The Use of Cloud-Resolving Simulations of Mesoscale Convective Systems to Build a Mesoscale Parameterization Scheme

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ABSTRACT

A method is described for parameterizing thermodynamic forcing by the mesoscale updrafts and downdrafts of mesoscale convective systems (MCSs) in models with resolution too coarse to resolve these drafts. The parameterization contains improvements over previous schemes, including a more sophisticated convective driver and inclusion of the vertical distribution of various physical processes obtained through conditional sampling of two cloud-resolving MCS simulations. The mesoscale parameterization is tied to a version of the Arakawa–Schubert convective parameterization scheme that is modified to employ a prognostic closure. The parameterized Arakawa–Schubert cumulus convection provides condensed water, ice, and water vapor, which drives the parameterization for the large-scale effects of mesoscale circulations associated with the convection. In the mesoscale parameterization, determining thermodynamic forcing of the large scale depends on knowing the vertically integrated values and the vertical distributions of phase transformation rates and mesoscale eddy fluxes of entropy and water vapor in mesoscale updrafts and downdrafts. The relative magnitudes of these quantities are constrained by assumptions made about the relationships between various quantities in an MCS's water budget deduced from the cloud-resolving MCS simulations. The MCS simulations include one of a tropical MCS observed during the 1987 Australian monsoon season (EMEX9) and one of a midlatitude MCS observed during a 1985 field experiment in the Central Plains of the United States (PRE-STORM 23–24 June).

1. Introduction

Zipser (1982) proposed the term mesoscale convective system (MCS) to refer to all organized precipitation systems on scales from 20 to 500 km that include deep convection during some part of their lifetime. The organization of convective cells onto larger scales frequently enables mesoscale circulations to develop. These mesoscale circulations are distinct from both the convective circulations within the cells themselves and the larger-scale synoptic circulations. MCSs contain several distinct circulations—convective updrafts, convective downdrafts, mesoscale updrafts, and mesoscale downdrafts—whose effects should be accounted for in those numerical models that do not resolve these circulations. Heretofore, mesoscale effects have not been included in GCM convective parameterization schemes, leading to errors in diagnosed heating and moistening. Wu (1993), for instance, assessed the errors that arise in parameterization schemes that fail to consider mesoscale effects. For instance, his Figs. 11 and 12 show vertical profiles of observed Q_1 and Q_2 and prognosed

Q_{1c} and Q_{2c} for MCSs observed over the south-central United States on 4 June 1985 and 10 June 1985. The quantities Q_1 and Q_2 are the residuals of the heat and moisture budgets of the resolvable motion, respectively, and as such represent the “apparent” heat source and moisture sink, respectively (Yanai et al. 1973). The quantities Q_{1c} and Q_{2c} represent the prognosed contributions of cumulus clouds (condensation, evaporation, and convective transports) to the environment. Wu's profiles of Q_1 , Q_2 , and Q_{1c} , Q_{2c} show that the Arakawa–Schubert convective scheme prognoses insufficient heating and drying in the upper troposphere and excessive heating and drying in the lower troposphere. These differences in the observed and prognosed profiles suggest the presence of additional heating and drying in the upper troposphere due to mesoscale updrafts and additional cooling and moistening in the lower troposphere due to mesoscale downdrafts. Estimated values of mesoscale effects Q_{1m} and Q_{2m} show positive values in the upper troposphere and negative values in the lower troposphere whose maximum magnitudes are roughly 20%–30% as large as the maximum magnitudes of Q_{1c} and Q_{2c} .

Increasingly, cloud-resolving numerical models have been used as a tool to assist in the formulation and testing of cloud parameterization schemes for larger-scale models. For example, Browning et al. (1993) discuss this approach in the context of the Global Energy

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and Water Cycle Experiment Cloud Systems study. Here, we discuss two MCS simulations—one tropical MCS and one midlatitude MCS—that are subsequently used to construct an MCS parameterization scheme. The simulations are three-dimensional and employ horizontal grid spacing fine enough that no convective parameterization scheme needs to be (or should be) used. Both simulations use multiple two-way interactive nested grids in order to achieve the minimum possible horizontal grid spacing yet include larger-scale horizontally inhomogeneous forcing. Both simulations explicitly simulate convection on their finest grid at a horizontal grid spacing of 1500–2000 m over an area of $\sim 17\,000\text{ km}^2$ for a time of 3–4 h.

The tropical MCS that we have simulated is the ninth MCS probed by research aircraft during the Equatorial Mesoscale Experiment (EMEX); this MCS is called EMEX9 and occurred on 2–3 February 1987. Mapes and Houze (1992) provide a detailed view of the horizontal structure of the 10 EMEX precipitation systems. Gamache et al. (1987), Bograd (1989), and Webster and Houze (1991) describe synoptic conditions for EMEX9. Wong et al. (1993) and Tao et al. (1993) have modeled EMEX9 two-dimensionally.

The midlatitude MCS that has been simulated is the MCS observed on 23–24 June 1985 during the Oklahoma–Kansas Preliminary Regional Experiment for STORM (Stormscale Operational and Research Meteorology Program), or PRE-STORM (Cunning 1986). Stensrud and Maddox (1988), Johnson et al. (1989), Johnson and Bartels (1992), and Bernstein and Johnson (1994) are among the researchers who have investigated various aspects of the 23–24 June MCS event. This MCS has been simulated by Olsson and Cotton (1997), who used a minimum grid spacing of $\sim 8\text{ km}$.

Here, we describe the application of the two aforementioned simulations to the construction of a parameterization for MCSs. In section 2 we discuss the details of the model that we have used to simulate these two MCSs, as well as the model setup for each simulation. In section 3 we discuss the results of the two simulations. In section 4 we describe the framework of the parameterization scheme. In section 5 we discuss the construction of the scheme. In section 6 we evaluate the performance of the scheme. We summarize our results in section 7.

2. The model

The nonhydrostatic version of the Regional Atmospheric Modeling System (RAMS; Pielke et al. 1992) is used. The model uses an isentropic analysis package to derive initial conditions and time-dependent lateral boundary conditions; this package interpolates pressure-level data onto 33 isentropic levels and applies the Barnes (1973) objective analysis scheme. Time-dependent model variables are the three velocity components, the perturbation Exner function, the ice–liquid potential

temperature, the total water mixing ratio, and the mixing ratio of rain droplets, snowflakes, pristine ice crystals, graupel particles, and aggregates. The bulk hydrometeors have prescribed exponential size distributions (see Cotton et al. 1986 or Flatau et al. 1989 for a full description of the microphysics parameterization). The model diagnoses vapor mixing ratio, cloud water mixing ratio, and potential temperature. The prognostic equations use a time-splitting technique; this allows the model to explicitly compute on a small time step those terms governing sound waves and to compute on a long time step those terms governing other processes. Horizontal time differencing for long time steps is a flux conservative form of second-order leapfrog (Tripoli and Cotton 1982).

The simulations employ the following boundary conditions: The lateral boundary conditions are the Klemp and Wilhelmson (1978a,b) radiative type, in which the normal velocity component specified at a lateral boundary is effectively advected from the interior, assuming a specified propagation speed. A Davies nudging condition causes model data at and near the lateral boundaries to be forced toward available observations. A rigid lid is used at the model top, in concert with a Rayleigh friction absorbing layer. The latter damps gravity wave and other disturbances that approach the top boundary. At the lower boundary there is horizontally variable topography. A vegetation parameterization (McCumber and Pielke 1981; Lee 1992) and an 11-layer prognostic soil model (Tremback and Kessler 1985) dictate fluxes of temperature and moisture over land surfaces. The soil model has 11 vertical levels located from -1 cm to -1 m . The parameterization of surface fluxes of momentum, heat, and moisture are designed according to surface similarity theory (Louis 1979). The Chen and Cotton (1983) radiation scheme accounts for longwave and shortwave radiative transfer, including the effects of liquid water and ice; the model updates radiative contributions to atmospheric and surface soil temperatures every 15 min. The Level 2.5w convective parameterization scheme (Weissbluth and Cotton 1993) is used, but only in the initial stages of each simulation. In each simulation, the cumulus parameterization was run for a short time on the finest grid in order to excite smaller scales of motion and assist in organizing convection on the meso- γ scale. Our use of the cumulus parameterization in this fashion can be thought of as an objective “hot bubble.”

3. Simulations

a. EMEX9

1) OBSERVATIONS

EMEX9 occurred during the most active phase of the 1987 Australian monsoon: 850-mb westerly winds over the EMEX9 region were on the order of 25 m s^{-1} . The prevailing synoptic feature at the time of EMEX9 was

a deep westerly monsoon trough extending from 500 mb to the surface, which oriented itself across northern Australia and into Papua New Guinea. The interaction of this synoptic-scale monsoon trough circulation with a mesoscale land breeze circulation provided a primary lifting mechanism for EMEX9. A composite sounding of the EMEX9 environment, assembled using all available aircraft and synoptic data, has a CAPE of 1484 J kg^{-1} and a bulk Richardson number of 51, typical of multicellular convection (Alexander and Young 1992).

The NOAA P-3, NCAR Electra, and CSIRO F-27 aircraft penetrated EMEX9 between 2100 UTC 2 February and 0100 UTC 3 February. The P-3's lower-fuselage Doppler radar observed two separate convective lines: an initial line oriented in a west-northwest to east-southeast direction that was more than 300 km long and a northwest to southeast-oriented convective line that was about 250 km long. The P-3 radar observed two types of embedded convection associated with EMEX9: upright and rearward sloping (Webster and Houze 1991). Both types extended to about 14.5 km and had a horizontal scale of about 40 km. Vertical motions in the upright convection were weak (on the order of $1\text{--}2 \text{ m s}^{-1}$) and were confined to upper levels. Vertical motions in the slanted convection were much stronger with maximum updraft strengths on the order of $7\text{--}9 \text{ m s}^{-1}$ at about 10 km. Radar observations of EMEX9's stratiform region indicated a more uniform precipitation field and a bright band near the freezing level. Stratiform cloud base was observed to be at 4.8 km with a top at about 15 km. Mean vertical motion in the stratiform region was upward above the freezing level and downward below the freezing level.

2) INITIALIZATION

The model is initialized with a special dataset prepared by Australia's Bureau of Meteorology (BOM) for the EMEX time period. The BOM analysis includes observations from 1) the normal network of sounding and surface stations, 2) a special network of sounding and surface stations established during EMEX, and 3) temperature and wind data retrieved from satellite and aircraft data. This dataset provides the horizontal wind components, temperature, and relative humidity at 1.25° latitude-longitude intervals and at 11 pressure levels. The model uses topographic data that have a horizontal spacing of $10'$ latitude-longitude on all grids. Over the ocean, horizontally variable sea surface temperatures for February 1987 from a $2^\circ \times 2^\circ$ latitude-longitude NMC (now the National Centers for Environmental Prediction) Climate Analysis Center dataset are used. A constant soil moisture is specified over land areas. A constant vegetation type of mixed woodland is specified over land areas in accordance with the land cover map of Wilson and Henderson-Sellers (1985).

3) SETUP

The model's horizontal grid spacing is 24 km on Grid 1, 6 km on Grid 2, and 1.5 km on Grid 3 (Fig. 1). All grids have 35 vertical levels, stretched from a spacing of 100 m near the surface to 1000 m at the model top (~ 22 km). The Rayleigh friction absorbing layer is in the top five model levels. The long time steps are 24, 8, and 4 s for Grids 1, 2, and 3, respectively, with short time steps half as long. The simulation is started at 1200 UTC (2230 LST over Grid 3) 2 February 1987. During the first hour of the simulation, only the coarsest grid is used. At that point, the two finer grids are activated. The Level 2.5w convective parameterization scheme is used on all grids until 1330 UTC (90 min into the simulation); thereafter, it is not used at all. The EMEX9 MCS is explicitly simulated for 4.5 hours between 1330 and 1800 UTC.

4) RESULTS

The simulated EMEX9 is triggered by strong low-level convergence of two flows: the mesoscale land breeze circulation and the strong synoptic-scale circulation. The simulated land breeze circulation extends from the surface to about 2 km. The robust northerly onshore flow (which has veered from the northwesterlies observed 2 h earlier) collides with a very weak near-shoreline flow along a line that nearly parallels the coast.

Examination of the vertical velocity fields on the fine grid shows that at 1500 UTC, convective cells are arranged in a WNW-ESE-oriented line, with 500-mb vertical velocities ranging from -9 to $+21 \text{ m s}^{-1}$. Over the next 3 h, the whole mass of convection propagates toward the northeast at $10\text{--}15 \text{ m s}^{-1}$ along the monsoon trough toward the south coast of Papua New Guinea. The generally linear WNW-ESE orientation of the convective elements becomes less well defined as time goes on. The range of vertical velocities remains about the same through the entire 3-h period, although increasingly broader areas of weakly positive vertical velocities become apparent as the stratiform region evolves. The simulated orientation and propagation speed of the convection are supported by the observations. The simulated vertical velocities also are consistent with observations, although they are slightly large compared to aircraft observations of vertical velocity for all EMEX MCSs. For example, at an altitude of 5 km, the simulated EMEX9 10% updraft vertical velocity (i.e., 10% of updrafts exceed that velocity) is approximately 6.5 m s^{-1} compared to the 5 m s^{-1} observed by Lucas et al. (1994) for all EMEX MCSs. Also at 5 km, the EMEX9 10% downdraft velocity is simulated to be 2.5 m s^{-1} compared to the 2 m s^{-1} observed by Lucas et al. (1994) for all EMEX MCSs. Figure 2 shows the 0.5 g kg^{-1} surface of condensate mixing ratio on Grid 3 at 1500, 1600, and 1700 UTC. Figure 3 shows the structure of the convection at 1800 UTC, with the horizontal dis-

FIG. 1. Grid setup for the EMEX9 simulation. The horizontal grid spacing is 24 km, 6 km, and 1.5 km on Grids 1, 2, and 3, respectively.

tribution of 500-mb vertical velocity in Fig. 3a and a vertical cross section of condensate in Fig. 3b.

b. PRE-STORM 23–24 June simulation

1) OBSERVATIONS

The PRE-STORM 23–24 June MCS formed under classic synoptic conditions. At 1200 UTC 23 June a cold front trailed from a deep surface low near Hudson Bay, with the front becoming stationary across the central United States. By 0000 UTC 24 June a surface low located in Nebraska had deepened a bit with a dryline extending south from it into western Kansas. A stationary front snaked its way southeastward into the low and east-northeastward out of the low (see Fig. 1 of Stensrud and Maddox 1988). At this time, low-level air south of the front was hot and moist, with surface temperatures as high as 39°C and surface dewpoints as high as 24°C. Meanwhile, at 850 mb a strong southerly jet (maximum wind speeds $>15 \text{ m s}^{-1}$) over the southern Great Plains states provided a continuing supply of warm, moist air. The 500-mb height field contained a broad, rather weak ridge over the continental United States, with a shortwave trough passing through the ridge over the central plains (see Fig. 2b of Stensrud and Maddox). The 850-mb warm temperature advection

and the 500-mb shortwave teamed up to provide the primary forcing for the PRE-STORM 23–24 June MCS.

Convective cells first formed around 1900 UTC 23 June along the dryline and front. Convective cells in northern Kansas and southern Nebraska moved toward the east-southeast; those in central and southern Kansas moved toward the south. By 0000 UTC 24 June, convection in the northeastern part of the area had consolidated into a large MCS in eastern Nebraska and Iowa along and to the south of the front (courtesy of an old outflow boundary). Another area of thunderstorms was located over west-central Kansas along the dryline and eventually blossomed into a smaller MCS. The fine grid zeroes in on the latter MCS.

2) INITIALIZATION

The model initial fields were obtained by compositing several different data sources. The large-scale background features were obtained from European Centre for Medium-Range Weather Forecasts (ECMWF) analyses, which provide the horizontal wind components, temperature, and relative humidity at 2.5° latitude–longitude intervals and at 1000, 850, 700, 500, 300, 200, and 100 mb. To resolve finer-scale features, additional rawinsonde and surface observations from the National Weather Service/Federal Aviation Administration operational station

FIG. 2. The 0.5 g kg^{-1} condensate surface on Grid 3 at 1500 UTC, 1600 UTC, and 1700 UTC. The vertical line on the southwest corner of the grid extends from the surface to 20 km. Perspective is from the northeast.

network and the PRE-STORM station network supplemented the ECMWF data. At the lower boundary the model uses topographic data, which has a horizontal spacing of $10'$ latitude–longitude on Grid 1 and $30''$ latitude–longitude on the other grids. A vegetation-type dataset from the National Center for Atmospheric Research with 11 primary vegetation types at a $5'$ latitude–longitude resolution is interpolated onto the model grids and then converted to the vegetation classification used in the model (the model recognizes 18 vegetation types). In addition, horizontally variable soil moisture is based on the soil moisture analysis in the U.S. Department of Agriculture

publication *Weekly Weather and Crop Bulletin* (WWCB). The WWCB soil moisture index data were manually transferred to a latitude–longitude gridded dataset at 1° horizontal spacing. This dataset was then filtered and interpolated onto the model grid where it was converted into a soil moisture percentage.

3) SETUP

The startup time is 1200 UTC (0600 LST over the fine grid) 23 June 1985. Between 0000 and 1900 UTC, the model uses three grids, with horizontal grid spacings

EMEX9 Grid #3 1800 UTC Contour from -4 to 26 by 1

w (m/s) at 500 mb

10 km

EMEX9 Grid #3 1800 UTC Contour from 0 to 0.012 by 0.005

Condensate mixing ratio

10 km

FIG. 3. The (a) 500-mb vertical velocity and (b) condensate mixing ratio on Grid 3 at 1800 UTC 2 February. The vertical cross section in (b) is a north-south slice through the center of the grid.

of 75 and 25 km. A grid with 8.333-km horizontal spacing is added at 1900 UTC. Then, after 0000 UTC, a cloud-resolving grid is added (2.083 km) while the coarse 75-km grid is eliminated. Thus, after 0000 UTC, the model's horizontal grid spacing on the three grids is 25 km, 8.333 km, and 2.083 km. Hereafter, these grids are called Grids 1, 2, and 3, respectively (Fig. 4). There

are 32 vertical levels, stretched from a spacing of 175 m near the surface to 1000 m at the model top (~ 21 km). During the first 9 h of the simulation, only the 75-km and 25-km grids are used. The 8.333-km grid is activated at 2100 UTC and the 2.083-km grid is activated at 0000 UTC. Between 1900 and 0000 UTC, the Level 2.5w convective parameterization scheme is used on the 25-km and 8.333-km grids; thereafter, it is not used at all. Thus, the PRE-STORM 23–24 June 1985 MCS is explicitly simulated (no convective parameterization) between 0000 and 0400 UTC.

4) RESULTS

Before discussing the structure of the MCS on the cloud-resolving grid, we will examine the evolution of the synoptic fields and parameterized convection during the first 12 h of the simulation, between 1200 UTC 23 June and 0000 UTC 24 June. The simulated surface low is located along the Nebraska–Kansas border in good agreement with the 2100 UTC analysis of Johnson et al. (1989). Simulated convection first occurred at 2000 UTC 23 June in east-central Iowa along the simulated cold front. Late morning surface reports from this region noted that heavy rain, damaging winds, and golf-ball-sized hail occurred over this area. As the simulated convective system drifted eastward, more moist soil in eastern Iowa resulted in depressed surface sensible heat fluxes and ambient CAPE, causing the parameterized convection to subside. By 2030 UTC, the parameterized convection over this region had largely finished, although resolved precipitation continued to fall for the next several hours as the simulated clouds drifted eastward along the surface front.

Shortly after 2000 UTC 23 June, new convection associated with subsequent MCS development formed just to the south of the surface front along a line extending from the eastern half of the Iowa–Missouri border westward into southeast Nebraska. Extremely hot, muggy surface air, high CAPEs, and sufficiently strong resolved vertical velocities at cloud base ($\sim 0.04 \text{ m s}^{-1}$) ripened conditions for convection across this region. Several tornadoes, damaging winds, and baseball-sized hail were reported over this area between 2100 and 0200 UTC 24 June. The simulated east–west convective line continued to intensify between 2030 and 2200 UTC and propagated southward as new convective bands formed along the low-level convergence zone where the outflow from the old convection met the low-level jet in northeast Kansas and southeast Nebraska. During this time, simulated rainfall rates in south-central Iowa exceeded 5 cm h^{-1} , in good agreement with observations. By 0000 UTC 24 June, the parameterized convection on the 8.333-km grid was concentrated in two general areas: southeast Nebraska/northeast Kansas and south-central Iowa/north-central Missouri.

Because the convection was propagating southward, it was decided to place the cloud-resolving grid in north-

FIG. 4. Grid setup for the PRE-STORM simulation. The horizontal grid spacing is 25 km, 8.333 km, and 2.083 km on Grids 1, 2, and 3, respectively.

central Missouri and let the convection move into the grid. During this time, in agreement with observations, the simulated convection propagated southward at about 15 m s^{-1} . Comparison of simulated convection with contemporaneous radar observations shows generally good agreement between the simulation and observations. Convection entered Grid 3 at 0100 UTC and propagated southward. The convective elements maintained their generally east-west orientation over the entire period. As for EMEX9, broad areas of weak positive vertical velocities trail the convective elements. Figure 5 shows the vertical velocities and condensate mixing ratio fields on Grid 3 at 0130, 0230, and 0330 UTC. Figure 6 shows the structure of the convection at 0200 UTC, with the horizontal distribution of 500-mb vertical velocity in Fig. 6a and a vertical cross section of condensate in Fig. 6b.

4. Parameterization framework

a. Overview

In this section, we describe the framework of the parameterization scheme for the mesoscale drafts of

MCSs. Cumulus convection will be driven by the Arakawa–Schubert parameterization (e.g., Arakawa and Cheng 1993) modified to account for the effects of convective downdrafts following Johnson (1976). In order to account for a more physically realistic coupling between cumulus convection and associated stratiform cloudiness, the scheme employs a prognostic closure (as opposed to a quasi-equilibrium closure), as described by Randall and Pan (1993). Randall and Pan point out that for the purpose of parameterizing mesoscale effects, an objection to the quasi-equilibrium closure is that it is necessary to group the contributions to the time change of the cloud-work function A (an integral of the temperature and moisture over the convectively active layer) into “convective” and “nonconvective” components. By treating the effects of convectively produced stratiform clouds as nonconvective processes, quasi-equilibrium closure incorporates some aspects of the convective feedback into the large-scale forcing. With the Randall and Pan prognostic closure, this problem is sidestepped, as it is no longer necessary to distinguish between large-scale forcing and the convective

FIG. 5. The 0.5 g kg^{-1} condensate surface on Grid 3 at 0130 UTC, 0230 UTC, and 0330 UTC 24 June. The vertical line on the southwest corner of the grid extends from the surface to 20 km. Perspective is from the southeast.

response. The Randall and Pan modification of the Arakawa–Schubert scheme employs a prognostic equation for the cumulus kinetic energy K ,

$$\frac{\partial K}{\partial t} = M_B A - \frac{K}{\tau_D}, \quad (1)$$

where M_B is the convective cloud-base mass flux, A is the cloud-work function, and τ_D is a dissipation time-scale. Cloud-base mass flux is related to K through

$$K = \alpha M_B^2, \quad (2)$$

where α is an empirical parameter that Randall and Pan describe as being generally proportional to the magnitude of the vertical shear of the horizontal wind. Here,

we have used $\alpha = 10^8 \text{ m}^4 \text{ kg}^{-1}$ and $\tau_D = 600 \text{ s}$, following Randall and Pan. The parameterized cumulus convection provides condensed water, ice, and water vapor, which drives a parameterization for the large-scale effects of mesoscale circulations associated with the convection.

b. Tendency equations

Donner (1993) discusses the effects of cumulus convection on the large-scale fields of potential temperature $\bar{\theta}$ and water-vapor mixing ratio $\bar{q}$. These effects may be obtained by decomposing these fields into large-scale and small-scale components and then averaging the ther-

modynamic and moisture equations over the large scale. In isobaric coordinates, tendency equations for these variables are

$$\frac{d\bar{\theta}}{dt} = \frac{\pi \bar{Q}_r^e}{c_p} + \frac{\pi \sum_{i=1}^6 L_i \bar{\gamma}_i^e}{c_p} + \frac{\pi \bar{Q}_r^*}{c_p} + \frac{\pi \sum_{i=1}^6 L_i \bar{\gamma}_i^*}{c_p} - \frac{\partial \bar{\omega}' \bar{\theta}'}{\partial p} - \nabla \cdot \bar{\mathbf{v}}' \bar{\theta}' \quad (3)$$

and

$$\frac{d\bar{q}}{dt} = - \sum_{i=1}^4 \frac{|L_i|}{L_i} \bar{\gamma}_i^e - \sum_{i=1}^4 \frac{|L_i|}{L_i} \bar{\gamma}_i^* - \frac{\partial \bar{\omega}' \bar{q}'}{\partial p} - \nabla \cdot \bar{\mathbf{v}}' \bar{q}', \quad (4)$$

where $\bar{Q}_r$ is the radiative heating, c_p is the specific heat at constant pressure, and

$$\pi = \left(\frac{p_0}{p} \right)^{R_d/c_p},$$

where $p_0 = 100$ kPa and R_d is the gas constant for dry air. The summations represent phase transformations. The values of latent heat include condensation (L_1), evaporation (L_2), deposition (L_3), sublimation (L_4), freezing (L_5), and melting (L_6), respectively. Likewise, the phase transformations include γ_1 (condensation), γ_2 (evaporation), γ_3 (deposition), γ_4 (sublimation), γ_5 (freezing), and γ_6 (melting). Cloud properties and those of their environment are denoted by asterisks and superscripts e , respectively, while primes denote departures from the large-scale average. Equations (3) and (4) show that the effect of cumulus convection and mesoscale drafts on the thermodynamic structure of the large-scale environment depends on fluxes and phase transformations.

To compute fluxes, one needs to recognize that the vertical eddy transport of a property χ is given by

$$\overline{\omega' \chi'} = \sum_{i=1}^N (\overline{\omega' \chi'})_i, \quad (5)$$

where

$$(\overline{\omega' \chi'})_i = \frac{a_i \omega_i^* \chi_i^{*'}}{1 - a_i} \quad (6)$$

and a_i is the fractional area occupied by clouds of the i th of N subensembles.

Large-scale phase transformations due to cumulus convection or mesoscale drafts are given by

$$\bar{\gamma}_i^* = \sum_{j=1}^N a_j \gamma_{i,j}^*, \quad (7)$$

where $\gamma_{i,j}^*$ is the rate of the i th phase transformation per unit mass in an updraft or downdraft belonging to subensemble j .

Thus, for any given subensemble, evaluating the forc-

PRE-STORM Grid #3 0200 UTC Contour from -8 to 20 by 2

PRE-STORM 0200 UTC Grid #3 Contour from 0 to 0.0105 by 0.0005

FIG. 6. The (a) 500-mb vertical velocity and (b) condensate mixing ratio on Grid 3 at 0200 UTC 24 June. The vertical cross section in (b) is a north-south slice through the center of the grid.

ing of the large-scale flow requires its cloud temperature, water-vapor mixing ratio, and vertical velocity (T^* , q^* , and ω^* , respectively); its fractional area a ; and the rates of phase transformations. The approach presented in this study, therefore, is analogous to that of Donner (1993) except that the explicit MCS simulations allow us to gain much more insight into the vertical distri-

butions of these processes. Obtaining insight on the shapes of the vertical profiles of processes discussed in the following section requires conditional sampling of mesoscale updrafts and mesoscale downdrafts within our explicit MCS simulations (see section 5a). The MCS simulations also provide guidance for values of several other parameters needed to close the parameterization; most of these parameters involve relationships between various quantities in an MCS's water budget and are described in more detail in the following subsections. Figure 7 shows a summary of the physical processes considered in the MCS parameterization.

c. Mesoscale processes

1) WATER VAPOR REDISTRIBUTION BY MESOSCALE UPDRAFTS

Mesoscale updrafts that occur in the stratiform region can advect water vapor. As this water vapor is advected upward, it can change phase and contribute to latent heat release in the mesoscale updraft. Following Donner (1993), this process is considered to be the sum of 1) the redistribution of water vapor provided by cumulus updrafts only (considered in this subsection) and 2) advection of water vapor present in the environment of the cumulus updrafts but not supplied by the updrafts (considered in the next subsection).

The cumulus parameterization provides the following inputs to the mesoscale parameterization:

- 1) Production rate of ice ($\text{g g}^{-1}\text{s}^{-1}$)
- 2) Production rate of liquid water ($\text{g g}^{-1}\text{s}^{-1}$)
- 3) Production rate of water vapor ($\text{g g}^{-1}\text{s}^{-1}$)
- 4) Top level of cumulus penetration (mb)

The water vapor provided by the cumulus updrafts is denoted by Q'_{mf} (units: $\text{g g}^{-1}\text{s}^{-1}$). The mesoscale updraft vertically distributes Q'_{mf} at all levels in the mesoscale updraft over a period of time. Cloud base of the parameterized mesoscale updraft (p_{zm}) occurs at the mean 0°C level (570 mb for EMEX9; 590 mb for PRE-STORM). The top of the mesoscale updraft (p_{zmm}) coincides with the level of deepest cumulus penetration (150 mb for EMEX9, 190 mb for PRE-STORM).

At pressure p , Q'_{mf} contributes $\int_0^{\tau_m} Q'_{mf}(t) dt$ to the vertically averaged water-vapor mixing ratio in the stratiform region over its lifetime τ_m . The water vapor is distributed uniformly between p and $p + \int_0^{\tau_m} \omega_m dt$. This integral is set to be 30 kPa (unless this would distribute water vapor above the top of the mesoscale updraft, in which case it is limited to the depth of the mesoscale updraft).

The quantity $\bar{q}_c(p)$ (units: g g^{-1}) represents the sum of the redistribution of Q'_{mf} by all subensembles at all levels. As such, it represents the contribution of cumulus updrafts to the water-vapor mixing ratio in the mesoscale region. The redistributed water-vapor mixing ratio $\bar{q}_c$ accumulates in the mesoscale region whose fractional

coverage is a_m . The redistributed water-vapor mixing ratio in the mesoscale region is then $\bar{q}_c/a_m$. This quantity may exceed saturation and then change phase.

2) DEPOSITION AND CONDENSATION WITHIN MESOSCALE UPDRAFTS

The parameterization of the summed deposition and condensation is again patterned after Donner (1993). The region within the mesoscale updraft is assumed to consist of ice and liquid water, which are furnished by deposition/condensation from water vapor and by transfer from the convective region. This transfer of ice and liquid is represented by C_A (units: $\text{g g}^{-1}\text{s}^{-1}$).

The first source of condensate is provided by the redistribution by the mesoscale updraft of water vapor that is provided by cumulus updrafts only. The rate of deposition/condensation by this process is

$$\frac{\bar{q}_l(p)}{a_m \tau_m} - \frac{q_s[T_m(p)]}{\tau_m}, \quad (8)$$

where q_s denotes saturation mixing ratio and T_m refers to the temperature in the mesoscale updraft.

The preceding process deals only with water vapor supplied by the cumulus updrafts; additional deposition/condensation occurs as large-scale water vapor in the mesoscale region surrounding the updrafts is lifted by mesoscale ascent. This process is parameterized in terms of the water-vapor mixing ratio at the base of the mesoscale updraft, which is conserved as it undergoes mesoscale ascent until deposition/condensation begins, when

$$\bar{q}(p_{zm}) + \frac{\bar{q}_l(p_{zm})}{2a_m} = q_s[T_m(p)]. \quad (9)$$

Deposition/condensation then proceeds at a rate,

$$\omega_m \frac{\partial q_s}{\partial p}$$

(the factor of 2 averages the water vapor from the cumulus updrafts over τ_m).

These two processes yield a value for the vertically integrated deposition plus condensation, $(\int_{p_{zmm}}^{p_{zm}} C_{mu} dp)$. The shapes of the vertical profiles of deposition and condensation and the ratio of vertically integrated deposition to vertically integrated condensation within the mesoscale updraft are then determined by examining the vertical structure of these processes within the conditionally sampled mesoscale updrafts of the explicit MCS simulations.

3) FREEZING IN MESOSCALE UPDRAFTS

Vertically integrated freezing in mesoscale updrafts is taken to be the sum of two processes. First, all liquid water transferred from the cumulus convection to the mesoscale updraft is assumed to freeze. This liquid wa-

FIG. 7. Summary of the physical processes considered in the MCS parameterization.

ter is provided by the Arakawa-Schubert cumulus parameterization. Second, all of the water formed through the condensation process described in the preceding subsection is assumed to freeze within the mesoscale updraft. The shape of the vertical profile of the net freezing within the mesoscale updraft is then determined by examining the vertical profile of net freezing in the conditionally sampled mesoscale updrafts of the explicit MCS simulations.

4) SUBLIMATION AND EVAPORATION IN MESOSCALE DOWNDRAFTS

As in the preceding subsection, the vertically integrated value of sublimation plus evaporation within the parameterized mesoscale downdraft is determined through consideration of an MCS's water budget, with the ratio of $g^{-1} \int_0^{p_s} E_{md} dp$ to $(g^{-1} \int_0^{p_s} C_{mu} dp + C_A)$ determined from a water budget of the explicit simulations.

The partitioning of E_{md} between sublimation and evaporation as well as the shapes of the vertical profiles of these two processes are also determined by examining the vertical profiles of sublimation and evaporation in the conditionally sampled mesoscale downdrafts of the explicit MCS simulations.

5) MELTING IN MESOSCALE DOWNDRAFTS

All mesoscale precipitation is assumed to fall out of the mesoscale updraft as ice. Melting may then come from one of two sources: 1) melting of all condensate that reaches the ground as rain (R_m) and 2) melting of ice that falls below 0°C and then evaporates in the mesoscale downdraft. The value of the first component (R_m) is computed from $\int (C_{mu} + C_A) dp$, along with an assumption regarding the precipitation efficiency of the mesoscale region [Eq. (17)]. These two sources are added to yield the magnitude of the vertically integrated melting in parameterized mesoscale downdrafts. The shape of the vertical profile of melting is then determined by examining the vertical profiles of melting within the conditionally sampled mesoscale downdrafts of the explicit MCS simulations.

6) MESOSCALE EDDY FLUXES OF ENTROPY AND MOISTURE

Eddy fluxes of entropy and water vapor are computed using (6), where the shapes of the vertical profiles of the perturbation potential temperature, perturbation water-vapor mixing ratio, and perturbation vertical velocity in parameterized mesoscale updrafts and downdrafts are determined through conditional sampling of the explicit simulations.

5. Scheme construction

a. Conditional sampling

Construction of the convective parameterization scheme described here requires vertical profiles of various quantities in conditionally sampled mesoscale updrafts and downdrafts. The first step in this conditional sampling process is to separate the simulated MCSs into convective and stratiform regions; mesoscale updrafts and mesoscale downdrafts are confined to the stratiform regions. Here, we employ the convective/stratiform partitioning technique of Tao et al. (1993). In the Tao et al. (1993) technique, model grid columns exhibiting a surface precipitation rate twice as large as the average value taken over the surrounding grid columns are identified as convective cells. For each core grid column, all adjacent grid columns are also taken to be convective. In addition, any grid column with a rain rate in excess of 25 mm h^{-1} is considered as convective regardless of the above criteria, and any grid column with no surface precipitation is considered convective if the

maximum updraft exceeds 5 m s^{-1} . All other grid columns precipitating at the surface are considered to be stratiform.

In both simulated MCSs, the stratiform region accounts for between 20% and 60% of total simulated precipitation, depending on the stage of the life cycle. Overall, for the EMEX9 simulation, the stratiform region contributes to approximately 44% of the total precipitation. By comparison, the Tao et al. (1993) simulation of EMEX9 resulted in a 42% contribution from the stratiform region. Both numbers are consistent with observations of other tropical MCSs (e.g., Houze 1977; Zipser et al. 1981). Overall, for the PRE-STORM 23–24 June simulation, the stratiform region contributes to approximately 32% of the total precipitation. Although there are no other independent simulations of the PRE-STORM 23–24 June MCS to compare with this number, it is consistent with observations from midlatitude MCSs (e.g., Johnson and Hamilton 1988).

After separating the MCSs into two-dimensional convective and stratiform regions, the mesoscale updrafts and mesoscale downdrafts are isolated within the stratiform region. The aims here are 1) to ensure that the mesoscale updrafts or downdrafts are spatially coherent and 2) to allow the mesoscale updrafts and downdrafts to have some vertical structure. In light of these requirements, the conditional sampling is done on a grid-point-by-gridpoint basis (not a column-by-column basis). A stratiform region grid point is considered to be within a mesoscale updraft if 1) there is condensate present, 2) the vertical velocity is upward, 3) conditions 1) and 2) are satisfied at *all* adjacent grid points, and 4) the grid point is located above the 0°C level. A stratiform region grid point is considered to be within a mesoscale downdraft if all of the following conditions are met: 1) the vertical velocity is downward, 2) condition 1) is satisfied at *all* adjacent grid points, and 3) the grid point is located below the 0°C level. Although the simulations show that some mesoscale updrafts occur below the freezing level and that some mesoscale downdrafts occur above the freezing level, conditional sampling of these drafts reveals that their net heating and drying rates are small; thus, they are neglected here. Vertical profiles of physical processes in these conditionally sampled mesoscale updrafts and downdrafts will be used to determine the shapes of vertical profiles of various physical processes as well as relationships between various components of an MCS's water budget. To build the MCS parameterization, EMEX9 simulation data are examined for 1700–1800 UTC and the PRE-STORM simulation data for 0300–0400 UTC.

b. Shape and depth of parameterized curves

The conditionally sampled data are used, in part, to provide insight into the shape and depth of the vertical profiles of various parameterized physical processes. To a first approximation, all of the curves are symmetric

with a sinusoidal form. Occasionally, however, the departure from symmetry may be significant. For instance, as one would expect, the vertical profiles of freezing rates in the conditionally sampled mesoscale updrafts of each MCS indicate that the peak freezing rate is located much closer to the bottom of the mesoscale updraft than to the top of the mesoscale updraft. To describe the precise shape of a curve, each curve is given a shape parameter. The shape parameter may range between 0 and 1, with a value of 0 indicating a vertical profile which has a maximum value at its bottom side, a value of 1 indicating a vertical profile which has a maximum value at its top side, and a value of 0.5 indicating a vertical profile which is perfectly symmetric. A depth parameter completes the description of the vertical profile. The depth parameter also may range between 0 and 1, with a value of 0 indicating that the profile has no depth and a value of 1 indicating that the profile extends fully through the extent of the mesoscale updraft or downdraft. All profiles for parameterized mesoscale downdrafts have a top at the 0°C level; all profiles for parameterized mesoscale updrafts have a base at the 0°C level.

Consider then a vertical profile of a process q with shape parameter S and depth parameter D , which is placed within a mesoscale updraft that extends from pressure level p_{zm} (bottom of the mesoscale updraft) to pressure level p_{zm} (top of the mesoscale updraft). The vertical profile of q extends from p_{bq} (bottom of curve q) to p_{pq} (top of curve q), where p_{bq} and p_{pq} are

$$p_{bq} = p_{zm} \quad (10)$$

and

$$p_{pq} = p_{zm} - D(p_{zm} - p_{tm}), \quad (11)$$

respectively. Then, if the quantity x varies from 0 at $p = p_{pq}$ to 1 at $p = p_{bq}$ as

$$x = \frac{p - p_{pq}}{p_{bq} - p_{pq}}, \quad (12)$$

the normalized value of q at pressure p is

$$q(p) = \sin\left(\frac{\pi x}{2S}\right) \quad (13)$$

for $x \leq S$ and

$$q(p) = \sin\left[\frac{\pi}{2(1-S)}(x + 1 - 2S)\right] \quad (14)$$

for $x > S$.

The vertical profile of each phase transformation process is assigned a shape and depth parameter. The conditionally sampled data provide guidance as to appropriate values. In practice, of course, one would not know exactly which values to use for these parameters—the numbers given in the following sections simply represent a reasonable range. The sensitivity of the param-

eterization to the prescribed values of these shape and depth parameters is examined in section 6.

c. Phase transformation rates

The bulk microphysics scheme used in the two MCS simulations is based on Cotton et al. (1986) and discussed in detail by Flatau et al. (1989). Here, we are particularly interested in conversions between the microphysical categories (cloud water, rain, pristine crystals, snow, graupel, and aggregates). Parameterized conversion processes include collection, vapor deposition/evaporation, melting, and ice nucleation (sorption/deposition, phoretic contact, and splintering).

The budget of condensed water in the mesoscale region of an MCS may be described by

$$\begin{aligned} R_m + \frac{1}{g} \int_0^{p_g} E_{me} dp + \frac{1}{g} \int_0^{p_g} E_{md} dp \\ = g^{-1} \int_0^{p_g} C_{mu} dp + C_A, \end{aligned} \quad (15)$$

where R_m is the mesoscale rainfall, E_{me} is the sublimation from mesoscale updrafts into the environment, E_{md} is the sublimation plus evaporation in mesoscale downdrafts, C_{mu} is the deposition plus condensation in mesoscale updrafts, and C_A is the condensate transferred from the convective region into the mesoscale region. The first term on the right side of (15) is determined as described in section 2c(2); the second term on the right side of (15) is provided by the convective parameterization. To complete the water budget, we need the ratios among the three terms on the left side of (15). Equation (15) may be rewritten as

$$a + b + c = 1, \quad (16)$$

where

$$a = \frac{R_m}{g^{-1} \int_0^{p_g} C_{mu} dp + C_A}, \quad (17)$$

$$b = \frac{g^{-1} \int_0^{p_g} E_{md} dp}{g^{-1} \int_0^{p_g} C_{mu} dp + C_A}, \quad (18)$$

and

$$c = \frac{g^{-1} \int_0^{p_g} E_{me} dp}{g^{-1} \int_0^{p_g} C_{mu} dp + C_A}. \quad (19)$$

Figure 6 shows the value of a , b , and c in the EMEX

TABLE 1. Values of MCS water budget parameters a , b , and c for cases A, B, and C of Leary and Houze (1980, LH); cases I and II of Gamache and Houze (1983, GH); Chong and Hauser (1989, CH); Roux and Ju (1990, RJ); Gallus and Johnson (1991, GH); EMEX9 (1700–1800 UTC mean, E9); and PRE-STORM 23–24 June (0300–0400 UTC mean, PS).

Parameter	LH-A	LH-B	LH-C	GH-I	GH-II	CH	RJ	GJ	E9	PS
a	0.91	0.50	0.50	0.64	0.64	0.52	0.61	0.54	0.92	0.87
b	0.09	0.10	0.10	0.16	0.16	0.16	0.11	0.20	0.06	0.10
c	0.00	0.40	0.40	0.20	0.27	0.38	0.28	0.27	0.02	0.13

(1700–1800 UTC) and PRE-STORM (0300–0400 UTC) simulations. For EMEX, a ranges between 0.91 and 0.93, b ranges between 0.05 and 0.06, and c ranges between 0.02 and 0.03. For PRE-STORM, a ranges between 0.80 and 0.92, b ranges between 0.07 and 0.14, and c ranges between 0.02 and 0.05. Table 1 lists values of a , b , and c determined by other investigators for various MCSs in the midlatitudes and Tropics.

The studies summarized in Table 1 use widely varying techniques to evaluate these quantities and therefore directly comparing one to another is dubious. Methodology aside, it is expected that there will also be physical differences in these quantities from one MCS to the next that will be particularly dependent on which stage of the MCS's life cycle is sampled. For example, consider the 10–11 June PRE-STORM MCS. At least two different values for the precipitation efficiency for this storm appear in the published literature. Gallus and Johnson (1991) observed a precipitation efficiency (or a value) of 0.54 for this system. On the other hand, Tao et al. (1993) obtained a value of 0.23 in their 2D simulation. Although the two values differ by more than a factor of 2, neither one is necessarily wrong. Note that our values are representative of the mature stage of the life cycle for each MCS, when precipitation efficiencies are approaching one.

1) DEPOSITION AND CONDENSATION IN MESOSCALE UPDRAFTS

In section 4c(2) we discuss how the magnitude of the vertically integrated vapor deposition plus condensation in the mesoscale updraft is parameterized. To complete the parameterization it is necessary to know 1) the ratio of the vertically integrated condensation to the vertically integrated deposition and 2) the shapes of the vertical profiles of deposition and condensation individually.

Figure 8 shows the ratio of vertically integrated condensation to vertically integrated deposition

$$\frac{\int C_{mu} dp}{\int D_{mu} dp}$$

in the conditionally sampled mesoscale updrafts for each MCS simulation for the 1-h periods under consideration. The values of $\int C_{mu} dp / \int D_{mu} dp$ range from about 0.1 to 0.3 for EMEX9 and 0.8 to 2.6 for PRE-STORM, and with time for a given simulation. Here, we are seeing the mature stage of each MCS. The parameter varies significantly between simulations. Still, within the extreme values given here the results of the parameterization are not particularly sensitive to the value of this parameter. This suggests that, in practice, the parameterization could be simplified to just consider the sum of deposition and condensation in mesoscale updrafts rather than treating the processes separately.

Figure 9 shows a time–height plot of the deposition rate in the conditionally sampled mesoscale updrafts for the EMEX9 (1700–1800 UTC) and PRE-STORM (0300–0400 UTC) simulations. Based on the character of the vertical profiles in Fig. 9, the parameterized vertical profile of deposition rate in the mesoscale updraft is assumed to have a shape parameter of 0.85 for EMEX and 0.50 for PRE-STORM and a depth parameter of 0.80 for both EMEX and PRE-STORM.

Figure 10 shows a time–height plot of the condensation rate in the conditionally sampled mesoscale updrafts for the EMEX9 (1700–1800 UTC) and PRE-STORM (0300–0400 UTC) simulations. Based on the character of the vertical profiles in Fig. 10, the parameterized vertical profile of condensation rate in the mesoscale updraft is assumed to have a depth parameter of 0.40 for EMEX and 0.70 for PRE-STORM and a

FIG. 8. Time evolution of the ratio of the vertically integrated condensation rate to the vertically integrated deposition rate in the conditionally sampled mesoscale updrafts for EMEX9 (solid) and PRE-STORM (dashed).

FIG. 9. Time evolution of the deposition rate in the conditionally sampled mesoscale updrafts for (a) EMEX9 and (b) PRE-STORM.

FIG. 10. Time evolution of the condensation rate in the conditionally sampled mesoscale updrafts for (a) EMEX9 and (b) PRE-STORM.

shape parameter of 0.85 for both EMEX and PRE-STORM.

2) FREEZING IN MESOSCALE UPDRAFTS

The magnitude of freezing in the parameterized mesoscale updraft is determined as discussed in section 4c(3). The shape of the freezing profile is determined

through conditional sampling of the explicit MCS simulations. Figures 11 and 12 show time–height plots of the freezing rate and the MCS water budget parameters a , b , and c for the EMEX9 and PRE-STORM simulations. Based on these vertical profiles, the shape parameter has a value of 0.75 for EMEX and 0.65 for PRE-STORM; the depth parameter is 0.6 for EMEX and 0.5

for PRE-STORM. Freezing, of course, only serves to modify the shape of the parameterized vertical profile of $d\bar{\theta}/dt$ and will have no effect on the parameterized vertical profile of $d\bar{q}/dt$.

3) SUBLIMATION AND EVAPORATION IN MESOSCALE DOWNDRAFTS

The magnitude of the vertical integral of sublimation plus evaporation in mesoscale downdrafts is given by

$$\frac{1}{g} \int_0^{p_g} E_{md} dp = b \left[\frac{1}{g} \int_0^{p_g} C_{mu} dp + C_A \right], \quad (20)$$

where b is as defined in (18). To determine how to partition this quantity into vertically integrated sublimation and vertically integrated evaporation individually, we look at the ratio between these quantities in the conditionally sampled downdrafts of the explicit simulations. Figure 13 shows the ratio of the vertically integrated evaporation in conditionally sampled mesoscale downdrafts to the vertically integrated sublimation in conditionally sampled mesoscale downdrafts for both EMEX9 and PRE-STORM. In each case, the vertically integrated evaporation rate exceeds the vertically integrated sublimation rate with the ratio ranging from 10 to 16 for EMEX9 and from 3 to 11 for PRE-STORM. The parameterization does show some sensitivity to the value of this parameter (e.g., see section 6b for a discussion of the sensitivity for PRE-STORM).

Figure 14 shows a time–height plot of the evaporation rate in the conditionally sampled mesoscale downdrafts for the EMEX9 (1700–1800 UTC) and PRE-STORM (0300–0400 UTC) simulations. Based on the character of the vertical profiles in Fig. 14, the vertical profile of evaporation rate in the parameterized mesoscale downdraft is assumed to have a shape parameter of 0.25 for EMEX and 0.85 for PRE-STORM. The depth parameter is 1.0 for each case.

Figure 15 shows a time–height plot of the sublimation rate in the conditionally sampled mesoscale downdrafts for the EMEX9 (1700–1800 UTC) and PRE-STORM (0300–0400 UTC) simulations. The vertical profiles in Fig. 15 yield a shape parameter of 0.50 for EMEX and 0.65 for PRE-STORM for sublimation in the parameterized mesoscale downdraft. The depth parameter is 0.20 for both EMEX and PRE-STORM.

4) MELTING

The magnitude of the vertically integrated melting in mesoscale downdrafts is computed as discussed in section 4c(6). That is, the melting depends on the magnitude of the parameter a , as defined in (17). Melting profiles for the conditionally sampled mesoscale downdrafts of each MCS are shown in Fig. 16. These profiles of melting in conditionally sampled mesoscale downdrafts yield a depth parameter of 0.60 for EMEX and

FIG. 11. Time evolution of the heating rate due to freezing in the conditionally sampled mesoscale updrafts for (a) EMEX9 and (b) PRE-STORM.

0.80 for PRE-STORM and yield a shape parameter of 0.45 for EMEX and 0.40 for PRE-STORM.

5) EDDY FLUX CONVERGENCES

The mesoscale eddy flux convergences of entropy and water vapor are the final terms needed to parameterize

FIG. 12. Time evolution of the MCS water budget parameters a , b , and c in the conditionally sampled mesoscale updrafts and downdrafts for EMEX9 and PRE-STORM. The quantities a , b , and c are defined in the text.

mesoscale heating and drying, respectively. The vertical profiles of the eddy flux entropy convergence show features analogous to a similar analysis (except for in a cumulus ensemble model) shown in Fig. 8 of Xu (1995). A time series of heating resulting from the eddy flux convergence of entropy for conditionally sampled updrafts and downdrafts for EMEX9 (Fig. 17a) shows that the important features of the profile are heating at the top of the troposphere and cooling of similar magnitude just beneath there. In the lower troposphere, particularly after 0345 UTC, there is a layer of cooling sandwiched between two layers of heating, as in Xu (1995). The

FIG. 13. Time evolution of the ratio of the vertically integrated evaporation rate to the vertically integrated sublimation rate in conditionally sampled mesoscale downdrafts for (a) EMEX9 (solid) and (b) PRE-STORM (dashed).

FIG. 14. Time evolution of the evaporation rate in the conditionally sampled mesoscale downdrafts for (a) EMEX9 and (b) PRE-STORM.

PRE-STORM 0300-0400 UTC time series of the vertical profile of heating due to eddy flux convergence of entropy (Fig. 17b) shows similar features to EMEX9 and Xu (1995). In both cases, the magnitude of heating from the eddy flux convergence term is relatively small in comparison to that by the phase change terms, except perhaps in the upper troposphere where the phase

FIG. 15. As in Fig. 14 but of the sublimation rate.

FIG. 16. As in Fig. 14 but of the melting rate.

change terms are small. The magnitude of the eddy heat flux convergences presented here are comparable to those found by Xu (1995). For EMEX9, the heating ranges from approximately -0.60 to $+0.75 \text{ K d}^{-1}$; for PRE-STORM, the heating ranges from -1.40 to $+0.75 \text{ K d}^{-1}$.

The vertical profiles of the eddy moisture convergence in the conditionally sampled regions also show features similar to those in Xu's Fig. 9. For EMEX9

(Fig. 18a) the mesoscale updrafts effect upper-tropospheric moistening and midtropospheric drying. In mesoscale downdrafts, the strongest feature is drying below 950 mb, which reaches a magnitude of about $0.75 \text{ g kg}^{-1} \text{ d}^{-1}$. For PRE-STORM (Fig. 18b), the features are similar. Mesoscale updrafts effect upper-tropospheric moistening and midtropospheric drying. In mesoscale downdrafts, moistening also overlies drying, with evi-

FIG. 17. Time evolution of the vertical profiles of the heating resulting from the eddy flux convergence of entropy for the conditionally sampled stratiform region of (a) EMEX9 (1700–1800 UTC) and (b) PRE-STORM 23–24 June (0300–0400 UTC). Dashed lines indicate negative contour values.

FIG. 18. Time evolution of the vertical profiles of drying resulting from the eddy flux convergence of water vapor for the conditionally sampled stratiform region of (a) EMEX9 (1700–1800 UTC) and (b) PRE-STORM 23–24 June (0300–0400 UTC). Dashed lines indicate negative contour values.

dence of a secondary drying layer around 850 mb, as seen in Xu's Fig. 9.

The mesoscale eddy flux convergence of a variable χ (here χ is either θ or q) is parameterized as specified in (6) except a_i is replaced by a_m , the parameterized fractional coverage of the stratiform region. That is,

$$\frac{\partial \omega' \chi'}{\partial p} = - \frac{a_m}{1 - a_m} \frac{\partial \omega'_m \chi'_m}{\partial p}. \quad (21)$$

Thus, to parameterize eddy flux convergence profiles the shapes and magnitudes of vertical profiles of ω' , θ' , and q' are specified according to the results of conditional sampling of mesoscale updrafts and downdrafts. The shape and depth parameters and maximum magnitudes of these profiles are summarized in Tables 2–4.

Fractional coverage of the stratiform region is parameterized based on observational evidence, following Leary and Houze (1980) and Donner (1993), who both

TABLE 2. Values of shape parameter, depth parameter, and maximum magnitudes of ω' in conditionally sampled mesoscale updrafts and downdrafts of the EMEX9 and PRE-STORM 23–24 June simulations.

	Depth parameter	Shape parameter	Max ω' value (m s^{-1})
EMEX9 updrafts	1.0	0.20	0.30
downdrafts	1.0	0.20	−0.20
PRE-STORM updrafts	1.0	0.50	1.40
downdrafts	1.0	0.20	−0.30

assumed that the stratiform region's fractional area is five times that of the convective region fractional area. Because the Arakawa–Schubert scheme does not predict cumulus fractional area, this is parameterized following Weissbluth and Cotton (1993). They parameterize cumulus updraft fractional coverage based on observational evidence of the diameter of the updraft core and its associated environment (which is assumed to comprise weaker updrafts as well as any compensating subsidence around the cloud). The area of the core and its associated environment (η) is stratified according to a bulk Richardson number (Weisman and Klemp 1982).

6. Scheme evaluation

a. EMEX9 simulation

The scheme described in previous sections requires an input sounding of pressure, temperature, and moisture, and then provides as output the vertical profiles of the mesoscale heating rate (shown here as K d^{-1}) and the mesoscale drying rate (shown as $\text{g kg}^{-1} \text{d}^{-1}$). The scheme will be tested by comparing its output with the diagnosed heating and drying resulting from phase transformations and eddy flux convergences in conditionally sampled mesoscale updrafts and downdrafts. For EMEX9, the input sounding will be the mean Grid 2 sounding at a given time.

It is acknowledged that all of the parameters discussed in section 5 vary from one MCS to the next and even vary during the life cycle of a particular MCS. Although the values of all the parameters discussed in section 5 were chosen by examining EMEX9 model output from 1700–1800 UTC, we have tested the scheme with soundings from 1400–1800 UTC. Figures 19 and 20 show a comparison of diagnosed and parameterized heating and drying rates, respectively, for 1400–1800 UTC. These results show that the parameterization performs well for the entire 4-h period. In particular, the parameterization is able to reproduce the maximum values of heating and drying diagnosed at 1400, 1700, and 1745 UTC. The only systematic errors, which appear in Figs. 19 and 20, are slightly overestimated cooling and moistening in mesoscale downdrafts. Although the magnitude of the errors is quite small ($\sim 0.5 \text{ K d}^{-1}$ too much cooling and $0.1\text{--}0.2 \text{ g kg}^{-1} \text{d}^{-1}$ too much mois-

TABLE 3. Values of shape parameter, depth parameter, and maximum magnitudes of θ' in conditionally sampled mesoscale updrafts and downdrafts of the EMEX9 and PRE-STORM 23–24 June simulations.

	Depth parameter	Shape parameter	Max θ' value (K)
EMEX9 updrafts	1.0	0.20	0.30
downdrafts	1.0	0.50	0.60
PRE-STORM updrafts	1.0	0.50	0.05
downdrafts	1.0	0.50	−0.10

tening), these errors appear over a layer about 200 mb deep in the mesoscale downdraft. Examination of the individual terms reveals that the main source of error in the updraft is the parameterized freezing rate, which reaches a maximum value of only about 1 K d^{-1} in the mesoscale updraft as compared to the diagnosed value of $\sim 3.5 \text{ K d}^{-1}$. However, there are no systematic errors in parameterized freezing rates—at some times the agreement is nearly perfect; at other times freezing is overpredicted.

At 1700 UTC, the maximum heating rate in the parameterized mesoscale updraft is about 8 K d^{-1} ; the maximum cooling rate in the parameterized mesoscale downdraft is about 3 K d^{-1} . The maximum drying rate in the parameterized mesoscale updraft is $2 \text{ g kg}^{-1} \text{d}^{-1}$; the maximum moistening rate in the parameterized mesoscale downdraft is $0.5 \text{ g kg}^{-1} \text{d}^{-1}$. By comparison, the maximum parameterized convective heating rate is 55 K d^{-1} and the maximum parameterized convective drying rate is $36 \text{ g kg}^{-1} \text{d}^{-1}$ at this time (not shown).

Although it is not the focus of this paper to evaluate the performance of the convective portion of the parameterization, it performed very well for both EMEX9 and PRE-STORM when compared to the diagnosed convective heating and drying. For example, for convective heating rates for EMEX9 at 1700 UTC, errors were less than 3 K d^{-1} at all vertical levels. On the other hand, when parameterized convective heating is compared to total heating (convective plus mesoscale), more significant errors appeared (as large as 15 K d^{-1}). These errors are the result of failing to parameterize mesoscale updrafts and downdrafts. When the mesoscale effects were parameterized, the errors were reduced significantly.

TABLE 4. Values of shape parameter, depth parameter, and maximum magnitudes of q' in conditionally sampled mesoscale updrafts and downdrafts of the EMEX9 and PRE-STORM 23–24 June simulations.

	Depth parameter	Shape parameter	Max q' value (g kg^{-1})
EMEX9 updrafts	1.0	0.40	0.07
downdrafts	1.0	0.30	−0.07
PRE-STORM updrafts	1.0	0.80	0.50
downdrafts	1.0	0.80	−0.50

FIG. 19. The (a) diagnosed and (b) parameterized heating rates for the EMEX9 simulation between 1400 and 1800 UTC.

FIG. 20. As in Fig. 19 but for drying rates.

b. PRE-STORM simulation

For PRE-STORM, as for EMEX9, the input sounding for each time will be the mean Grid 2 sounding. The scheme has been built using conditionally sampled data from Grid 3 data of the PRE-STORM simulation from 0300–0400 UTC. Figure 21 shows the diagnosed and parameterized heating rates for the PRE-STORM simulation for between 0100 and 0400 UTC. Because the

contribution of the mesoscale eddy fluxes are so small, the structure of the corresponding drying profiles strongly resembles the heating profiles (not shown). Compared to EMEX9, the parameterized heating and drying rates show relatively little variation with time.

Although heating and drying values in the updraft are generally well simulated, there are more significant errors in the downdraft, with excessive cooling and drying predicted at most times. Here, the primary source of error is that the parameterized evaporation rates are

FIG. 21. As in Fig. 19 but for heating rates for the PRE-STORM simulation between 0100 and 0400 UTC.

too large compared to the observed evaporation rates. After 0300 UTC, the value of the b parameter used in the parameterization is just too large, resulting in too much evaporation. At times prior to 0300 UTC more of this evaporation should have been treated as sublimation, which would not only improve the magnitude of the cooling but the shape profile as well. Refer again to Fig. 13, which shows the ratio of the vertically integrated evaporation to the vertically integrated subli-

mation during the hour of conditional sampling. In fact, for the previous hours of the simulation, this ratio was consistently lower. So a deceptively high value of this ratio was used at most times. This also explains why the errors are most significant in the lowest 200 mb; for PRE-STORM the conditional sampling shows that most of the evaporation is concentrated here (see Fig. 14).

The errors in the updraft heating at the later times in Fig. 21 are related to underestimates in parameterized deposition rates. See, for example, Fig. 9, which shows continuously increasing deposition rates at later times in the simulation compared to previous times. The errors in the mesoscale parameterization for PRE-STORM highlight the difficulties in assigning a single value to water budget parameters, which are actually life cycle dependent.

The parameterized mesoscale heating and drying tendencies for the PRE-STORM simulation account for a smaller fraction of the total heating and drying tendencies than for EMEX9. At 0400 UTC, the maximum heating rate in the parameterized mesoscale updraft is about 10 K d^{-1} ; the maximum cooling rate in the parameterized mesoscale downdraft is about 3.8 K d^{-1} . The maximum drying rate in the parameterized mesoscale updraft is $5.0 \text{ g kg}^{-1} \text{ d}^{-1}$; the maximum moistening rate in the parameterized mesoscale downdraft is $1.0 \text{ g kg}^{-1} \text{ d}^{-1}$. By comparison, the maximum parameterized convective heating rate is 165 K d^{-1} and the maximum parameterized convective drying rate is $34 \text{ g kg}^{-1} \text{ d}^{-1}$ at this time (not shown). Thus, although the absolute magnitudes of the mesoscale heating and drying are about the same for EMEX9 and PRE-STORM, their magnitudes relative to the convective heating and drying are greater for EMEX9. These results are consistent with those of Wu (1993). He found, for instance, that for six PRE-STORM observation times, the maximum magnitude of mesoscale heating was about 17 K d^{-1} compared to a maximum magnitude of convective heating of about 125 K d^{-1} (his Figs. 21–23). For six GARP Atlantic Tropical Experiment observation times, on the other hand, Wu (1993) found a maximum mesoscale heating rate of 4 K d^{-1} versus a maximum convective heating rate of 15 K d^{-1} .

As discussed above, the values of all parameters determined through conditional sampling would be expected to vary among different MCSs and throughout the life cycle of a single MCS. So far we have seen how sensitive the scheme is to the variation of these parameters through an MCS's life cycle. What happens if we now *completely* change these parameters, but still maintain plausible values? To answer this question, we will run the parameterization using 1) the PRE-STORM soundings for 0100 to 0400 UTC and 2) the EMEX9 parameters (all water budget parameters, depth, and shape parameters for phase transformations). The parameterized heating tendencies for this situation are compared to the PRE-STORM diagnosed heating and drying in Fig. 22. At 0400 UTC, the maximum para-

FIG. 22. The parameterized heating rates for the PRE-STORM simulation between 0100 and 0400 UTC, where PRE-STORM parameters are replaced with the EMEX9 parameters (see text).

meterized heating in the mesoscale updraft is underestimated by $\sim 4 \text{ K d}^{-1}$, with the maximum value displaced about 100 mb too low. The maximum parameterized cooling in the mesoscale downdraft at 0400 UTC is overestimated by about a factor of 2. For the 0400 UTC drying profiles (not shown), the nature of the errors is similar, with errors of $1\text{--}2 \text{ g kg}^{-1} \text{ d}^{-1}$ in the mesoscale updraft, and $<0.5 \text{ g kg}^{-1} \text{ d}^{-1}$ in the mesoscale downdraft. Actually, in the downdraft, the errors for this test are smaller than the for the test shown in Fig. 21. In the updraft, the mean error is slightly larger with the main problem being a skewing of the shape of the heating and drying to lower levels in the mesoscale updraft, as is characteristic of the EMEX9 mesoscale updrafts. Whether these errors are acceptable in the context of GCM simulations can only be determined when the present scheme is applied to global simulations. This work is on going, as the scheme is presently being interfaced with a global version of RAMS.

7. Summary

We have described the application of cloud-resolving simulations of two MCSs to the construction of an MCS parameterization scheme. The MCS simulations include one of a tropical MCS observed during the 1987 Australian monsoon season (EMEX9) and one of a midlatitude MCS observed during a 1985 field experiment in the central plains of the United States (PRE-STORM 23–24 June). For both simulations, the grid spacing on the finest grid is fine enough (1500 m for EMEX9, 2083 m for PRE-STORM) that no convective parameterization scheme is required.

A framework has been described for parameterizing the mesoscale updrafts and downdrafts of MCSs in mod-

els with resolution too coarse to resolve these drafts. The parameterization is analogous to the formulation of Donner (1993), with improvements of a more sophisticated convective driver (the Arakawa–Schubert convective scheme with convective downdrafts) and inclusion of the vertical distribution of various physical processes obtained through conditional sampling of two cloud-resolving MCS simulations. The mesoscale parameterization is tied to a version of the Arakawa–Schubert convective parameterization scheme that is modified to employ a prognostic closure, as described by Randall and Pan (1993). The parameterized Arakawa–Schubert cumulus convection provides condensed water, ice, and water vapor that drive the parameterization for the large-scale effects of mesoscale circulations associated with the convection.

The mesoscale thermodynamic parameterization is developed with the idea that determining thermodynamic forcing of the large scale depends on knowing the vertically integrated values and the vertical distributions of the following quantities: 1) deposition and condensation in mesoscale updrafts, 2) freezing in mesoscale updrafts, 3) sublimation and evaporation in mesoscale downdrafts, 4) melting in mesoscale downdrafts, and 5) mesoscale eddy fluxes of entropy and water vapor. The relative magnitudes of these quantities are constrained by assumptions made about the relationships between various quantities in an MCS's water budget deduced from three-dimensional cloud-resolving MCS simulations. The conditional sampling of the fine grid data of each MCS simulation attempts to identify mesoscale updrafts and mesoscale downdrafts within the stratiform region of each system. Once these mesoscale updraft and downdrafts are identified, vertical profiles of physical processes in these conditionally sampled mesoscale updrafts and downdrafts are used to determine the shapes of vertical profiles of various physical processes as well as relationships between various components of an MCS's water budget. The scheme is then tested by comparing the heating and drying tendencies produced by feeding it mean soundings from the simulations with tendencies diagnosed from the conditional sampling of the simulations.

A key issue that has not been addressed here is how the mesoscale parameterization scheme should be activated. Although the Arakawa–Schubert scheme can tell us whether deep convection is expected, it does not tell us whether it will be organized on the mesoscale or not. In practice, a simple way to activate the mesoscale component of the scheme could be to track the vertically integrated “mesoscale kinetic energy” (MKE). The term MKE has been used previously in the context of mesoscale circulations that develop in response to land surface heterogeneities (Avissar and Chen 1993). The MKE is an extension of the concept of cumulus kinetic energy (CKE) discussed by Lord and Arakawa (1980) and Randall and Pan (1993), among others. An equation for subgrid-scale MKE can be developed in a way anal-

ogous to Randall and Pan's equation for subgrid-scale CKE (as Avissar and Chen have done). The mesoscale parameterization could then either be turned on or off if the MKE exceeds a certain threshold or could have its tendencies modulated by the magnitude of the MKE. This is consistent with the concept that an MCS represents a more balanced form of convection (Olsson and Cotton 1997). That is, a signature of a balanced MCS would be the amount of kinetic energy residing in the mesoscale drafts. The MKE would have two fundamental sources/sinks. The first would be a source/sink from CKE and the second would be an internal source/sink due to the parameterized mesoscale circulation itself. Subsequent papers will describe the activation of the MCS parameterization in detail as well as the parameterization of momentum acceleration associated with MCSs.

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