2 - Q_R$ represents the convergence of the eddy transport of moist static energy.

In deriving (1)-(3), it is assumed that the Reynolds conditions hold with sufficient accuracy (e.g., Kampé de Férriet 1951; Monin and Yaglom 1971; Cotton and Anthes 1989). This may be justified when we imagine a horizontal area that is large enough to contain an ensemble of cumulus clouds, but small enough to be regarded as a fraction of the large-scale system (Yanai et al. 1973). The presence of mesoscale circulations, however, may cause ambiguity in the interpretation of $Q_1$ and $Q_2$ (e.g., Anthes 1977).

Integrating (1) and (2) from $p_T$ (pressure at the cloud top or tropopause) to $p_e$ (pressure at the surface), we obtain

$$\langle Q_1 \rangle = \langle Q_R \rangle + LP + S - \langle \nabla \cdot s \vec{v} \rangle,$$

$$\langle Q_2 \rangle = \langle L (P - E) + L \langle \nabla \cdot q \vec{v} \rangle,$$

where

$$\langle (\cdot) \rangle = \frac{1}{g} \int_{p_T}^{p_e} (\cdot) dp,$$

$P$ is the rate of precipitation, and $S$ and $E$ are the rates of sensible heat flux and evaporation from the surface. Note that storage effects (e.g., McNab and Betts 1978; Gallus and Johnson 1991) are neglected in the right-hand sides of (4) and (5). From (4) and (5) we obtain

$$\langle Q_1 \rangle - \langle Q_2 \rangle = \langle Q_R \rangle + LE - \langle \nabla \cdot h \vec{w} \rangle.$$  

With the neglect of horizontal eddy flux terms, (1)-(6) were widely used to measure the collective effects of subgrid-scale processes such as cumulus convection (e.g., Ninomiya 1968, 1971; Nitta 1972, 1975, 1977; Yanai et al. 1973, 1976; Nitta and Esbensen 1974; Luo and Yanai 1984; Kuo and Anthes 1984; Gallus and Johnson 1991; Wu and Yanai 1991b). The justification of ignoring horizontal eddy flux terms is that the net lateral transports across the boundary of the fixed area by cumulus convection are negligible compared to the horizontal transports by the large-scale motion (Arakawa and Schubert 1974). We cannot, however, rule out possible contributions from mesoscale eddies (Wu 1992).

b. Expressions for cumulus ensemble effects

As has been shown in many studies (e.g., Ooyama 1971; Yanai et al. 1973; Arakawa and Schubert 1974; Cheng and Yanai 1989), the contributions of cumulus clouds (condensation, evaporation, and convective transports) to the environment are given by

$$Q_{1c} = \delta (s_D - \bar{s} - Ll_D) - M_c \frac{\partial \bar{s}}{\partial p},$$

$$Q_{2c} = -L \delta (q_D - \bar{q} + l_D) + LM_c \frac{\partial \bar{q}}{\partial p},$$

where $M_c$ is the cloud mass flux, $\delta$ the total mass detrainment from clouds, and $l$ the mixing ratio of liquid water. The subscript $(\cdot)_D$ expresses the value in the detrainment air. The first terms on the right-hand sides of (7) and (8) express the effects of excess heat and moisture detrainment from cumulus clouds and evaporation of detrainment cloud water. The second terms represent the vertical advection of heat and moisture by the part of the environmental vertical motion that compensates the convective mass flux. The reader is referred to Cheng (1989a, b) for a detailed description of the inclusion of convective-scale downdrafts in (7) and (8).

The total mass flux of a cumulus ensemble at a given level is represented by

$$M_c = \int_{0}^{\lambda(p)} \frac{\partial p}{\partial p} m_B(\lambda) \eta(p, \lambda) d\lambda,$$

where $\lambda$ is a constant fractional rate of entrainment, which is assumed to characterize the statistical features of all clouds in a subensemble (Arakawa and Schubert 1974); $m_B(\lambda) d\lambda$ is the cloud-base mass flux of the subensemble due to the clouds whose fractional rate of entrainment are between $\lambda$ and $\lambda + d\lambda$, $\eta(p, \lambda)$ is the mass flux normalized by $m_B(\lambda)$, and $\lambda_D(p)$ is the $\lambda$ of clouds that have the detrainment level at pressure $p$.

The total mass detrainment from the clouds in the layer between $p$ and $p + dp$ can be expressed by

$$\delta(p) = m_B(\lambda_D(p)) \eta(p, \lambda_D(p)) \frac{d\lambda_D(p)}{dp},$$

that is, the mass flux of the subensemble due to clouds whose $\lambda$ is between $\lambda_D(p)$ and $\lambda_D(p) + [d\lambda_D(p)/dp] dp$.

Here, we have assumed that the detrainment occurs only at the top of clouds in each subensemble.

The reader is referred to Cheng (1989b) for a detailed description of the determination of cumulus ensemble properties in (7) and (8). In the following, two methods are described to obtain $m_B(\lambda)$.

c. Diagnostic method to obtain cumulus ensemble effects

Following Cheng and Yanai (1989), we assume that the observed $Q_1 - Q_2$ and $Q_3$ are contributed from the effects of cumulus convection ($Q_{1c}$ and $Q_{2c}$) and the mesoscale circulations ($Q_{1m}$ and $Q_{2m}$), that is,
\[ Q_1 - Q_R = Q_{1c} + Q_{1m}, \quad (11) \]
\[ Q_2 = Q_{2c} + Q_{2m}. \quad (12) \]

Considering the stratiform character of the mesoscale cloud systems, we assume
\[ Q_{1m} \approx L(e_m - e_m), \quad (13) \]
where \( e_m \) and \( e_m \) are the condensation and evaporation rates associated with either mesoscale updrafts or downdrafts. Therefore, from (11) and (12) we have
\[ Q_1 - Q_2 - Q_R = Q_{1c} - Q_{2c}. \quad (14) \]

Using (7), (8), (9), and (10), we can rewrite (14) as
\[ Q_1 - Q_2 - Q_R = (h_D - h) m_B(\lambda_D(p)) \eta(p, \lambda_D(p)) \frac{d \lambda_D(p)}{dp} \]
\[ - \frac{\partial}{\partial p} \int_0^{\lambda_D(p)} m_B(\lambda) \eta(p, \lambda) d\lambda. \quad (15) \]

Equation (15) is an integral equation with respect to \( m_B(\lambda) \), originally derived by Nitta (1975). Once \( m_B(\lambda) \) is obtained by solving (15), \( M_c(p) \) and \( \delta(p) \) are obtained from (9) and (10), respectively. Then \( Q_{1c} \) and \( Q_{2c} \) are respectively determined from (7) and (8). Note that (15) uses only a combined quantity, \( Q_1 - Q_2 - Q_R \), as input. The diagnosed \( Q_{1c} \) and \( Q_{2c} \) will not generally agree with the observed \( Q_1 - Q_R \) and \( Q_2 \). Only when the cumulus contributions \( Q_{1c} \) and \( Q_{2c} \) are accurately obtained and \( Q_{1m} \) and \( Q_{2m} \) are small, can we reproduce the observed \( Q_1 - Q_R \) and \( Q_2 \) separately.

**d. Semiprognostic method to obtain cumulus ensemble effects**

The semiprognostic approach proposed by Lord (1982) is a one-step prediction of cumulus ensemble effects for a given large-scale condition. In contrast to the diagnostic method, the observed \( Q_1 \) and \( Q_2 \) are not used to determine the cloud base mass flux. Instead, a relationship derived from the "cloud work function quasi equilibrium" (Arakawa and Schubert 1974) is used to determine the cloud base mass flux. The cloud work function is defined by the kinetic energy generation per unit \( m_B(\lambda) \) by the thermal buoyancy, that is,
\[ A(\lambda) = \int_{p_T}^{p_B} \frac{1}{\rho c_p T(p)} \eta(p, \lambda) [s_{sc}(p, \lambda) - \bar{s}_v(p)] dp, \quad (16) \]
where \( s_{sc} \) and \( \bar{s}_v \) are the virtual static energy in the clouds of a cumulus subensemble and the environment, respectively; \( T \) is the environmental temperature; \( p_B \) and \( p_T \) are the pressures of cloud base and cloud top, respectively.

Assuming that the destabilization due to large-scale thermodynamic processes is nearly balanced by the stabilizing effect of cumulus convection, we have
\[ \frac{dA(\lambda)}{dt} = \left[ \frac{dA(\lambda)}{dt} \right]_C + \left[ \frac{dA(\lambda)}{dt} \right]_{LS} = 0, \quad (17) \]
where the subscript \( C \) and \( LS \) represent the cloud effects and large-scale effects, respectively.

From (17), an integral equation governing \( m_B(\lambda) \) can be obtained in the form
\[ \int_0^{\lambda_{\text{max}}} K(\lambda, \lambda') m_B(\lambda') d\lambda' + F(\lambda) = 0, \quad (18) \]
where \( \lambda_{\text{max}} \) is the maximum value of \( \lambda \) and
\[ F(\lambda) = \left[ \frac{dA(\lambda)}{dt} \right]_{LS}. \quad (19) \]

denotes the large-scale forcing. The mass flux kernel \( K(\lambda, \lambda') \) expresses the rate of increase of the cloud work function for type \( \lambda \) cloud through the modification of the environment by type \( \lambda' \) cloud, per unit \( m_B(\lambda) \) (Arakawa and Schubert 1974).

Following Lord (1982), the large-scale forcing for each cloud type is calculated from
\[ F(\lambda) = \frac{A'(\lambda) - A(\lambda)}{\Delta t'}. \quad (20) \]

In (20) \( A(\lambda) \) is calculated from the temperature and moisture fields at a given observation time, \( A'(\lambda) \) is obtained from the fields modified by the large-scale thermodynamic processes such as the horizontal and vertical advection and radiative heating over a time \( \Delta t' \).

The mass flux kernel \( K(\lambda, \lambda') \) is calculated from
\[ K(\lambda, \lambda') = \frac{A''(\lambda) - A(\lambda)}{m_B(\lambda') \Delta t'}, \quad (21) \]
where \( A''(\lambda) \) is determined from the fields modified through heat and moisture transports by an arbitrarily chosen cloud-base mass flux \( m_B(\lambda') \) of type \( \lambda' \) cloud over a time \( \Delta t'' \). In this study, for both calculations of large-scale forcing and mass flux kernel, all subcloud-layer processes are ignored by assuming that the total effects of subcloud-layer processes balance those of cumulus clouds.

With the \( F(\lambda) \) and \( K(\lambda, \lambda') \) from (20) and (21), the spectrum of cloud-base mass flux \( m_B(\lambda) \) can be obtained from a discretized form of Eq. (18). Then \( M_c(p) \) and \( \delta(p) \) are determined from (9) and (10); \( Q_{1c} \) and \( Q_{2c} \) can be predicted by (7) and (8).

**3. Data**

The observational datasets from two field experiments, GATE in the tropics and PRE-STORM in the midlatitudes, are used in this study. To analyze the
cumulus–environment interactions under various conditions, we selected six major convective events during GATE, which included three nonsquall clusters and three squall clusters (see Table 1). Deep convection and heavy precipitation were associated with each of these events (e.g., Houze and Betts 1981; Sui and Yanai 1986). We also selected four convective events during PRE-STORM, which included MCCs and squall lines (see Table 1). Associated with these convective systems, heavy rainfall, hail, tornadoes, and/or severe wind gusts were observed in the network and strong convective echoes were detected by the radar network (e.g., Johnson and Hamilton 1988; Rutledge et al. 1988; Biggerstaff and Houze 1991; Stumpf et al. 1991). The following sections give more detailed descriptions of these datasets.

### a. GATE

The datasets derived from Phase III of GATE in 1974 are used in this study. The wind, temperature and relative humidity data are based on upper-air soundings from the A/B- and B-array ships (Fig. 1). The average distance between B-array ships is ~150 km. These data were analyzed by Ooyama and Esbensen using a statistical interpolation scheme (Ooyama 1987). The grid system has a resolution of 0.5° lat and 0.5° long within the domain 4°–13°N, 19°–28°W (Fig. 1). Three-hourly data over a 20-day period are interpolated at 19 pressure levels from 991 to 76 mb. Since the analyzed wind data have a horizontal wavelength cutoff at 450 km, the detailed structure of mesoscale circulations in cloud clusters is not resolved explicitly. But circulations on the scale of large cloud clusters can be identified in the data (Esbensen et al. 1982). Therefore, the budget residuals calculated by (1) and (2) using this dataset contain effects of all unresolvable processes with horizontal scales less than 450 km. The \( Q_1 \) and \( Q_2 \) values were calculated at the centers of 1° × 1° grid boxes formed by the thick solid lines shown in Fig. 1. The radiative heating rates \( Q_R \) were calculated by Cox and Griffith (1979) and were given on grid boxes of 0.5° by 0.5° covering the domain shown by dashed lines in Fig. 1. In this study, the \( Q_R \) values are horizontally smoothed and the smoothed \( Q_R \) values are defined at the centers of the 1° × 1° grid boxes. For more detailed descriptions of this data, the reader is referred to Esbensen et al. (1982), Sui and Yanai (1986), and Sui et al. (1989).

### b. OK PRE-STORM

OK PRE-STORM was conducted from 1 May through 27 June 1985 (e.g., Cunning 1986). The data obtained during this field program are considered to be the most extensive for midlatitude mesoscale systems. The rawinsonde data from 15 National Weather Service (NWS) sites and 12 supplemental sites (Fig. 2) were checked by subjective analyses and interpolated to constant pressure levels from 1000 to 100 mb with an interval of 25 mb at three-hour intervals. The infrared satellite imagery and the composite analyses of digitized data obtained from the NWS WSR-57 radars located at Wichita, Kansas, and Oklahoma City, Oklahoma, are used to identify the convective systems. Detailed descriptions of these data were presented by Meifin and Cunning (1985) and Blanchard (1990). The rainfall data were originally collected at 5-minute intervals from the PRE-STORM surface mesonetwork (with spacing between stations about 50 km), which included 42 sites of NCAR PAM (Portable Automated

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**TABLE 1. Mesoscale convective events selected for this study.**

<table>
<thead>
<tr>
<th>Observation time</th>
<th>Type</th>
<th>Data source</th>
</tr>
</thead>
<tbody>
<tr>
<td>1800 UTC 2 Sep 1974</td>
<td>Nonsquall cluster</td>
<td>GATE</td>
</tr>
<tr>
<td>1800 UTC 4 Sep 1974</td>
<td>Squall cluster</td>
<td>GATE</td>
</tr>
<tr>
<td>1500 UTC 5 Sep 1974</td>
<td>Nonsquall cluster</td>
<td>GATE</td>
</tr>
<tr>
<td>2100 UTC 11 Sep 1974</td>
<td>Squall cluster</td>
<td>GATE</td>
</tr>
<tr>
<td>1500 UTC 12 Sep 1974</td>
<td>Squall cluster</td>
<td>GATE</td>
</tr>
<tr>
<td>2100 UTC 13 Sep 1974</td>
<td>Nonsquall cluster</td>
<td>GATE</td>
</tr>
<tr>
<td>0000 UTC 4 Jun 1985</td>
<td>MCC</td>
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</tr>
<tr>
<td>0300 UTC 11 Jun 1985</td>
<td>Squall line</td>
<td>PRE-STORM</td>
</tr>
<tr>
<td>0600–0900 UTC 22 Jun 1985</td>
<td>Squall line</td>
<td>PRE-STORM</td>
</tr>
</tbody>
</table>

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**FIG. 1.** Horizontal grid system for wind, temperature, relative humidity, and radiation data (see text). Also shown are the positions of A/B-array ships (solid circles), B-array ships (open circles), and a C-array ship (open triangle) during Phase III of GATE.
Mesonetwork-II) and 42 sites of NSSL SAM (Surface Automated Mesonetwork).

A version of the Barnes objective analysis scheme (Barnes 1973) supplied by the National Oceanic and Atmospheric Administration/National Severe Storms Laboratory (NOAA/NSSL) is employed to analyze the horizontal wind components, geopotential height, temperature, and water vapor mixing ratio at the surface and on constant pressure levels from 1000 to 100 mb. Since the average station spacing is approximately 150 km, the radius of influence in the objective analysis scheme is chosen to allow ~42% response at a wavelength of 300 km. The grid system used in PRE-STORM is the same as that used in GATE. The data are interpolated onto 15 × 15 horizontal grid points. The gridded data have a horizontal resolution of 0.5 degree in latitude and longitude within the domain 33°–40°N, 94°–101°W. Finally, the data are vertically smoothed and the vertical resolution is reduced to 19 levels from the surface to the 100-mb level with an interval of 50 mb.

4. Large-scale environmental conditions

In general, the environment that leads to development of cumulus convection is very different between the tropics and midlatitudes. The purpose of the comparison of tropics and midlatitudes is to reveal the similarities and differences of cumulus–environment interaction processes under widely different environmental conditions, and to test the generality of existing cumulus parameterization schemes. Since the PRE-STORM dataset is comparable to the GATE dataset in both horizontal and vertical resolution, we can compare the large-scale environmental conditions in the tropics and midlatitudes. The similarities and differences of the features of cumulus and mesoscale stratiform cloud effects in the tropical and midlatitude convective systems will be discussed in section 8.

Figures 3a and 3b show scatter diagrams of the mean vertical shear versus the cloud work function for clouds whose tops reach near 200 mb in the convective systems observed during GATE and PRE-STORM, respect-
Fig. 3. Scatter diagrams of the mean vertical shear versus the cloud work function for deep clouds in (a) GATE and (b) PRE-STORM. "O" represents nonsquall clusters or MCCs, "+" represents squall clusters or squall lines.

The cloud work function defined by Arakawa and Schubert (1974) [see Eq. (16)] is a generalized measure of moist convective instability in the large-scale environment. The mean vertical shear is defined by

$$MVS = \frac{1}{\Delta \rho} \int_{p_T}^{p_B} \left| \frac{\partial v}{\partial \rho} \right| dp,$$

where $\Delta \rho = p_B - p_T$, and $p_B$ and $p_T$ are the pressures at the cloud base and top, respectively. The remarkable feature seen in the figures is that the mean vertical shear for the GATE cases has an upper limit $\sim 8 \times 10^{-2}$ m s$^{-1}$ mb$^{-1}$, while the mean vertical shear for the PRE-STORM cases has a lower limit $\sim 4 \times 10^{-2}$ m s$^{-1}$ mb$^{-1}$. The mean vertical shear in MCCs and squall lines of PRE-STORM is generally larger than that in nonsquall clusters of GATE. Due to the existence of the low-level easterly jets, the mean vertical shear in squall clusters is generally larger than that in nonsquall clusters of GATE. The stronger mean shear in the midlatitudes suggests that the dynamic interaction between cumulus convection and the large-scale processes is more significant in the midlatitudes than in the tropics, and results in the well-organized convection seen there in the satellite and radar pictures.

Figures 3a and 3b also show that the values of the cloud work function in both GATE and PRE-STORM are mainly confined between 1000 and 2500 J kg$^{-1}$. This means that the moist convective instability of the environment in both GATE and PRE-STORM is similar despite the difference in temperature and moisture structure between the tropics and midlatitudes. It is also noted that the values of cloud work function for the MCCs in PRE-STORM are larger than those for the nonsquall clusters in GATE, suggesting a more unstable environment for the MCCs than for the nonsquall clusters.

Figures 4a and 4b show scatter diagrams of the mean vertical shear versus the large-scale forcing for clouds reaching near 200 mb in mesoscale convective systems observed during GATE and PRE-STORM, respectively. The large-scale forcing shown in Fig. 4 is computed by taking the change of cloud work function due to the large-scale horizontal and vertical advections of temperature and moisture only. As seen in Fig. 4a, the scatter points for GATE are confined to a small area and the large-scale forcing is less than 10 kJ kg$^{-1}$ day$^{-1}$. However, the PRE-STORM points (Fig. 4b) are widely

Fig. 4. As in Fig. 3 except for the mean vertical shear versus the large-scale forcing for deep clouds in (a) GATE and (b) PRE-STORM.
scattered and the large-scale forcing often exceeds 30 kJ kg$^{-1}$ day$^{-1}$. This means that the destabilization (or the generation of moist convective instability) by the large-scale advective processes in PRE-STORM can be much larger than that in GATE. Therefore, the quasi-equilibrium assumption implies stronger stabilization by the convection in PRE-STORM than in GATE.

5. Activity of cumulus convection diagnosed from observations

The horizontal distributions of the vertically integrated heat source $\langle Q_1 \rangle$ and moisture sink $\langle Q_2 \rangle$ for a MCC observed at 0000 UTC 4 June 1985 are shown in Figs. 5a and 5b. This case has been analyzed in several observational studies (e.g., Smull and Augustine 1989; Stumpf et al. 1991). As can be seen from the figures, large positive values of $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ are over Kansas with a center at 38°N, 97°W. The heating and drying are clearly associated with large-scale divergence at 200 mb (Fig. 5e). Both the wind speed minimum (Fig. 5d) and divergent wind (Fig. 5e) at this level suggest the presence of a vertical mixing process in the convective region. From (4) and (5), the good agreement between the horizontal distributions of $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ (also shown by the small difference between them in Fig. 5c) suggests that the heating and drying are primarily due to condensation processes in the MCC, because the sensible heat flux and evaporation from the surface are small. The diagnosed rainfall rates from $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ are the same order as the observed rainfall rate ($\sim 10$ mm h$^{-1}$), which is calculated over three-hour periods centered at 0000 UTC using the PAM and SAM data.

To identify the mesoscale convective system and examine the nature of precipitation in more detail, the infrared satellite image at 0000 UTC 4 June and the radar analysis at 0010 UTC 4 June are shown in Fig. 6. The satellite picture shows the MCC over Kansas and northern Oklahoma (Fig. 6a). The radar composite (Fig. 6b) shows that the precipitation consists of both convective and stratiform components. A narrow but intense convective line extends from southern Kansas southward to northern Oklahoma. An extensive stratiform precipitation region extends to the west and north of a broader but less intense convective band oriented from southwest to northeast over Kansas. The whole pattern of reflectivity corresponds well to that

![Fig. 5](image-url). Horizontal distributions of the vertically integrated (a) heat source $\langle Q_1 \rangle$, (b) moisture sink $\langle Q_2 \rangle$, (c) $\langle Q_2 \rangle - \langle Q_2 \rangle$ (units: 10$^3$ W m$^{-2}$). (d) The observed streamlines and isolach (units: m s$^{-1}$), and (e) the streamlines of divergent wind and divergence (units: 10$^{-8}$ s$^{-1}$) at 200 mb for the MCC at 0000 UTC 4 June 1985 during PRE-STORM.
of $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$. This further confirms that the observed heating and drying are contributed from both convective and mesoscale stratiform precipitation.

Figure 7 shows the horizontal distributions of $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$, and the associated wind and divergence fields at 200 mb for a squall line at 0300 UTC 11 June 1985. Many observational and numerical studies have been done to examine the structure and dynamics of
the squall line (e.g., Johnson and Hamilton 1988; Rutledge et al. 1988; Zhang and Gao 1989; Zhang et al. 1989). The linear patterns of $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ (Figs. 7a and 7b) are clearly associated with the axis of upper-level divergence oriented from southwest to northeast and the diffuseness of streamlines of the divergent wind (Fig. 7e). The isolates in Fig. 7d show strong horizontal shear and do not show a wind speed minimum as in the MCC case. The diagnosed rainfall rate estimated from $\langle Q_1 \rangle$ (averaged about 10 mm h$^{-1}$) corresponds very well with the observed rainfall rate calculated over three-hour periods centered at 0300 UTC using the PAM and SAM data, but the rainfall rate estimated from $\langle Q_2 \rangle$ is about 20% lower than the observed one. It is noted that the difference between $\langle Q_1 \rangle$ and $\langle Q_2 \rangle$ is relatively larger (Fig. 7c) compared that for the MCC (Fig. 5c). Since the sensible heat flux and evaporation from the surface are small, it is speculated that the horizontal eddy flux divergence may be responsible for this difference [see Eq. (6)].

The possible effect of the horizontal eddy flux divergence term seems related to the organization of the convective system. Figures 8a and 8b show the infrared satellite image and the radar analysis at 0300 UTC 11 June. The satellite picture (Fig. 8a) shows that the upper-level cloud shield covers much of Kansas and northwestern Oklahoma with a well-organized convective region (temperature $<-70^\circ$). The radar reflectivity pattern (Fig. 8b) clearly shows the convective line oriented from southwest to northeast. The stratiform precipitation is present to the rear of the convective line.

The horizontal distributions of the vertically integrated heat source and moisture sink for the MCC and squall line are well correlated with the radar reflectivity pattern. Similar features also appear in other cases (listed in Table 1) observed during PRE-STORM (not shown).

6. Contributions of cumulus ensemble to large-scale budget residuals

In this section, we will apply the diagnostic and semiprognostic methods to the PRE-STORM data to estimate the contributions of the cumulus ensemble to the large-scale heat and moisture budget residuals in midlatitude convective systems. Detailed analyses will be made mainly for the MCC at 0000 UTC 4 June and the squall line at 0300 UTC 11 June. Both are well documented by many case studies (e.g., Johnson and Hamilton 1988; Rutledge et al. 1988; Smull and Au-
Fig. 8. (a) The infrared satellite image and (b) radar composite at 0300 UTC 11 June.

a. Diagnosed cumulus ensemble effects

Figure 9a shows the vertical profiles of the observed $Q_1$ and $Q_2$ (solid lines) and the $Q_{1C}$ and $Q_{2C}$ diagnosed from $Q_1 - Q_2$ using the cumulus ensemble model with the updraft only (dashed lines) and with both the updraft and downdraft (dotted lines) averaged over the domain 37°–39°N, 96°–98°W for the MCC at 0000 UTC 4 June. The observed heating shows a maximum at 450 mb, less heating in the lower troposphere, and a slight cooling near the cloud base. On the other hand, the profile of apparent moisture sink shows different features. The observed drying shows its principal maximum at 550 mb, and secondary peaks at 400 mb and 850 mb. The separation of the peak levels between the profiles of $Q_1$ and $Q_2$ suggests the presence of vertical eddy transport of heat and moisture by cumulus convection [see Eq. (3)]. This is supported by the profiles of diagnostically obtained $Q_{1C}$ and $Q_{2C}$. As shown in Fig. 9a, the cumulus heating and drying explain most of the observed heating and drying in the upper layers. The maximum $Q_{1C}$ is at the same levels as the maximum $Q_1$. Examining further each term on the right-hand side of (7) and (8), we find that the heating and drying due to the environmental subsidence compensating the cloud mass flux dominates the cumulus contributions (not shown). Comparing the profiles of $Q_1$, $Q_2$ and $Q_{1C}$, $Q_{2C}$, it is also clearly seen that insufficient heating and drying appear in the upper troposphere, and excessive heating and drying in the lower troposphere. These suggest the presence of additional mesoscale stratiform heating and drying ($Q_{1m} \approx Q_{2m}$) in the upper troposphere and cooling and moistening due to the convective-scale and mesoscale downdrafts in the lower troposphere. More extensive analyses of convective-scale downdraft effects and mesoscale stratiform cloud effects are given in sections 6b and 7, respectively.

Figure 9b shows the vertical profiles of the diagnosed convective mass flux $M_c$ and the observed large-scale mass flux $\bar{M}$ for the MCC. The significant feature seen in the figure is that $M_c$ exceeds $\bar{M}$ below the 600-mb level, but $M_c$ is less than $\bar{M}$ above that level. This indicates that, in the lower troposphere, the upward mass flux in the cumulus ensemble is larger than the mass flux required from the large-scale horizontal convergence. Since $\bar{M} - M_c = M + M_m$, where $M$ is the mass flux of environment and $M_m$ the mass flux of mesoscale stratiform clouds, the positive value of $\bar{M} - M_c$ in the upper troposphere suggests the upward mass flux is associated with mesoscale stratiform clouds.

The vertical profiles of $Q_1$ and $Q_2$ (solid lines) and $Q_{1C}$ and $Q_{2C}$ (dashed lines) in the squall line at 0300 UTC 11 June, which are averaged over the domain 36°–38°N, 98°–100°W, are shown in Fig. 10a. Note that there is also considerable difference between the profiles of $Q_1$ and $Q_2$. The maximum heating (reaching ~250 K day$^{-1}$) (agreeing with Gallus and Johnson 1991) is at 400 mb, while the principal drying is at 600 mb. As in the MCC case, the general agreement between the profiles of $Q_1$ and $Q_{1C}$ and $Q_2$ and $Q_{2C}$ shows that the contributions of cumulus convection are dominant in the observed budget residuals. However, insufficient heating and drying in the upper troposphere and excessive heating and drying in the lower troposphere again suggest possible contributions from the mesoscale stratiform precipitation and convective-scale downdrafts to the budget residuals. Figure 10b shows
that $M_e$ is larger than $\bar{M}$ below the 500-mb level, but the reverse is true above that level.

Diagnostic analyses of other cases (listed in Table 1) observed during PRE-STORM were also done (figures not shown). We find that the $Q_{1e}$ and $Q_{2e}$ given by (7) and (8), with the diagnostically obtained cloud mass flux $M_e$ and detrainment $\delta$ using the spectral cumulus ensemble model, are capable of approximately reproducing a variety of vertical distributions of cumulus heating and drying in midlatitude mesoscale convective systems in spite of the existence of mesoscale stratiform clouds. The cloud mass flux in midlatitude convective systems is mainly contributed to from deep clouds, supporting previous studies (e.g., Lewis 1975; Grell et al. 1991).

b. Effects of convective-scale downdrafts

The convective-scale downdrafts induced by the accumulation of rain within updrafts were first found during the Ohio–Florida Thunderstorm Project (Byers and Braham 1949). Observations during GATE also showed that the convective-scale downdraft is an important component of tropical cloud clusters (e.g., Zipser 1977; Houze 1977; Houze and Betts 1981). Recent studies using the data obtained during PRE-STORM also show the existence of convective-scale downdrafts in the mesoscale convective systems (e.g., Rutledge et al. 1988; Stumpf et al. 1991). The consecutive development of new convective cells triggered by the downdraft outflow may be an important process leading to the organization of cumulus convection (e.g., Rotunno et al. 1988). Considering the potential significance of the convective-scale downdrafts, several attempts have been made to include the impacts of downdrafts in the diagnostic studies of cumulus effects (e.g., Johnson 1976; Nitta 1977; Cheng 1989b). In this section, the diagnostic cumulus ensemble model with both the convective-scale updraft and downdraft (Cheng 1989a,b) is applied to the PRE-STORM dataset to examine the effects of convective-scale downdrafts on the large-scale heat and moisture budget of midlatitude convective systems.

The dotted lines in Figs. 9 and 10 show the vertical profiles of $Q_{1e}$, $Q_{2e}$, and $M_e$, diagnosed from the observed $Q_1 - Q_2$ using the model with convective-scale downdrafts for the convective region of the MCC and squall line, respectively. The convective-scale downdraft in the 4 June MCC has been identified from the Doppler radar analysis by Smull and Augustine (1989). Using Doppler radar observations, Rutledge et al. (1988) and Biggerstaff and Houze (1991) showed that an intense convective-scale downdraft also existed in the 11 June squall line. For the MCC, the diagnosed downdraft mass flux is about half the updraft mass flux (not shown) and results in less total mass flux in the lower troposphere (dotted lines in Fig. 9b). The cumulus heating and drying in the lower layer are reduced by 20% (comparing dotted lines with dashed lines in Fig. 9a). For the squall line, the inclusion of downdrafts results in a total convective mass flux about half the updraft mass flux, dramatically decreasing the discrepancy between the observed and diagnosed drying in the lower troposphere, and reducing the cumulus heating by about 50% (comparing dotted lines with dashed lines in Fig. 10).

We further examined the separate contributions of convective-scale updrafts and downdrafts to the profiles of the diagnosed $Q_{1e}$ and $Q_{2e}$ (not shown). The convective-scale downdrafts tend to cool and moisten the lower troposphere for both the MCC and squall line. Near cloud base, the cooling and moistening due to the downdrafts is about half of the heating and drying due to the updrafts. The ascending environmental air, which compensates the downdraft mass flux, contributes to the major cooling and moistening, while the effect of the detrainment from the downdrafts is negligibly small in the lower cloud layer.
Similar results are obtained by applying the model to other PRE-STORM cases (listed in Table 1). The inclusion of convective-scale downdrafts indeed reduces the magnitudes of cumulus heating and drying in the lower troposphere. However, differences between the observed $Q_1$ and diagnosed $Q_{1c}$, and between the observed $Q_2$ and diagnosed $Q_{2c}$, remain in the upper and lower troposphere.

c. Predicted cumulus ensemble effects

In the diagnostic analysis, the cumulus heating and drying are estimated by using the diagnostically obtained cloud-base mass flux $m_B$ which is derived from the observed $Q_1 - Q_2$ [see Eq. (15)]. Besides the vertical eddy transport of heat and moisture, the observed $Q_1 - Q_2$ may include other unresolved processes, such as the horizontal eddy transport of heat and moisture, which may affect the determination of the cumulus effects (Wu 1992). Therefore, we further apply the semiprognostic analysis, which is independent of the observed $Q_1$ and $Q_2$, to predict the cumulus heating and drying from the observed environmental conditions.

Figures 11 and 12 show the observed $\tilde{M}$, $Q_1$, and $Q_2$ (solid lines) and the predicted $M_{cz}$, $Q_{1c}$, and $Q_{2c}$, by the Arakawa–Schubert (A–S) parameterization scheme with (dashed lines) and without (dotted lines) the quasi-equilibrium assumption for the MCC and squall line, respectively. Comparing these figures with Figs. 9 and 10, it is interesting to note that the predicted and diagnosed cumulus heating and drying possess very similar features. The predicted cumulus heating and drying produce the major parts of the observed heating and drying. This indicates that the A–S cumulus parameterization scheme performs quite well in the environment of midlatitude convective systems. It is also noted that discrepancies exist between the observed $Q_1$.

Fig. 11. Vertical profiles of (a) observed $Q_1$ and $-Q_2$ (solid) and (b) $\tilde{M}$ (solid), and (a) predicted $Q_{1c}$ and $-Q_{2c}$ and (b) $M_{cz}$ using the A–S scheme with the quasi-equilibrium assumption (dashed) and with the observed $dA/dt$ (dotted) for the MCC at 0000 UTC 4 June.

Fig. 12. As in Fig. 11 but for the squall line at 0300 UTC 11 June.
and \( Q_2 \) and the predicted \( Q_{1c} \) and \( Q_{2c} \) in the upper and lower troposphere.

We further examined the impacts of the observed time change of the cloud work function \( dA/dt \) on the predicted cumulus effects. Following Lord (1982), the observed \( dA/dt \) is calculated by

\[
\frac{dA}{dt} = \frac{A(t + \Delta t) - A(t - \Delta t)}{2\Delta t},
\]

where \( \Delta t = 3h \) is the time interval of observation.

Figures 13a and 13b show the large-scale forcing \( F \) [see Eqs. (19) and (20)] and the observed time change of cloud work function \( dA/dt \) for the MCC and the squall line, respectively. The magnitudes of \( F \) are much larger than the values of \( dA/dt \) for the squall line case (Fig. 13b). Even the relatively weaker large-scale forcing for the MCC case is larger than the observed time change of cloud work function (Fig. 13a). For a given large-scale forcing, \( dA/dt = 0 \) means that the stabilization by cumulus convection exactly balances the destabilization by the large-scale processes, while \( dA/dt < 0 \) means that the stabilization by convection exceeds the destabilization by the large-scale processes and vice versa. Thus, from (17) the inclusion of the observed \( dA/dt \) tends to enhance cumulus convection in the MCC, while it inhibits convection in the squall line case. The results are to slightly increase the predicted cumulus heating and drying in the MCC case (dotted lines in Fig. 11), while slightly decreasing the predicted cumulus heating and drying in the squall line case (dotted lines in Fig. 12). The small differences between the profiles of the predicted cumulus heating and drying with and without the effect of nonequilibrium of cloud work function shown in Figs. 11 and 12 further support that the quasi-equilibrium assumption of cloud work function is approximately valid in the environment of midlatitude convective systems. Therefore, underestimates of the observed \( Q_1 \) and \( Q_2 \) in the upper troposphere and overestimates in the lower troposphere must come from physical processes not included in the A–S scheme, such as convective-scale downdrafts and mesoscale stratiform clouds.

We also applied the semiprognostic analysis to other cases observed during PRE-STORM. Figures 14a and 14b show the mean vertical profiles of the observed \( M_f, Q_1 \), and \( Q_2 \) and those of the predicted \( M_f, Q_{1c}, \) and \( Q_{2c} \) obtained from six observation times (Table 1). The profiles of predicted \( Q_{1c} \) and \( Q_{2c} \) are similar to those of observed \( Q_1 \) and \( Q_2 \) (Fig. 14a). But insufficient heating and drying appear in the upper troposphere and excessive heating and drying in the lower troposphere, suggesting that mesoscale stratiform precipitation and convective-scale downdrafts must be also considered. It is also seen that \( M_f \) exceeds \( M \) below the 500-mb level, but \( M_f \) is less than \( M \) above that level (Fig. 14b). Examination of the cloud-base mass flux \( m_b(p_b) \) according to cloud types (Fig. 14c) shows that the total convective mass flux \( M_f \) is mainly contributed from deep clouds. This is consistent with the results shown in section 6a.

7. Effects of mesoscale stratiform clouds

The results shown in Figs. 9 and 10 demonstrate that the cumulus ensemble model including both the convective-scale updrafts and downdrafts can closely approximate the observed \( Q_1 \) and \( Q_2 \) using the formulation given by (7) and (8). Differences still exist, however, between the observed \( Q_1 \) and \( Q_2 \) and diagnosed \( Q_{1c} \) and \( Q_{2c} \). An interesting feature in Figs. 9 and 10 is that \( Q_1 - Q_{1c} \) is approximately equal to \( Q_2 - Q_{2c} \) in the upper troposphere. This suggests the contributions of mesoscale stratiform clouds to \( Q_1 \) and \( Q_2 \). We estimate \( Q_{1m} \approx Q_{2m} \) by \( [(Q_1 - Q_{1c}) + (Q_2 - Q_{2c})]/2 \). Figure 15 shows the vertical profile of \( Q_{1m} \) \( [\approx Q_{2m}, \text{see Eqs. (11)–(13)}] \) for the MCC (solid) and the squall line (dashed). The estimated \( Q_{1m} \) shows
positive values in the upper troposphere and negative values in the lower troposphere. The profile of \( Q_{1m} \) for the squall line is similar to that obtained using a different method by Gallus and Johnson (1991).

The mass fluxes of the cumulus environment \( \bar{M} + M_m (= \bar{M} - M_s) \) are shown in Fig. 16 for the MCC and the squall line. For the MCC, the mass fluxes of the cumulus environment are upward above 600 mb with the maximum of \( \sim 40 \text{ mb h}^{-1} \) \((\sim 20 \text{ cm s}^{-1})\) near 300 mb and downward below 600 mb with the maximum of \( \sim 15 \text{ mb h}^{-1} \) \((\sim 5 \text{ cm s}^{-1})\) near 900 mb for the MCC. For the squall line, the mass fluxes of the cumulus environment are upward above 500 mb with the maximum of \( \sim 50 \text{ mb h}^{-1} \) \((\sim 25 \text{ cm s}^{-1})\) near 300 mb and downward below 500 mb with the maximum of \( \sim 30 \text{ mb h}^{-1} \) \((\sim 10 \text{ cm s}^{-1})\) near 900 mb. Positive values of \( \bar{M} + M_m \) imply the presence of mass flux associated with mesoscale updrafts, while negative values of \( \bar{M} + M_m \) suggest the presence of mass flux associated with mesoscale downdrafts (Cheng and Yanai 1989). Therefore, the mesoscale heating and drying in the upper troposphere due to condensation are associated with the mesoscale updraft. On the other hand, the mesoscale cooling and moistening in the lower troposphere due to evaporation are related to the mesoscale downdraft.
These results are qualitatively consistent with the findings of the observational studies using Doppler radar data. For the 4 June MCC, the Doppler radar analysis by Stumpf et al. (1991) displays the mesoscale updraft above 5.5 km (with maximum speed 60 cm s⁻¹ near 8.5 km) and the mesoscale downdraft below 5.5 km (with maximum speed 65 cm s⁻¹ near 3 km). For the 11 June squall line, the Doppler radar analysis by Rutledge et al. (1988) shows the mesoscale updraft above 6 km (maximum speed 30 cm s⁻¹ near 10.5 km in a vertical motion profile averaged over ~1.5 h periods) and the mesoscale downdraft below 6 km (maximum speed 50 cm s⁻¹ near 3.5 km). Using data composited over 5-h periods, Biggerstaff and Houze (1991) showed that the peak of maximum upward motion (30 cm s⁻¹ near 300 mb) in the mesoscale stratiform region is higher than the peak of maximum upward motion in the convective region (80 cm s⁻¹ near 400 mb).

To show the mesoscale stratiform cloud effects more clearly, we further examine the quantity $H$, which was defined by Cheng and Yanai (1989) as follows:

$$ H = Q_2 - Q_{1c} \left( -L \frac{\partial \bar{q}}{\partial p} \right) / \left( \frac{\partial \bar{e}}{\partial p} \right). \quad (24) $$

Using (11), (12), and (13) we can write (24) as

$$ H = Q_{2m} \left( \frac{\partial h}{\partial p} \right) / \left( \frac{\partial \bar{e}}{\partial p} \right) + Q_{2c} - Q_{1c} \left( -L \frac{\partial \bar{q}}{\partial p} \right) / \left( \frac{\partial \bar{e}}{\partial p} \right). \quad (25) $$

From (25), (7), and (8) we can show that the subsidence terms of $Q_{1c}$ and $Q_{2c}$ have no contribution to $H$ and the detrainment terms of $Q_{1c}$ and $Q_{2c}$ contribute mainly negative values to $H$. Therefore, $H$ represents the effects of mesoscale processes and detrainment from...
cumulus clouds. From (25), it is also seen that in the upper troposphere \((\partial h/\partial p)/(\partial s/\partial p) > 0\), the condensation \((Q_{2m} > 0)\) associated with the mesoscale updraft contributes positive values to \(H\). In the lower troposphere \((\partial h/\partial p)/(\partial s/\partial p) < 0\), the evaporation \((Q_{2m} < 0)\) associated with the mesoscale downdraft contributes positive values to \(H\). Figures 17a and 17b show the horizontal distributions of \(H\) at 300 mb for the MCC and the squall line, respectively. Positive values of \(H\) are found in the west and northeast parts of the MCC and in the northwest part of the squall line. Positive values of \(H\) are also found at 650 mb in the area of organized convection (Figs. 18a and 18b). These results clearly show the existence of condensation and evaporation effects associated with mesoscale stratiform clouds in both systems.

8. Comparison between the tropics and midlatitudes

a. Quasi-equilibrium assumption

Figures 19a and 19b show the scatter diagrams of the observed time change of cloud work function versus the large-scale forcing for clouds whose tops reach near 200 mb in mesoscale convective systems observed during GATE and PRE-STORM, respectively. The scatter points in both figures are distributed around the zero line, showing that the quasi-equilibrium assumption is approximately valid in both the GATE and PRE-STORM cases. Comparing Fig. 19a with 19b, we also find that the large-scale forcing in the PRE-STORM cases is much stronger than that in the tropics. The large-scale forcing has a maxima of 9 kJ kg\(^{-1}\) day\(^{-1}\) in the GATE and 60 kJ kg\(^{-1}\) day\(^{-1}\) in the PRE-STORM. On the other hand, the values of \(dA/dt\) are confined between \(-5\) kJ kg\(^{-1}\) day\(^{-1}\) and 5 kJ kg\(^{-1}\) day\(^{-1}\) in GATE and between \(-10\) kJ kg\(^{-1}\) day\(^{-1}\) and 10 kJ kg\(^{-1}\) day\(^{-1}\) in PRE-STORM. Therefore, the quasi-equilibrium assumption holds better in the PRE-STORM cases than in the GATE cases.

b. Cumulus ensemble effects

Figures 20 and 21 show the mean vertical distributions of the observed \(M\), \(Q_1\), \(Q_2\), \(M_c\), \(Q_{1c}\), and \(Q_{2c}\) diagnosed from \(Q_1 - Q_2\) using the model with convective-scale downdrafts for GATE and PRE-STORM (see Table 1), respectively. It can be seen that the vertical distribution of cumulus heating in the tropics is somewhat uniform (Fig. 20a), while the distribution in the midlatitudes has a sharp peak at 400 mb (Fig. 21a). The distributions of cumulus drying also show a difference between the two regions, that is, the peak of drying is located at a lower level in the tropics (Fig. 20a) than in the midlatitudes (Fig. 21a). The relative contributions of the downdraft mass flux to the total convective mass flux are about the same in both GATE (Fig. 20b) and PRE-STORM (Fig. 21b) cases, although the intensity of downdrafts in the PRE-STORM cases is much stronger than that in the GATE cases. Near the cloud base, the downdraft mass flux is approximately one-half the updraft mass flux in both the GATE and PRE-STORM cases. Further examining the contributions of the updraft and downdraft to the diagnosed \(Q_{1c}\) and \(Q_{2c}\) in the GATE and PRE-STORM cases (not shown), it is found that the downdrafts cool and moisten the lower troposphere. The environmental ascent, which compensates the downdraft mass flux, contributes to the major cooling and moistening, while the effect of detrainment from the downdraft is negligibly small in the lower cloud layer in both GATE and PRE-STORM.

Figures 22a and 22b show the mean cloud-base mass fluxes for the GATE and PRE-STORM cases, respectively. Both deep and shallow clouds exist in the tropical convective systems (Fig. 22a), while only deep clouds

![Fig. 19. Scatter diagrams of the observed time change of cloud work function versus the large-scale forcing for deep clouds in mesoscale convective systems observed during (a) GATE and (b) PRE-STORM.](attachment:fig19.png)
appear in the midlatitude convective systems (Fig. 22b). These results are consistent with the results obtained by previous studies (e.g., Yanai et al. 1973; Lewis 1975; Nitta 1978; Grell et al. 1991).

c. Mesoscale stratiform cloud effects

Figure 23 shows the mean vertical profiles of $Q_{1m}$ ($\approx Q_{2m}$) for both GATE and PRE-STORM. It is shown in Fig. 23a that the heating in the upper troposphere peaks at 300 mb and the cooling in the middle troposphere peaks at 500 mb for the GATE mean. Normally the cooling peak is in the lower troposphere (e.g., Houze 1989). Since $Q_{1m}$ is estimated from differences between the observed residuals and the diagnosed cumulus effects, our method may underestimate the cooling below 600 mb. For PRE-STORM (Fig. 23b) heating appears in a deep layer above 600 mb with a peak at 350 mb and cooling occurs below with a peak at 700 mb. Note that the mean heating rate in PRE-STORM is about three times that in GATE.

Figure 24 shows the vertical profiles of the mass flux $\dot{M} + M_m$ ($= \dot{M} - M_e$) for both GATE and PRE-STORM. We find that positive values of $\dot{M} + M_m$ appear above 400 mb and negative values of $\dot{M} + M_m$ below for GATE (Fig. 24a). In PRE-STORM, $\dot{M} + M_m$ shows positive values above 600 mb and negative values below (Fig. 24b). We also see that the peak value of $\dot{M} + M_m$ in PRE-STORM is about three times that in GATE. Clearly, the heating and cooling shown in Fig. 23 are related to the mesoscale updraft and downdraft in stratiform clouds. These results are qualitatively consistent with those shown in a review paper by Houze (1989).
9. Summary and conclusions

In this study, we examined the cumulus–environment interaction in midlatitude convective systems by performing diagnostic and semiprog nostic analyses of the OK PRE-STORM data. The dataset was obtained from a network with an average station separation of ~150 km, which is comparable in horizontal resolution to the GATE dataset. The similarities and differences of cumulus and mesoscale stratiform cloud effects in the midlatitudes and tropics are also discussed.

Large-scale environment conditions obtained from PRE-STORM and GATE data show generally larger vertical wind shear, large-scale forcing, and moist convective instability in the midlatitude MCCs and squall lines than in the tropical cloud clusters. The stronger mean shear suggests that the dynamic interaction of cumulus convection with the large-scale environment is more significant in the midlatitudes than tropics, and results in well-organized midlatitude convection as seen in satellite and radar pictures. The larger moist convective instability suggests a more unstable environment for the midlatitude than the tropical convective systems. Stronger stabilization by cumulus convection is required to balance the destabilization by the large-scale vertical advection in the midlatitudes than in the tropics.

The large-scale heat and moisture budget residuals in the PRE-STORM cases are much larger than those in the GATE cases. The horizontal distributions of the vertically integrated heat source and moisture sink for the MCCs and squall lines correspond very well with the radar reflectivity patterns, suggesting the contributions of convective and mesoscale stratiform precipitation to the observed heating and drying.

The results obtained from the diagnostic and semiprog nostic analyses using the PRE-STORM data show that the contributions of cumulus convection are dominant in the observed heat and moisture budget residuals. The physical processes through which the cumulus ensemble interacts with the large-scale environment are basically the same in the tropics and mid-latitudes, that is, a cumulus ensemble modifies the thermodynamical fields of its environment primarily through the subsidence of the environmental air that compensates the convective mass flux and through the detrainment of heat and water substance from clouds.

It is shown that only deep clouds appear in the mid-latitude convective systems, while both deep and shallow cumuli exist in the tropical systems. The effects of convective-scale downdrafts are to cool and moisten the lower troposphere in the midlatitudes as in the tropics (Cheng 1989b). The Arakawa–Schubert quasi-equilibrium assumption is valid and holds better in the
midlatitudes than in the tropics since the large-scale forcing is much stronger.

The results also show the contributions of mesoscale stratiform clouds to the vertical distributions of the observed heating and drying in the upper and lower troposphere for midlatitude convective systems, even though they are small compared with the cumulus effects. The isolated stratiform cloud effects show the heating and drying in the upper troposphere due to the condensation associated with the mesoscale updraft, and the cooling and moistening in the lower troposphere due to the evaporation associated with the mesoscale downdraft. Both cumulus and stratiform cloud effects are stronger in midlatitude convective systems than in tropical systems.

The presence of mesoscale circulations raises a question concerning the separation of space and time scales between cumulus convection and large-scale processes, which is tacitly assumed in the theory of cumulus–environment interaction (e.g., Frank 1983; Arakawa and Chen 1987; Cheng and Yanai 1989). This problem was discussed in Wu (1992). Another problem is the effect of vertical wind shear on the cumulus–environment interaction which was examined in Wu and Yanai (1991a) and Wu (1992).

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REFERENCES


——, 1989b: Effects of downdrafts and mesoscale convective organization on the heat and moisture budgets of tropical cloud


