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AN OBSERVATIONAL AND MODELING STUDY
OF MESOSCALE CONVECTIVE SYSTEM
ELECTRIFICATION

by Terry J. Schuur

Steven A. Rutledge, P.I.

**Colorado
State
University**

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by

Terry J. Schuur

Department of Atmospheric Science

Colorado State University

Fort Collins, CO 80523-1371

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ABSTRACT

AN OBSERVATIONAL AND MODELING STUDY OF MESOSCALE CONVECTIVE SYSTEM ELECTRIFICATION

The electrical structure and development of two Mesoscale Convective Systems (MCSs) observed during the 1991 Cooperative Oklahoma Profiler Studies (COPS91) experiment are analyzed through kinematic, microphysical, and electrical data sets. Profiles of the vertical electric field structure (from which vertical profiles of charge density are derived using an approximation to Gauss' Law) were obtained from a series of balloon-borne electric field meter (EFM) flights into each MCS. Two unique electric field structures were found. In both cases, the EFM data indicate that the MCS charge structure was characterized by horizontally extensive regions of charge and charge density magnitudes on the order of what is observed in convective cores ($\leq 5 \text{ nC/m}^3$). However, the data also indicate that the vertical electric field profiles were each related to unique MCS precipitation and kinematic structures, with a more complicated 5-layered charge profile (at $T \leq 0^\circ\text{C}$) associated with the "symmetric" MCS and a less complicated 3-layered charge profile (at $T \leq 0^\circ\text{C}$) associated with the "asymmetric" MCSs.

The observational analysis identified several kinematic, thermodynamic, and microphysical differences between the two systems that offer at least some explanation for their different electrical structures. First, ice particles detrained from the convective line of the symmetric MCS had much shorter "residence times" in the unfavorable growth/charging region associated

with the transition zone downdraft. The residence time of ice particles in the transition zone downdraft was much larger in the asymmetric case. Second, upon entering the MCS stratiform region, ice particles in the symmetric system entered a kinematic/microphysical environment that was more conducive to in situ charging. There are also indications that gravitational separation may be responsible for some of the charge transitions in the electric field profiles.

Model simulations of the symmetric MCS were conducted using a 2-D, time-dependent numerical model with bulk microphysical parameterizations. A number of charging mechanisms were considered, based on past and more recent laboratory studies. The simulations suggested that non-inductive ice-ice charge transfer in the low-liquid water content regime characteristic of MCS stratiform regions is sufficient to account for observed charge density magnitudes, and as much as 70% of the total stratiform charge (with the remaining 30% being the result of charge advection from the convective line). This result is consistent with the fraction of total stratiform precipitation generated by mesoscale ascent (in situ condensate production), thereby suggesting a linkage between stratiform electrification and stratiform precipitation processes. The model also indicates that, once these charge densities are achieved, the sink of charge resulting from fallspeed divergence becomes approximately equal to the rate of charge generation. This might lead to the quasi-steady layered structure that is commonly seen in the observations. Both non-inductive charging parameterizations (from Takahashi, 1978 and Saunders et al., 1991) reproduce some of the observed stratiform charge features. The best results, however, were obtained when charge advection and non-inductive charging processes were allowed to act in unison.

Terry J. Schuur
Department of Atmospheric Science
Colorado State University
Fort Collins, CO 80523
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TABLE OF CONTENTS

1. Introduction	1
1.1 Introduction and motivation for research	1
1.2 Background	3
1.2.1 Mesoscale Convective Systems	3
1.2.1.1 MCS dynamical structure.....	3
1.2.1.2 MCS microphysical structure.....	8
1.2.1.3 MCS electrical structure	10
1.2.2 Cloud electrification and charging theories	15
1.2.2.1 Overview.....	15
1.2.2.2 The non-inductive mechanism	18
1.2.2.3 Other charging mechanisms.....	19
1.3 Scientific objectives of this study.....	20
2. Data analysis	30
2.1 Data sources.....	30
2.1.1 Electric field meter data	30
2.1.2 Radar data.....	35
2.1.3 Microphysical data	37
2.1.4 Supplemental data.....	40
2.2 Coulomb's Law model	40
3. Kinematic, microphysical and electrical observations	53
3.1 The symmetric MCS of 24 May 1991	53
3.1.1 Environmental conditions and MCS development.....	53
3.1.2 MCS kinematic structure	55
3.1.2.1 P-3 flight track.....	55
3.1.2.2 P-3 Doppler observations of horizontal flow structure.....	55
3.1.2.3 P-3 Doppler composites of vertical motion	58
3.1.2.4 Ground-based Doppler observations	62
3.1.3 MCS microphysical structure.....	64
3.1.3.1 Optical array probe data	64
3.1.3.2 Cloud water	65
3.1.3.3 Particle trajectories	66
3.1.4 MCS electrical structure	67
3.1.4.1 EFM observations.....	67
3.1.4.2 Lightning observations	71
3.2 The asymmetric MCS of 8 May 1991	72
3.2.1 Environmental conditions and MCS development.....	72
3.2.2 MCS kinematic structure	74
3.2.2.1 P-3 flight track.....	74

3.2.2.2 P-3 Doppler observations of horizontal flow structure.....	75
3.2.2.3 P-3 Doppler composites of vertical motion	76
3.2.3 MCS microphysical structure.....	78
3.2.3.1 Optical array probe data	78
3.2.3.2 Cloud water	79
3.2.3.3 Particle trajectories	79
3.2.4 MCS electrical structure	80
3.2.4.1 EFM observations.....	80
3.2.4.2 Lightning observations	82
4. Comparison of symmetric and asymmetric mesoscale convective systems.....	116
4.1 Electrical structures	116
4.2 Kinematic and microphysical structures	119
4.3 Conceptual models.....	122
4.4 Comparison to other profiles	124
5. The numerical model.....	134
5.1 Model description	135
5.2 Model microphysics.....	137
5.2.1 Mixing ratios	138
5.2.2 Cloud ice concentrations	141
5.2.2.1 Prognostic equation for cloud ice concentration	141
5.2.2.2 Ice initiation.....	142
5.2.2.3 Ice multiplication	143
5.2.3 Other microphysical modifications	145
5.2.4 Cloud water	146
5.3 Model electrification	147
5.3.1 Continuity equations	147
5.3.2 Electric fields.....	149
5.3.3 Charge advection	150
5.3.4 Charging parameterizations.....	151
5.3.4.1 Non-inductive charging	151
5.3.4.2 Evaporation-condensation charging	153
5.3.4.3 Melting charging.....	155
5.4 Pulsing and data averaging.....	156
6. Numerical simulations.....	167
6.1 Kinematics and microphysics of the control run.....	168
6.2 Charge advection	173
6.3 Non-inductive charging.....	175
6.3.1 Takahashi (1978).....	175
6.3.1.1 Without charge advection.....	175
6.3.1.2 With charge advection.....	177
6.3.2 Saunders et al. (1991).....	179
6.3.2.1 Without charge advection.....	179
6.3.2.2 With charge advection.....	180
6.3.3 Charge budget analysis	182
6.4 Evaporation-condensation charging	184
6.5 Melting charging.....	186

7. Model sensitivity tests	219
7.1 Convective pulsing.....	219
7.2 Convective heating.....	221
7.3 Collections efficiencies.....	224
7.4 Charge advection	226
8. Summary and conclusions.....	259
8.1 Observations.....	259
8.2 Modeling	262
8.3 MCS structure and evolution.....	267
8.4 Conclusions and recommendations for future research	268
References	272
A. Microphysical parameterization	292
A.1 Mass weighted fallspeeds	292
A.2 Sources and sinks of water continuity variables	293
A.3 Source terms for the water continuity variables.....	297
B. List of symbols	299

LIST OF TABLES

Table 3.1	Charge analysis of the 0549 UTC, 24 May 1991 (symmetric) MCS electric field profile.	83
Table 3.2	Charge analysis of the 0705 UTC, 24 May 1991 (symmetric) MCS electric field profile.	83
Table 3.3	Charge analysis of the 0659 UTC, 8 May 1991 (asymmetric) MCS electric field profile.	83
Table 3.4	Charge analysis of the 0755 UTC, 8 May 1991 (asymmetric) MCS electric field profile.	83
Table 6.1	Takahashi (1978) non-inductive charging statistics.....	187
Table 6.2	Saunders et al. (1991) non-inductive charging statistics.....	187
Table 6.3	Comparison of Takahashi (1978) non-inductive charging with charge advection from the convective line.....	187
Table 6.4	Comparison of Saunders et al. (1991) non-inductive charging with charge advection from the convective line.	187

LIST OF FIGURES

Fig. 1.1 Conceptual model of mid-level horizontal cross-section through (a) a typical symmetric MCS, and (b) a typical asymmetric MCS. (From Houze et al., 1989).....	22
Fig. 1.2 Conceptual model of the kinematic and precipitation structure of a symmetric MCS as viewed in a vertical cross-section oriented normal to the convective line. (From Houze et al., 1989).....	23
Fig. 1.3 Conceptual model of the microphysical structure of a symmetric MCS as viewed in a vertical cross-section oriented normal to the convective line. (From Houze, 1989).....	24
Fig. 1.4 Conceptual model of the electrical structure of a symmetric MCS as viewed in a vertical cross-section oriented normal to the convective line. (From Schuur et al., 1991).....	25
Fig. 1.5 Schematic showing the tripole thunderstorm charge structure. (From Levin and Tzur, 1986).....	26
Fig. 1.6 Positive and negative rime charging zones as a function of temperature ($^{\circ}\text{C}$) and cloud water content (g m^{-3}) from the laboratory experiment of Takahashi (1978).	27
Fig. 1.7 Positive and negative rime charging zones as a function of temperature ($^{\circ}\text{C}$) and cloud water content (g m^{-3}) from the laboratory study of Jayaratne et al. (1983).....	28
Fig. 1.8 Positive and negative rime charging zones as a function of temperature ($^{\circ}\text{C}$) and effective liquid water content (g m^{-3}) from the laboratory experiment of Saunders et al. (1991).	29
Fig. 2.1 Locations of NSSL Norman (NRO) and Cimarron (CIM) Doppler radars and electric field meter (EFM) launch locations.....	44
Fig. 2.2 Electric field meter used during COPS91.	45
Fig. 2.3 Balloon-borne instrument train consisting of a 1200 g balloon, parachute, sonde, and electric field meter. (From Shepherd et al., 1996).....	46

Fig. 2.4 Electric field profile obtained from a balloon-borne electric field meter flight launched near the leading edge of the radar bright band of the 24 May 1991 (symmetric) MCS at 0549 UTC.	47
Fig. 2.5 (a) Basic geometry of fore-aft (FAST) scanning with respect to the P-3 aircraft. (b) Two-dimensional area swept out by FAST scanning during a quasi-linear flight leg.	48
Fig. 2.6 Coulomb's Law model test of EFM ascent through charge disks of (a) 2.5 km, (b) 5.0 km, (c) 10.0 km, (d) 20.0 km, (e) 40.0 km, and (f) $\gg 40$ km (infinite, determined theoretically) in width.	49
Fig. 2.7 Coulomb's Law model test of vertical and horizontal electric field components resulting from EFM ascent through a 10 km diameter disk of charge at locations (a) 1 km from the center, (b) 2 km from the center, (c) 3 km, (d) 4 km, (e) 5 km, and (f) 6 km (i.e., outside of the disk of charge) from the center of the disk.	50
Fig. 2.7 continued.	51
Fig. 2.8 Coulomb's Law model test of electric field profile depicted in Fig. 2.4 using (a) infinite (theoretically derived from Fig. 2.4 using Gauss' Law) (b) 10 km, (c) 8 km, (d) 6 km, (e) 4 km, and (f) 2 km wide disks of charge.	52
Fig. 3.1 Composite low-level radar reflectivity patterns of the 24 May 1991 (symmetric) MCS at (a) 0300, (b) 0400, (c) 0500, (d) 0600, (e) 0700, and (f) 0800 UTC.	84
Fig. 3.2 (a) NOAA P-3 aircraft flight track for 24 May 1991 (symmetric) MCS plotted with respect to the state of Oklahoma. (b) NOAA P-3 aircraft pressure altitude for 24 May 1991 (symmetric) MCS plotted with respect to time.	85
Fig. 3.3 Low-level radar reflectivity pattern (from CIM) at 0515 UTC, 24 May 1991 and CG lightning strike locations from 050730-052230 UTC, 24 May 1991.	86
Fig. 3.4 Dual-Doppler radar analysis at (a) 0.9, (b) 1.9, (c) 2.9, (d) 3.9, (e) 4.9, (f) 5.9, (g) 6.9, and (h) 7.9 km MSL.	87
Fig. 3.4 continued.	88
Fig. 3.5 Composite line-normal profile of (a) radar reflectivity (dBZ), and (b) vertical velocity (cm s^{-1}) constructed from the 0408 UTC, 24 May 1991 P-3 flight leg.	89
Fig. 3.6 Composite line-normal profile of (a) radar reflectivity (dBZ), and (b) vertical velocity (cm s^{-1}) constructed from all 24 May 1991 P-3 transition zone/stratiform region flight legs.	90
Fig. 3.7 Same as Fig. 3.5 except plotted with respect to the leading edge of the stratiform precipitation region.	91

Fig. 3.8	Same as Fig. 3.6 except plotted with respect to the leading edge of the stratiform precipitation region.	92
Fig. 3.9	Single-Doppler vertical cross-section of 24 May 1991 (symmetric) MCS (a) reflectivity (dBZ), and (b) system-relative radial velocity constructed along a 270° radial (i.e., normal to the convective line) at 0540 UTC, and (c) a conceptualization of the mesoscale vertical velocity profile.....	93
Fig. 3.10	NOAA P-3 shadow graph imagery obtained at a temperature of -8.5°C in 24 May 1991 (symmetric) MCS.....	94
Fig. 3.11	Schematic of ice particle trajectories in 24 May 1991 (symmetric) MCS overlaid on low-level (CIM) radar reflectivity at 0500 UTC.	95
Fig. 3.12	(a) Location of 0549 UTC 24 May 1991 electric field meter launch (indicated by arrow) overlaid on composite of low-level radar reflectivity patterns of the 24 May 1991 (symmetric) MCS at 0545 UTC. (b) Electric field profile from the 0549 UTC electric field meter flight. (c) Charge densities (nC m^{-3}) derived from the 0549 UTC electric field meter profile using Gauss' Law (Eqn. 2.3).	96
Fig. 3.13	Same as Fig. 3.12 except for 0705 UTC 24 May 1991 electric field meter launch/flight.....	97
Fig. 3.14	Same as Fig. 3.12 except for 0712 UTC 24 May 1991 electric field meter launch/flight.....	98
Fig. 3.15	Vertical cross-sections of 24 May 1991 (symmetric) MCS radar reflectivity for (a) 0549, (b) 0705, and (c) 0712 electric field meter launches.....	99
Fig. 3.16	Composite depiction of 0549 (solid line), 0705 (dashed line), and 0712 (dotted line) UTC 24 May 1991 (symmetric) MCS electric field meter profiles.	100
Fig. 3.17	Cloud-to-ground (CG) lightning flashes and bright band level radar reflectivity at 0512 UTC.....	101
Fig. 3.18	Cloud-to-ground (CG) lightning flashes and low-level radar reflectivity at (a) 0300, (b) 0400, (c) 0500, and (d) 0600 UTC 24 May 1991 (symmetric MCS).	102
Fig. 3.19	Composite low-level radar reflectivity patterns of the 8 May 1991 (asymmetric) MCS at (a) 0200, (b) 0300, (c) 0400, (d) 0500, (e) 0600, and (f) 0700 UTC.....	103
Fig. 3.20	(a) NOAA P-3 aircraft flight track for 8 May 1991 (asymmetric) MCS plotted with respect to the state of Oklahoma. (b) NOAA P-3 aircraft pressure altitude for 8 May 1991 (asymmetric) MCS plotted with respect to time.....	104
Fig. 3.21	Low-level radar reflectivity pattern (from CIM) at 0615 UTC, 8 May 1991 and CG lightning strike locations from 060730-062230 UTC, 8 May 1991.....	105

Fig. 3.22 Dual-Doppler radar analysis at (a) 0.9, (b) 1.9, (c) 2.9, (d) 3.9, (e) 4.9, (f) 5.9, (g) 6.9, and (h) 7.9 km MSL.....	106
Fig. 3.22 continued.....	107
Fig. 3.22 continued.....	108
Fig. 3.22 continued.....	109
Fig. 3.23 Composite line-normal profile of (a) radar reflectivity (dBZ), and (b) vertical velocity (cm s^{-1}) constructed from all 8 May 1991 P-3 transition zone/stratiform region flight legs.....	110
Fig. 3.24 NOAA P-3 shadow graph imagery obtained at a temperature of -8.5°C in 8 May 1991 (asymmetric) MCS.	111
Fig. 3.25 Schematic of ice particle trajectories in 8 May 1991 (asymmetric) MCS overlaid on low-level (CIM) radar reflectivity at 0615 UTC.	112
Fig. 3.26 (a) Location of 0659 UTC 8 May 1991 electric field meter launch (indicated by arrow) overlaid on composite of low-level radar reflectivity patterns of the 8 May 1991 (asymmetric) MCS at 0700 UTC. (b) Electric field profile from the 0659 UTC electric field meter flight. (c) Charge densities (nC m^{-3}) derived from the 0659 UTC electric field meter profile using Gauss' Law (Eqn. 2.3).....	113
Fig. 3.27 Same as Fig. 3.26 except for 0755 UTC 8 May 1991 electric field meter launch/flight.....	114
Fig. 3.28 Cloud-to-ground (CG) lightning flashes and low-level radar reflectivity at (a) 0500, (b) 0600, (c) 0700, and (d) 0800 UTC 8 May 1991 (asymmetric MCS).....	115
Fig. 4.1 Symmetric MCS electric field profiles at (a) 0549, and (b) 0705 UTC with 0, -12 , and -40°C temperature levels overlaid.....	126
Fig. 4.2 Mean composited vertical velocity of (a) transition zone, and (b) stratiform precipitation region for symmetric (solid line) and asymmetric (dashed line) MCSs, respectively.	127
Fig. 4.3 Charge layers from symmetric and asymmetric MCS overlaid on figure depicting depositional growth regime. (Adapted from Hallett, 1984).....	128
Fig. 4.4 Charge layers from symmetric MCS overlaid on figure depicting mean and maximum aggregate sizes. (Adapted from Hobbs et al., 1974, and Young, 1993).....	129
Fig. 4.5 Conceptual model depicting (a) kinematic, (b) microphysical, and (c) electrical structure of symmetric MCS.....	130

Fig. 4.6 Same as Fig. 4.6, except for asymmetric MCS.	131
Fig. 4.7 Electric field plotted with respect to temperature for the 18 June 1987 transition zone electric field profile. (From Schuur et al., 1991).....	132
Fig. 4.8 Vertical electric field profiles obtained in the MCS stratiform regions of the 4 June 1989 and 7 June 1989 MCSs (From Hunter et al., 1992).....	133
Fig. 5.1 Schematic of the model domain. (Adapted from Gallus, 1993).....	158
Fig. 5.2 Peak heating rate used to initialize model buffer zone.	159
Fig. 5.3 Peak hydrometeor mixing ratios used to initialize model buffer zone.	160
Fig. 5.4 Ice crystal fallspeeds. (from Kajikawa, 1972).....	161
Fig. 5.5 Collection efficiencies for thin plates collecting water droplets (from Pitter and Pruppacher, 1974).....	162
Fig. 5.6 Schematic showing grid sizes used in this modeling study.....	163
Fig. 5.7 Charge profiles used to initialize the buffer zone region. (a) Charge densities (nC m^{-3}) of snow and cloud ice. (b) Total charge density (nC m^{-3}).....	164
Fig. 5.8 Polynomial fit of charging current ($\text{A cm}^{-2} \times 10^{-16}$) versus temperature ($^{\circ}\text{C}$) from the data of Dong and Hallett (1992).....	165
Fig. 5.9 Pulsing amplitude used for convective line hydrometeor, heating, and electrical profiles.	166
Fig. 6.1 Control run radar reflectivities (dBZ) computed from precipitation-sized hydrometeor fields at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.	188
Fig. 6.2 Control run horizontal winds (m s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.....	189
Fig. 6.3 Same as Fig. 6.2 except model output is averaged over two convective pulses.....	190
Fig. 6.4 Control run vertical velocities (cm s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.....	191
Fig. 6.5 Same as Fig. 6.4 except model output is averaged over two convective pulses.....	192
Fig. 6.6 Control run snow mixing ratios (g kg^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.....	193
Fig. 6.7 Same as Fig. 6.6 except model output is averaged over two convective pulses.....	194

Fig. 6.8 Control run cloud water liquid water content (g m^{-3}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.	195
Fig. 6.9 Same as Fig. 6.8 except model output is averaged over two convective pulses.	196
Fig. 6.10 Control run cloud ice concentration (l^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.	197
Fig. 6.11 Control run average ice mass ($\text{kg} \times 10^{-11}$) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.	198
Fig. 6.12 Total charge density magnitudes (nC m^{-3}) for charge advection model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.	199
Fig. 6.13 Model snow/aggregate fallspeeds.	200
Fig. 6.14 Total charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.	201
Fig. 6.15 Same as Fig. 6.14 except model output is averaged over two convective pulses.	202
Fig. 6.16 Snow charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.	203
Fig. 6.17 Cloud ice charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.	204
Fig. 6.18 Same as Fig. 6.14 except with the addition of charge advection.	205
Fig. 6.19 Same as Fig. 6.18 except model output is averaged over two convective pulses.	206
Fig. 6.20 Grid averaged total charge density (nC m^{-3}) for the simulation depicted in Fig. 6.18 at 180 minutes into the simulation.	207
Fig. 6.21 Vertical charge density profile (nC m^{-3}) (a) from symmetric MCS observations (cf. Fig. 3.12), and (b) constructed though the model grid depicted in Fig. 6.18 at 180 minutes into the simulation and a location 35 km behind the convective line.	208
Fig. 6.22 Total charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.	209

Fig. 6.23	Same as Fig. 6.22 except model output is averaged over two convective pulses.....	210
Fig. 6.24	Snow charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.....	211
Fig. 6.25	Cloud ice charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.....	212
Fig. 6.26	Same as Fig. 6.22 except with the addition of charge advection.....	213
Fig. 6.27	Same as Fig. 6.26 except model output is averaged over two convective pulses.....	214
Fig. 6.28	Grid averaged total charge density (nC m^{-3}) for the simulation depicted in Fig. 6.26 at 180 minutes into the simulation.....	215
Fig. 6.29	Vertical charge density profile (nC m^{-3}) (a) from symmetric MCS observations (cf. Fig. 3.12), and (b) constructed though the model grid depicted in Fig. 6.18 at 180 minutes into the simulation and a location 35 km behind the convective line.	216
Fig. 6.30	Charge density magnitudes (nC m^{-3}) for the Dong and Hallett (1992) evaporation-condensation charging mechanism at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.....	217
Fig. 6.31	Charge density magnitudes (nC m^{-3}) for the Drake (1968) melting charging mechanism at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation.....	218
Fig. 7.1	Horizontal winds (m s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation.....	229
Fig. 7.2	Vertical velocities (cm s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation.....	230
Fig. 7.3	Snow mixing ratios (g kg^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation.....	231
Fig. 7.4	Cloud water liquid water content (g m^{-3}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation.....	232
Fig. 7.5	Cloud ice concentration (l^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation.....	233
Fig. 7.6	Total charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation.....	234

Fig. 7.7 Total charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation.....	235
Fig. 7.8 Horizontal winds (m s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation.....	236
Fig. 7.9 Vertical velocities (cm s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation.....	237
Fig. 7.10 Snow mixing ratios (g kg^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation.....	238
Fig. 7.11 Cloud water liquid water content (g m^{-3}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation.....	239
Fig. 7.12 Cloud ice concentration (l^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation.....	240
Fig. 7.13 Total charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation.	241
Fig. 7.14 Total charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation.	242
Fig. 7.15 Horizontal winds (m s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Peak convective heating used is $40^{\circ}\text{C h}^{-1}$	243
Fig. 7.16 Vertical velocities (cm s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Peak convective heating used is $40^{\circ}\text{C h}^{-1}$	244
Fig. 7.17 Horizontal winds (m s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies.....	245
Fig. 7.18 Vertical velocities (cm s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies.....	246
Fig. 7.19 Snow mixing ratios (g kg^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies.....	247

Fig. 7.20 Cloud water liquid water content (g m^{-3}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies.	248
Fig. 7.21 Cloud ice concentration (l^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies.	249
Fig. 7.22 Average ice mass ($\text{kg} \times 10^{-11}$) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies.	250
Fig. 7.23 Total charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies.	251
Fig. 7.24 Total charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies.	252
Fig. 7.25 Charge profiles used to initialize the buffer zone region for CA (a) Charge densities (nC m^{-3}) of snow (solid line) and cloud ice (dashed line). (b) Total charge density (nC m^{-3}).	253
Fig. 7.26 Total charge density magnitudes (nC m^{-3}) for advection of the charge profile depicted in Fig. 7.25 at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies.	254
Fig. 7.27 Same as 7.26 except with Takahashi (1978) non-inductive charging included.	255
Fig. 7.28 Vertical charge density profile (nC m^{-3}) (a) from symmetric MCS observations (cf. Fig. 3.12), and (b) constructed through the model grid depicted in Fig. 7.27 at 180 minutes into the simulation and a location 35 km behind the convective line.	256
Fig. 7.29 Same as 7.26 except with Saunders et al. (1991) non-inductive charging included.	257
Fig. 7.30 Vertical charge density profile (nC m^{-3}) (a) from symmetric MCS observations (cf. Fig. 3.12), and (b) constructed through the model grid depicted in Fig. 7.29 at 180 minutes into the simulation and a location 35 km behind the convective line.	258

CHAPTER 1

INTRODUCTION AND BACKGROUND

1.1 Introduction and motivation for research

Observational studies over the last few decades have shown that individual thunderstorms in both the midlatitudes and the tropics often are organized into mesoscale convective weather systems that can cover several thousand square kilometers (e.g. Newton, 1950; Fujita, 1955). These conglomerates, which are commonly referred to as Mesoscale Convective Systems (MCSs), often exhibit extensive regions of both convective and stratiform precipitation. Over the years, much has been learned of the precipitation and kinematic structures of MCSs. In particular, some studies have shown that the precipitation structure of MCSs can often be classified into distinct categories (e.g. Houze et al., 1989; Blanchard, 1990; Houze et al., 1990) with two of the more common types of mesoscale organization being: 1) a quasi-linear convective line trailed by a symmetric region of stratiform precipitation, and 2) a bowed convective line trailed by an asymmetric region of stratiform precipitation (with the stratiform precipitation region commonly associated with a mesoscale vortex at the northern end of the MCS). A conceptual model depicting the typical precipitation and kinematic structures of these two types of systems is shown in Fig. 1.1 (from Houze et al., 1989).

By comparison, the electrical structure and evolution of MCSs is less understood. Initially it was noted that MCSs have a tendency to develop “bipolar” cloud-to-ground (CG) lightning patterns wherein negative CG flashes are predominantly located in the MCS’s convective region and positive CG flashes in the trailing stratiform precipitation (e.g., Orville et al., 1988; Rutledge

and MacGorman, 1988). More recently, several studies have documented the electrical structure of MCS stratiform clouds using balloon-borne electric field meter data (e.g., Schuur et al., 1991; Hunter et al., 1992; Marshall and Rust, 1993; Stolzenburg et al., 1994; Stolzenburg, 1996). These studies have consistently shown that electric fields and derived charge densities are of the same order of magnitude as those observed in convective cores (i.e., $>100 \text{ kV m}^{-1}$ and $>3 \text{ nC m}^{-3}$, respectively). Furthermore, Marshall and Rust (1993), who analyzed electric field data collected in several different MCSs, noted that the two distinct MCS types listed above (i.e., symmetric and asymmetric) exhibited distinct electrical structures. Symmetric type MCSs were found to have a complicated 5-layered charge structure (at $T \leq 0^\circ\text{C}$) while asymmetric type MCSs were found to have a simpler 3-layered structure (at $T \leq 0^\circ\text{C}$).

Combined, these recent discoveries of bipolar lightning patterns, large stratiform region electric fields and charge densities, and repeatable electrical structures (which appear to be related to MCS type), raise many questions regarding the processes that lead to MCS electrification. In this study we address the concurrent kinematic, microphysical, and electrical structures of each MCS type (symmetric and asymmetric, both observed during the 1991 Cooperative Oklahoma Profiler Studies experiment) in an attempt to better understand the kinematic and microphysical differences that might account for the different electrical structures. No previous observational study of MCS electrification has combined such comprehensive data sets. Additionally, we investigate the electrification of a symmetric type MCS with a two-dimensional, bulk-microphysical-electrical model. In this chapter, we first review studies of MCS dynamical, microphysical, and electrical structures. Then, in order to provide a framework from which the numerical modeling might be better understood, we review thunderstorm charging theories and discuss their relevance to the present work. Finally, we outline the goals of this thesis.

1.2 Background

1.2.1 Mesoscale Convective Systems

In the observational component of this study (Chapters 2, 3 and 4), we develop conceptual models depicting the combined kinematic, microphysical, and electrical structures of two distinct types of MCS: symmetric and asymmetric. To provide the necessary background for understanding the observational analyses, we now review recent work on MCS dynamical, microphysical, and electrical structure and evolution.

1.2.1.1 MCS dynamical structure

Though observational studies have identified several different types of mesoscale organization (e.g., Houze et al., 1990), this review focuses on the simplest and most widely studied type of MCS: the symmetric (Fig. 1.1a), leading line/trailing stratiform system (e.g., Houze, 1977; Zipser, 1977; Ogura and Liou, 1980; Gamache and Houze, 1982, 1983, 1985; Smull and Houze, 1985, 1987a; Rutledge et al., 1988; Biggerstaff et al., 1991a; Schuur et al., 1991, and many others). Recently, the asymmetric (Fig. 1.1b) type MCS, which is occasionally associated with a mesoscale swath of damaging winds and referred to as a bow-echo, has also received much attention (e.g., Johnson and Bartels, 1992; Duke and Rogash, 1992; Zhang, 1992; Brandes and Ziegler, 1993; Jorgensen and Smull, 1993; Weisman, 1993; Skamarock et al., 1994; Scott and Rutledge, 1995).

As can be seen by comparing Figs. 1.1a and 1.1b, the primary difference between the two types of systems lies in their dimensionality. Symmetric type MCSs (Fig. 1.1a) are rather two-dimensional since they typically exhibit quasi-linear convective and stratiform precipitation structures that vary little from one end of the system to the next. Additionally, though a line-parallel component of airflow commonly exists in symmetric type systems, it is generally rather

weak when compared to the line-normal component of airflow. Asymmetric type MCSs (Fig. 1.1b), on the other hand, are highly three-dimensional since their convective line often exhibits a “bowed” structure and the bulk of their stratiform precipitation is commonly located towards the northern end of the system. The line-parallel component of airflow in asymmetric type systems is typically rather strong and a mesoscale vortex is often found at the northern end of the system. While MCSs are commonly classified as being either one “type” or the other, some studies have shown that symmetric type MCSs occasionally evolve into asymmetric MCSs in their dissipating stages (e.g., Rutledge et al., 1988; Houze et al., 1989; Loehrer and Johnson, 1995). A more comprehensive review of MCS precipitation and kinematic structure can be found in Rutledge (1991), Houze (1993), and Smull (1995).

Typical MCS line-normal precipitation and kinematic structures are depicted in Fig. 1.2 (from Houze et al., 1989; oriented along line AB in Fig. 1.1). The MCS is commonly characterized by three distinct precipitation regions: the convective line, the transition zone, and the stratiform precipitation region. In general, the kinematic structure is dominated by two airflow regimes: a front-to-rear directed flow that ascends rapidly in the convective line, bridges the transition zone at the mid-levels and protrudes into the trailing stratiform region, and a rear inflow jet (e.g., Smull and Houze, 1987b) that enters the system’s trailing stratiform boundary and slopes downward towards the convective line, often contributing to the low-level cold pool behind the leading gust front (e.g., LaFore and Moncrieff, 1989). Occasionally a shallow region of front-to-rear flow, the result of low-level divergence of the convective and transition zone downdrafts, is observed immediately above the surface in the transition zone and stratiform precipitation regions.

Since the microphysical, and electrical data sets presented in this study are likely representative of contributions from each of the three aforementioned MCS precipitation regions,

we now examine in detail the precipitation and kinematic structures of each region individually. The first MCS precipitation region, the convective line, has been shown to exert many influences on the trailing stratiform region's structure and development through the advection of hydrometeors (Gamache and Houze, 1982; Smull and Houze, 1985; Rutledge and Houze, 1987; Houze, 1989; Gallus and Johnson, 1995a), convective heating (Gamache and Houze, 1985; Gallus and Johnson, 1991; Gallus and Johnson, 1995b), and momentum (Moncrieff, 1981; LeMone, 1983; Smull and Houze, 1987a; Gallus and Johnson, 1992). The convective line, in turn, typically consists of a series of cells at different stages of development. The first radar echo in this sequence often corresponds to a developing convective cell located immediately above the gust front. This is commonly followed by a mature convective cell that is associated with the MCS's heaviest precipitation rates, strongest vertical motions, and highest radar reflectivities, and a series of weaker, dissipating cells (not shown in Fig. 1.2), which are occasionally observed at increasingly higher levels as their remnants are detrained rearward into the stratiform region by the ascending front-to-rear flow regime (Houze et al., 1989). Sun et al. (1993) used a thermodynamic retrieval to show that these detrained convective cells were warmer than their environment and, therefore, positively buoyant.

Trailing the convective line is the transition zone (e.g., Ligda, 1956; Sommeria and Testud, 1984; Smull and Houze, 1985), which is an approximately 30-50 km wide region of weak radar reflectivity and light precipitation that precedes the heavier stratiform precipitation. Several studies have shown that the transition zone is a region of deep mesoscale descent (e.g., Houze and Rappaport, 1984; Smull and Houze, 1985; Heymsfield and Schotz, 1985; Srivastava et al., 1986; Roux, 1988; Rutledge et al., 1988; Biggerstaff et al., 1993; Cifelli and Rutledge, 1994, and others). Several hypothesis have been advanced to explain the transition zone. Ligda (1956), who referred to the transition zone region as a "precipitation-free zone" or the "dry slot",

suggested that hydrometeors that fell into the transition zone either evaporated quickly or were caught up in some circulation that prevented them from falling out in the transition zone region. Smull and Houze (1985, 1987a) later suggested that either 1) enhanced subsidence leading to sublimation and evaporation, or 2) fallspeed sorting, whereby heavier hydrometeors fall out close to the convective line and lighter ice particles are advected significantly further rearward, were responsible for the transition zone reflectivity deficit. The fallspeed sorting hypothesis was tested by Rutledge and Houze (1987) who varied graupel fallspeeds in a numerical model and determined that fallspeed sorting can explain at least a portion of the transition zone reflectivity deficit.

More recently, in a detailed study that examined the source of the observed transition zone subsidence, Biggerstaff and Houze (1993) concluded that the transition zone downdraft often consists of two distinct parts: an upper level downdraft that appeared to be mechanically forced (partially by gravity waves excited by penetrative convective updrafts) and a lower level downdraft that appeared to be forced by microphysical processes (sublimation, evaporation, and melting). They further suggested that the weaker reflectivity observed in the transition zone might be the result of the inability of ice particles to form and grow in the deep transition zone downdraft, thereby resulting in less aggregation (and weaker reflectivities) than is commonly observed to its rear in the heavier stratiform precipitation. In a retrieval study, Braun and Houze (1994) concluded that transition zone reflectivity minimum was largely the result of the suppression of hydrometeor growth (rather than sublimation) in the transition zone downdraft.

Following the transition zone is the stratiform precipitation region, typically identified by a horizontal band of locally higher reflectivities (located immediately below the 0°C isotherm) known as the radar “bright band” (e.g., Battan, 1973). The radar bright band, in turn, generally delineates the location of the most intense stratiform precipitation. Unlike the transition zone,

which is characterized by a region of deep mesoscale descent, the vertical motion structure of the stratiform precipitation region is characterized by a broad region of mesoscale ascent overlying a deep, unsaturated mesoscale downdraft (e.g., Ogura and Liou, 1980; Gamache and Houze, 1982; Smull and Houze, 1985; Srivastava et al., 1986; Rutledge et al., 1988; Biggerstaff and Houze, 1991a; Cifelli and Rutledge, 1994, and many others). These studies have also indicated that the magnitudes of these mesoscale updrafts and downdrafts are generally quite weak, usually less than 1 m s^{-1} , with the sloping transition level between updraft and downdraft occurring at a height of 0-2 km above the radar bright band (Houze, 1989).

The source and location of the radar bright band has been the topic of much research. Smull and Houze (1985) hypothesized that the position and intensity of the radar bright band was related to the aggregation, and melting of the preferred fallout location of a mesoscale swath of ice particles advected from the convective line. In a numerically modeling study, Rutledge and Houze (1987) determined that both ice particle advection and the depositional growth of those particles in the mesoscale updraft was necessary to explain stratiform precipitation rates. They referred to this two part process as the MCS feeder-seeder mechanism. More recently, Matejka and Schuur (1991) composited Doppler radar data in the stratiform region of a symmetric type MCS and found that a local maximum in the stratiform region mesoscale updraft occurred near the leading edge of the heaviest stratiform precipitation. By examining trajectories of ice particles that passed through this updraft maximum, they concluded that enhanced depositional growth of ice particles passing through this region explained the location and intensity of the stratiform bright band in the case they studies rather well. Enhanced depositional growth in this region is also implied by Tao and Simpson (1989), whose modeling work indicated that the majority of depositional growth occurred within 40-50 km of the convective line.

1.2.1.2 MCS microphysical structure

Observational studies of the microphysical structure of MCSs are less common compared to studies of their kinematic structure. A cross-section depicting the typical microphysical structure of a symmetric type MCS is shown in Fig. 1.3 (from Houze, 1989; oriented along a line analogous to that of Fig. 1.2). In this review, we focus on the microphysical structure and development of the MCS stratiform region. The mid-level microphysical structure of a midlatitude MCS convective line has been documented by Heymsfield and Hjelmfelt (1984) and of a tropical MCS line by Houze and Churchill (1984) and Gamache (1990).

As can be seen from Fig. 1.3 (and comparing it with the precipitation and kinematic structures described above in Section 1.2.1.1), ice generated in convective line updrafts is advected rearward in the MCS's front-to-rear flow, bridging the transition zone and distributed into the mesoscale updraft. Ice particles continue to grow by vapor deposition in the mesoscale updraft in a seeder-feeder type process (e.g., Rutledge and Houze, 1987). However, the evolution of these hydrometeors as they pass through the weaker reflectivity, transition zone region is less well known. As discussed in the previous section, it is widely accepted that deep subsidence in the transition zone results in 1) sublimation (e.g., Smull and Houze, 1985), or 2) suppressed growth (e.g., Biggerstaff and Houze, 1993; Braun and Houze, 1994) of hydrometeors. The in situ aircraft observations of Yeh et al. (1991) indicate that hydrometeor concentrations are as much as 4 to 6 times higher in the transition zone than in the stratiform region and that the mean size of the hydrometeors in the transition zone are smaller by as much as a factor of two. This difference in mean ice size would likely explain the smaller reflectivities observed in the transition zone region, though the actual mechanism responsible for these smaller ice sizes remains unknown.

In addition to depositional growth, ice particles in the stratiform region also grow by aggregation (Houze and Churchill, 1984; Yeh et al., 1991). Evidence has also been presented to support riming in the stratiform region (Leary and Houze, 1979; Yeh et al., 1991). In particular, aggregation has been noted to be significant in the 0 to -12°C layer where dendrites interlock to form larger snowflakes and continued aggregation takes place as the ice particles become coated with a pseudo-film of water, thereby enhancing their collection efficiency and producing large aggregates. The largest snowflakes are typically observed at temperatures just below freezing (Young, 1993). In a study that focused on the microphysical evolution above, through, and below the melting layer, Willis and Heymsfield (1989) noted that, despite the simultaneous reduction in aggregate concentrations at warmer temperatures, the concentrations of small ice particles increased to concentrations of $>100 \text{ L}^{-1}$ at -2°C. Since this level is commonly within the mesoscale downdraft (e.g., Houze, 1989), water saturated conditions are often suppressed and ice multiplication due to the Hallett-Mossop rime splintering mechanism (Hallett and Mossop, 1974) was considered an unlikely source of ice multiplication. Rather, Willis and Heymsfield speculated that the higher ice concentrations were the result of mechanical fracturing (e.g., Hobbs and Farber, 1972; Vardiman, 1978; Dong et al., 1994). As the aggregates continue to fall through the 0°C isotherm, they begin to melt and acquire an outer layer of liquid water that greatly increases the radar reflectivity (e.g., Battan, 1973). Below this "bright band" region, radar reflectivity decreases as continued melting results in smaller sizes, eventual drop breakup, evaporation, and fallspeed divergence. Rutledge and MacGorman (1988) speculated that the aggregation process may play a role in subsequent charge separation by a non-inductive transfer of charge during ice/ice collisions (addressed in more detail later) through its ability to produce larger ice particles.

Riming growth in MCS stratiform regions remains more controversial due to the question of whether MCS mesoscale updrafts are strong enough to produce and sustain sufficient cloud water. However, recent aircraft observations (e.g., Yeh et al., 1991, and reported in this study) indicate that cloud water contents in mesoscale updrafts are frequently in the $0.2\text{--}0.3\text{ g m}^{-3}$ range; localized pockets of cloud water as high as 0.5 g m^{-3} are also occasionally observed. It is likely that much of this cloud water is generated in regions of higher updraft velocity that are commonly embedded within the mean mesoscale updraft (Rutledge and MacGorman, 1988). Other studies have observed either lightly rimed hydrometeors in the stratiform region (e.g., Yeh et al., 1991) or inferred the presence of riming (and small graupel contents) above the radar bright band by examining drop size distributions at warmer temperatures (e.g., Leary and Houze, 1979). Small graupel concentrations (0.07 L^{-1}) were also observed by Yeh et al. (1991). In addition, numerical modeling studies have either diagnosed (Rutledge, 1986; Rutledge and Houze, 1987) or predicted (Gallus and Johnson, 1995a) the presence of small amounts of cloud water in the MCS stratiform regions.

1.2.1.3 MCS electrical structure

While the electrical structure of individual thunderstorms have been investigated for many years, studies of midlatitude MCS electrical properties is a comparatively new area of research. In somewhat related work, Imyanotov et al. (1972) found electric fields of up to 40 kV m^{-1} in nimbostratus clouds in the Soviet Union. It isn't clear that the nimbostratus clouds studied by Imyanotov et al., however, were as dynamically active as the MCSs commonly observed over the great plains of the United States. Chauzy et al. (1980) and Chauzy et al. (1985) launched balloon-borne electric field meters into MCS over southwestern France and West Africa, respectively. Their stratiform region electric field meter data, all of which ended at approximately 6 km of altitude, indicated electric fields as high as 60 kV m^{-1} and a negative

region of charge near cloud base. Chauzy et al. (1985) further observed that ground-level electrostatic field changes (associated with lightning discharges) occurred both in the transition zone region and the stratiform precipitation region to its rear.

In the United States, the advent of magnetic direction-finder lightning detection networks (e.g., Orville et al., 1983; Mach et al., 1986) in the early 1980's spawned interest in MCS electrification by providing detailed information on the temporal and spatial variability of lightning within MCSs. Rust et al. (1985) and Goodman and MacGorman (1986) studied CG flash rates in Mesoscale Convective Complexes (a subset of MCSs, Maddox, 1980) and found peak flash rates of 48 and 60 min⁻¹, respectively. In studies that examined the spatial distribution of MCS lightning flashes, Orville et al. (1988) and Rutledge and MacGorman (1988) observed that MCSs tend to develop "bipolar" lightning patterns, wherein negative polarity CG flashes (i.e., those that lower negative charge to ground) occurred primarily in the MCS convective region and the bulk of the positive CG flashes (i.e., those that lower positive charge to ground) were situated in the trailing stratiform precipitation. Rutledge and MacGorman suggested that the advection of positive charge on small ice particles from the upper levels of the convective line might be responsible for the production of positive CG flashes in the MCS stratiform region. However, they also speculated that in situ charge generation by a non-inductive transfer of charge during ice/ice collisions (discussed later) represented another possible charge source. Other characteristics of CG lightning in MCSs have been investigated by Petersen and Rutledge (1992) who, in a study of tropical MCSs over northern Australia, found that positive peak currents of CG flashes in the stratiform region were much larger than those from the convective line. They argued that this indicated that the CG flashes in MCS stratiform regions were likely tapping a large reservoir of charge, afforded by horizontally extensive stratiform regions.

In a study that used lightning and surface-based electric field data to document the electrical structure of several winter storms and two warm season MCSs, Engholm et al. (1990) concluded that the lightning bipoles observed in their study were largely attributable to active charge separation over the system's entire area, not just in the deep convection associated with the convective line. To test the in situ charge generation hypothesis, Rutledge et al. (1990) developed a one-dimensional numerical model to examine the non-inductive transfer of charge resulting from ice/ice collisions in a low liquid water environment. They found that their model produced an "inverted dipole" structure (i.e., negative charge over positive charge) and charge density magnitudes of 3 nC m^{-3} . Based on this finding, they speculated that in situ charge generation might account for both the positive and negative CG flashes that are commonly seen in MCS stratiform regions. The charge advection hypothesis, on the other hand, was investigated in more detail by Rutledge et al. (1993) who employed a simple plume model to examine the expansion of advected charge by both turbulent diffusion and Coulomb forces, concluding that the charge advection mechanism did not offer a plausible explanation for lightning bipoles. Rutledge and Petersen (1994) presented calculations from a simple one-dimensional model that diagnosed the presence of cloud water at temperatures between 0 and -20°C . These model results, combined with radar and lightning observations of several MCSs led them to conclude than in situ charging was likely occurring in MCS stratiform regions.

Several studies have investigated the electrical structure of MCS stratiform clouds by examining balloon-borne electric field meter (EFM) data. Schuur et al. (1991) presented an electric field profile obtained from a balloon-borne EFM launched into the transition zone of a symmetric type MCS. Their data showed a complicated stratiform charge structure consisting of several layers of charge, electric field magnitudes that exceeded 100 kV m^{-1} , and charge densities (derived using an approximation of Gauss's Law) of nearly 4 nC m^{-3} . A conceptual model

depicting the charge structure derived from that EFM sounding, overlaid on the concomitant MCS precipitation and kinematic structure (oriented along a line analogous to that of Figs. 1.2 and 1.3) is shown in Fig. 1.4. Schuur et al. further examined their data with respect to the microphysical and electrical models of Rutledge and Houze (1987) and Rutledge (1990), respectively, and concluded that two layers of charge observed in their EFM sounding (negative above positive at 6.5 km) were representative of the inverted dipole structure modeled by Rutledge et al. (1990). They also found, however, that at least some of the charge in their profile could only be explained by charge advection from the convective line.

Several studies have since examined multiple electric field meter profiles in MCSs. Hunter et al. (1992) compared electric field profiles collected in two separate MCSs and found that the profiles exhibited essentially the same vertical charge structure. Marshall and Rust (1993) compared charge profiles derived from multiple EFM flights into several MCSs and found that two distinct MCS charge structures exist: 1) a rather complicated 5-layered charge structure characteristic of symmetric type MCSs, and 2) a simpler 4-layered charge structure characteristic of asymmetric type MCSs. They referred to these two types of electric field profiles as Type A and Type B, respectively, and remarked that MCS stratiform charge structure was much more repeatable than is commonly observed in individual convective cells. They speculated that the underlying cause for this repeatability might be related to MCS flow structure and microphysics. Bateman et al. (1995) presented particle size and charge data that were collected on several of the stratiform EFM flights presented by Marshall and Rust (1993). Interestingly, their data indicated that many locations contained nearly equal amounts of positive and negative charge rather than a single, dominant polarity of charge. Furthermore, by comparing their data to the concurrent EFM measurements, they concluded that the bulk of the

charge in MCS stratiform regions is carried on hydrometeors smaller than 0.8 mm, which was the minimum detectable size of their instrument.

In a study of multiple electric field meter profiles obtained in a single MCS, Stolzenburg et al. (1994) concluded that charge regions in an MCS stratiform cloud are horizontally extensive. The charge transition found near the 0°C isotherm, which is a feature that appears to be common to nearly all MCS EFM soundings (e.g., Marshall and Rust, 1993), was likely the result of one or more in situ charge generation mechanisms. They also examined the turbulent diffusion of charge advected from a line-source and concluded, in disagreement with the work of Rutledge et al. (1993), that advection of charge from the convective line is indeed a viable charging mechanism over distances characteristic of MCS stratiform regions. More recently, Stolzenburg (1996) composited several EFM soundings, which were classified as updraft or non-updraft soundings based on their rate of ascent, to examine the characteristic electrical structure of several MCS precipitation regions (but focusing on the convective line). From the composited soundings, Stolzenburg concluded that characteristic charge structures exist in both the updraft and non-updraft regions of MCS convective lines. Combining these composited soundings with others obtained in the trailing MCS's stratiform region, Stolzenburg developed a conceptual model that argued for the role of charge advection in leading to stratiform region charge structures.

Shepherd et al. (1996) examined electric field profiles from several MCS stratiform regions, focusing on the electric field structure through the 0°C isotherm, isothermal layer (e.g., Willis and Heymsfield, 1989), and radar bright band. Sharp charge transitions near the freezing level have been noted previously by Marshall and Rust (1993) and Stolzenburg et al. (1994). From examining 7 stratiform cases and 5 non-stratiform cases, Shepherd et al. concluded that melting

charging (e.g., Simpson, 1909; Drake, 1968) was the primary cause of the high charge densities found at or below the freezing level in electrified clouds.

1.2.2 Cloud electrification and charging theories

In the numerical modeling component of this study (Chapters 5, 6, and 7), we address a wide number of potential electrification mechanisms, ranging from the advection of charge from the convective line to a variety of in situ charge generation processes. In order to provide a framework from which the modeling work might be better understood, we now review thunderstorm electrification hypotheses.

1.2.2.1 Overview

Cloud electrification is a topic that has intrigued scientists for many centuries (see Uman, 1987 for a review) but it is only in the past 60 years that high quality in situ electrical measurements of convective clouds have been possible (e.g., Simpson and Scrase, 1937). From those early measurements, a conceptual model of thunderstorm charge structure has evolved in which a primary region of negative charge is located beneath an upper region of positive charge (with the centers of charge located at temperatures of approximately -15°C and -40°C , respectively). In addition, a weak region of positive charge is often found near cloud base. This charge structure, commonly referred to as the tripole conceptual model (see Fig. 1.5), is still widely accepted (Williams, 1989). More recent measurements (e.g., Marshall and Rust, 1991), however, have indicated that thunderstorm charge structure may often be much more complicated. These newfound variations in convective cloud charge structure are perplexing and further complicate our ability to understand the basic processes leading to the electrification of thunderstorms and stratiform regions of MCSs. Here we provide a background of cloud

electrification research to provide a framework from which the modeling component of this study might be better understood.

In general, the proposed mechanisms by which thunderclouds develop regions of net charge can broadly be grouped into three categories: 1) convective charging, 2) inductive charging, and 3) non-inductive charging (a more complete review of these and other mechanisms can be found in Beard and Ochs, 1986). The convective mechanism (e.g., Grenet, 1947; Vonnegut, 1953, Wagner and Telford, 1981) relies upon the convective transport of ions within the cloud. With this mechanism, it is suggested that negative ions drifting towards the surface as part of the Wilson current (Wilson, 1920) become attached to the upper cloud boundary, thereby greatly reducing their mobility and creating an upper region of negative (screening layer) charge. This negative charge is then transported downwards by convective downdrafts and overturning turrets at the cloud boundaries. At the same time, positive charge produced by corona emission at the Earth's surface (in response to the building negative charge in the lower portion of the cloud) is attracted to the lower cloud boundary and carried upward by convective updrafts. This newly formed upper positive charge center, in turn, attracts even more negative ions to the upper cloud boundary. Hence a positive feedback loop is established. Several recent studies, however, have shown that the convective theory fails to explain many observed thunderstorm charge and lightning features (e.g., Williams, 1989). Because of these deficiencies, and also due to computational constraints (addressed later), we neglect atmospheric ions in the numerical modeling component of this study. Screening layer formation (e.g., Brown et al., 1971), however, is crudely parameterized using an approach similar to that presented by Ziegler et al. (1991).

Unlike the convective mechanism, the inductive (e.g., Elster and Geitel, 1913; Illingworth and Latham, 1977; Mason, 1988) and non-inductive (e.g., Reynolds et al., 1957; Takahashi,

1978; Saunders et al., 1991) mechanisms require the presence and interaction of precipitation-sized hydrometeors. With the inductive mechanism, it is suggested that a downward-directed electric field polarizes hydrometeors such that their upper (lower) surfaces are negatively (positively) charged. Rebounding collisions between hydrometeors should then result in a transfer of positive (negative) charge to the upward (downward) moving particle, thereby resulting in a strengthening of the existing electric field as the particles gravitationally separate. It should be noted that the effectiveness of the inductive mechanism is highly dependent on the strength of the electric field. That is, as the electric field increases, the degree of hydrometeor polarization and the resultant transfer of charge also increase. For this reason, the inductive mechanism fails to adequately explain the initial electrification of clouds since electric fields at that time are usually quite low (where the fair weather field is approximately -100 V m^{-1}). Nevertheless, the inductive mechanism has been shown to be an efficient charge transfer mechanism once high electric fields are established (e.g., Kuettner et al., 1981). For this reason, the inductive charging mechanism is investigated in the numerical modeling component of this study, though only as a sensitivity test after other processes, namely non-inductive charging, have already established high (i.e., $>10 \text{ kV m}^{-1}$) electric fields.

Of the three primary charging mechanisms listed above, the non-inductive mechanism is widely accepted as the most viable candidate to explain thunderstorm electrification (Latham, 1981; Illingworth, 1985; Williams, 1989). Over the years, numerous laboratory studies (e.g., Reynolds et al., 1957; Takahashi, 1978; Jayaratne et al., 1983; Keith and Saunders, 1990; Saunders et al., 1991, and many others) have consistently demonstrated that rebounding graupel/ice collisions in the presence of supercooled cloud water droplets results in a small amount of charge transfer (1-100 fC), where the polarity of charge transferred to the graupel (and opposite to the cloud ice) is related to the temperature and liquid water content (LWC) of

the cloud. Though the process by which clouds electrify is far from being a resolved issue, numerous observational studies of convective clouds have offered support for the non-inductive mechanism (Williams, 1989). Nevertheless, despite all of the supporting evidence for the non-inductive mechanism, much debate remains on 1) the temperature/LWC parameter space that defines the polarity of charge transferred, and 2) the physical mechanism of charge transfer. Since the non-inductive mechanism is extensively tested in the numerical modeling component of this study, it is reviewed here in more detail than the convective and inductive mechanisms.

1.2.2.2 The non-inductive mechanism

The non-inductive charging mechanism was first investigated by Reynolds et al. (1957) who found that a riming graupel pellet (simulated by rotating an ice sphere through a laboratory cloud of ice crystals and supercooled drops) charged negatively at temperatures around -25°C . Church (1966) conducted a similar experiment and found that the polarity of charge separated was a function of both temperature and LWC, with the graupel charging negatively at -15°C in low LWC, but positively in high LWC. Takahashi (1978) extended this work by determining charge transfer over a wide range of temperatures and LWCs. The data from his study (see Fig. 1.6) indicated that graupel charges positively at all temperatures warmer than -10°C . At colder temperatures, however, the graupel charged positively for both low and high LWCs, but negatively at intermediate LWCs.

Since those early studies, the majority of laboratory work has been conducted at the University of Manchester and has produced results that, at times, differed substantially from those of the earlier studies. Jayaratne et al. (1983) investigated the charge transfer to graupel by moving a riming target rod through a chamber of ice crystals at speeds of 1.5 to 3 m s^{-1} , which results in a much slower and realistic impact velocity than used in the earlier work. They found

(see Fig. 1.7) that the simulated graupel charged positively at temperatures warmer than a “reversal temperature” and negatively at temperatures below it. The reversal temperature, in turn, was related to the LWC. A comparison (cf. Figs. 1.6 and 1.7) of the results of Jayaratne et al. (1983) with those of Takahashi (1978) shows marked disagreement in the polarity of charge that is transferred at low and intermediate LWCs and temperatures between -10 and -20°C. More recent studies (e.g., Jayaratne and Saunders, 1985; Keith and Saunders, 1990; Saunders et al., 1991) have determined that the magnitude and polarity of charge transferred during graupel/ice collisions is dependent on a number of other factors, including 1) the effective LWC, which they defined as the proportion of the droplet spectrum captured by the target, 2) the ice crystal size, and 3) the collision impact velocity. Data from the Saunders et al. (1991), which quantified polarity (and magnitude as determined by a set of corresponding equations) of charge transfer with respect to the effective LWC, is shown in Fig. 1.8. It is clear that much uncertainty remains regarding the magnitude and polarity of charge transferred during ice/ice collisions. Several recent laboratory (e.g., Baker et al., 1987) and theoretical (e.g., Williams et al., 1991) studies have suggested that the vapor growth state on the surface of the colliding ice particles is responsible for determining the polarity of charge transfer. A more detailed review of non-inductive charging and proposed mechanisms for the actual transfer of charge may be found in Saunders (1994).

1.2.2.3 Other charging mechanisms

Numerous other proposed charging mechanisms do not clearly fall into the convective, inductive, or non-inductive categories. Since some of these mechanisms, however, may be applicable to MCS stratiform region electrification (and are therefore included in the modeling component of this study), we now present a brief review of some of these mechanisms. One class of microphysical charging theories suggests that droplets undergoing condensational growth

(e.g., Takahashi, 1973) or ice particles undergoing depositional growth (e.g., Findeisen, 1940; Rydock and Williams, 1991; Dong and Hallett, 1992) obtain a small amount of charge. Though the actual process by which charge is transferred by this mechanism is the subject of debate, the various studies have conclusively shown that ice particles in a state of deposition charge positively and those in a state of sublimation charge negatively (Saunders, 1994). Recently, Dong and Hallett (1992) conducted a laboratory study that quantified the magnitude of this charge transfer by this mechanism and concluded that charging current is typically on the order of 10^{-16} amps cm^{-2} s^{-1} . However, Dong and Hallett also found that the magnitude of charge transferred was highly dependent on 1) whether earlier riming had taken place, and 2) the ion concentration of the cloud, which is very uncertain.

Another class of microphysical charging mechanisms relevant to MCS electrification studies is charging during melting (Dinger and Gunn, 1946; Drake, 1968). According to Drake (1968), melting hydrometeors often develop convection currents that result in the transfer of bubbles to the surface. When the bubbles burst, a small amount of electric charge is then observed to be separated, resulting in positively charged melt water. Shepherd et al. (1996) concluded that this mechanism was likely responsible for the charge observed near the freezing level of MCS stratiform regions.

1.3 Scientific objectives of this study

The primary objectives of this study are twofold. First, we seek to observationally investigate the concurrent kinematic, microphysical, and electrical structures of two MCSs, one with a symmetric structure and one with an asymmetric structure. Second, we investigate electrification mechanisms that might account for observed MCS electrical structures through numerical modeling simulations. In particular, some of the questions investigated in the observational component of this study are:

- Are there identifiable kinematic and microphysical differences that can explain the unique (and repeatable) symmetric and asymmetric electric field structures identified by Marshall and Rust (1993)?
- Are the charge transitions within a specific type of MCS (symmetric or asymmetric) related to any identifiable kinematic and microphysical features?
- Do the composite kinematic, microphysical, and electrical observations offer any insight into the actual mechanisms responsible for MCS electrification?
- How might the different electrical structures be related to lightning discharges?

Some of the questions investigated in the modeling component of this study are:

- What are the relative contributions of 1) charge advection from the convective line, and 2) in situ charge generation due to local microphysical interactions?
- What local microphysical charging mechanisms, if any, are important?
- How might an MCS's kinematic flow structure influence the distribution of charge in the stratiform region?

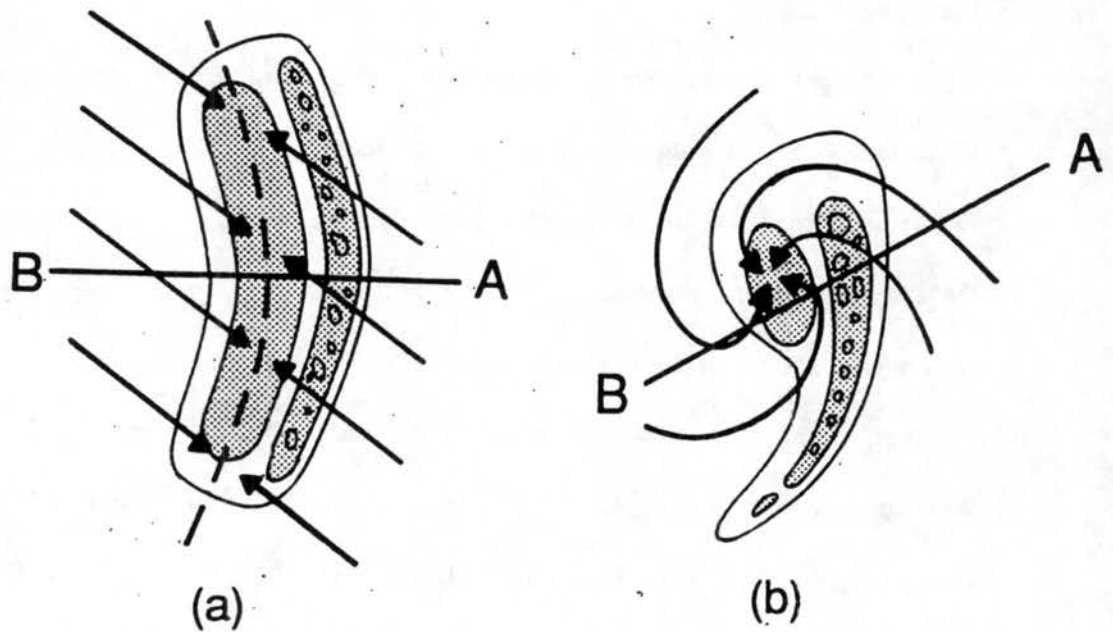


Fig. 1.1. Conceptual model of mid-level horizontal cross-section through (a) a typical symmetric MCS, and (b) a typical asymmetric MCS. In each case, the mid-level, system-relative airflow is displayed on top of the low-level radar reflectivity pattern. Light shading indicates regions of higher reflectivity. (From Houze et al., 1989).

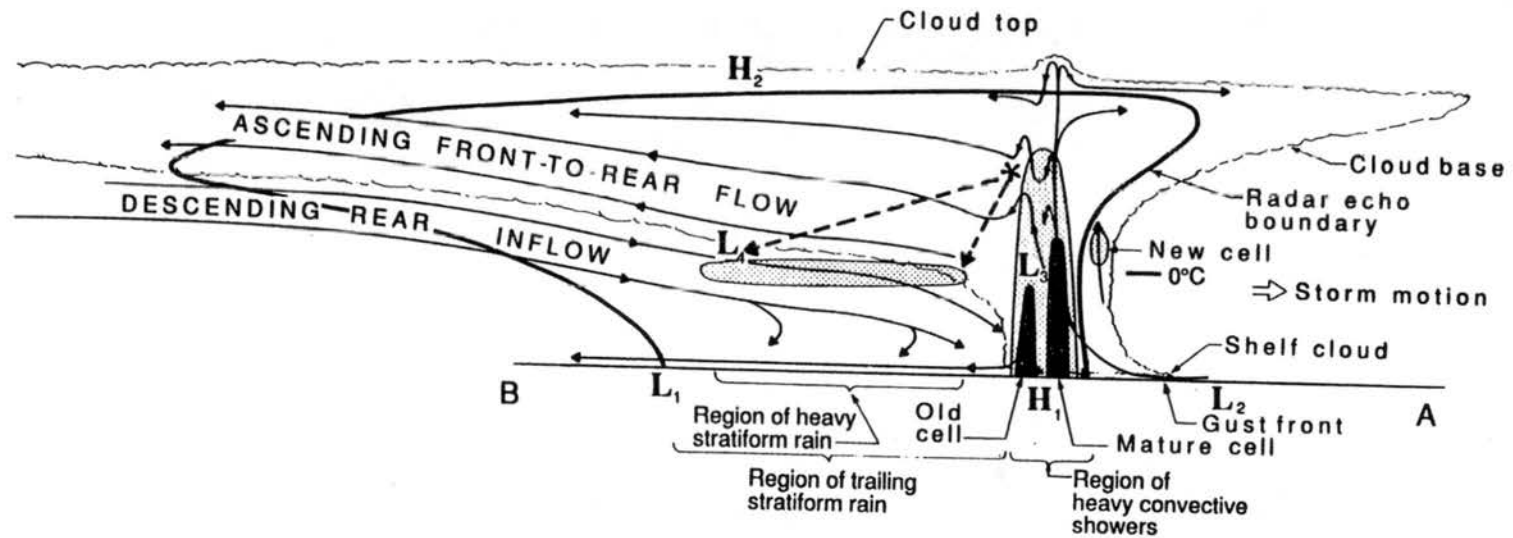


Fig. 1.2. Conceptual model of the kinematic and precipitation structure of a symmetric MCS as viewed in a vertical cross-section oriented normal to the convective line (cf. Fig. 1.1). (From Houze et al., 1989).

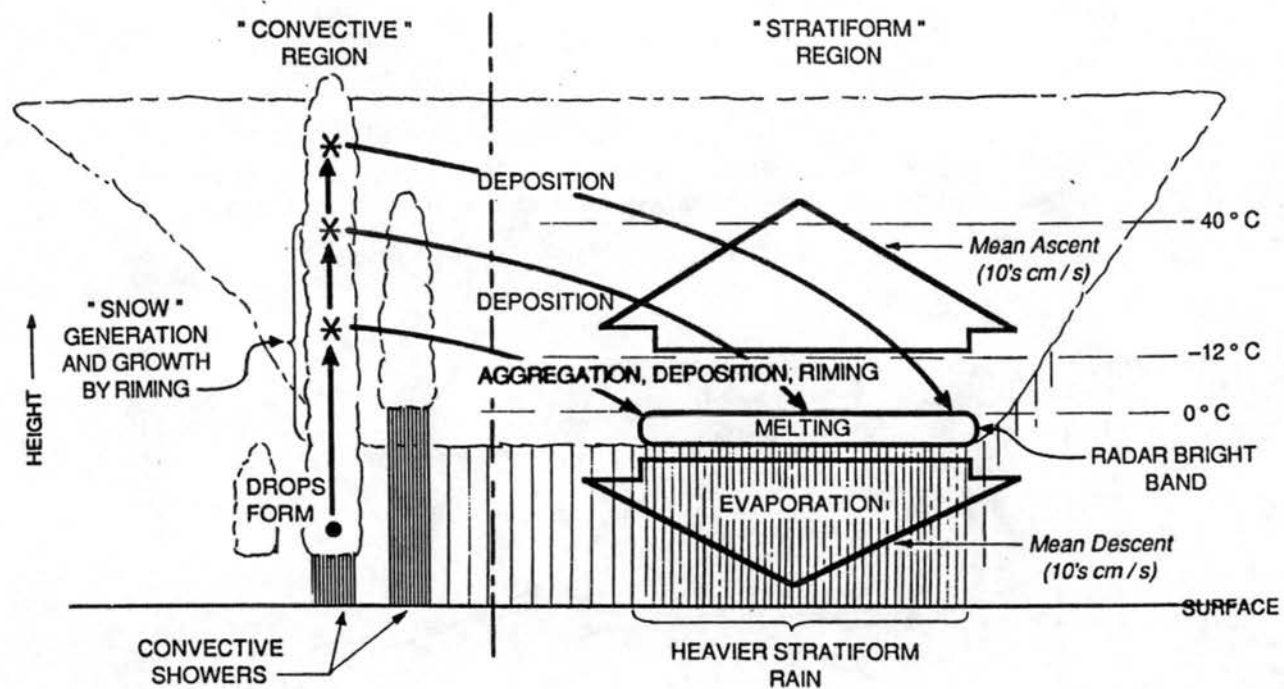


Fig. 1.3. Conceptual model of the microphysical structure of a symmetric MCS as viewed in a vertical cross-section oriented normal to the convective line (cf. Fig. 1.1). (From Houze, 1989).

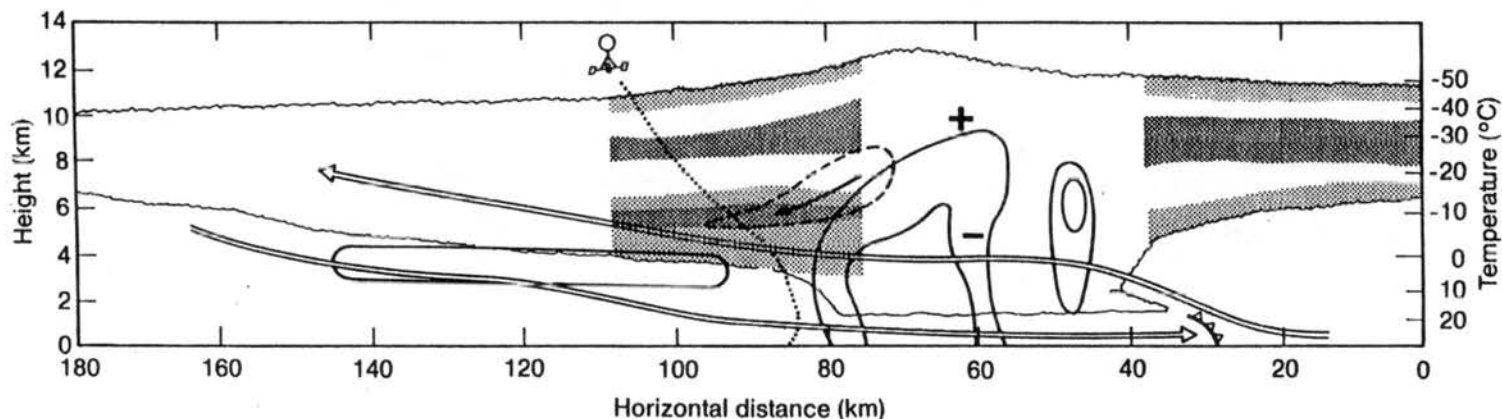


Fig. 1.4. Conceptual model of the electrical structure of a symmetric MCS as viewed in a vertical cross-section oriented normal to the convective line (cf. Fig. 1.1). Light shading indicates regions of negative charge while dark shading represents positive charge. Heavy lines enclose higher reflectivity region (i.e., >35 dBZ) within the convective and stratiform regions. Streamlines depict the primary components of the mesoscale airflow (cf. Fig. 1.2). Local maximum in graupel mixing ratio (adapted from Rutledge and Houze, 1987) is delineated by the dashed line within the transition zone; arrow within this region indicates the approximate system-relative path followed by graupel particles detrained from the convective line (see text). Dotted line depicts the path of the balloon-borne EFM. (From Schuur et al., 1991).

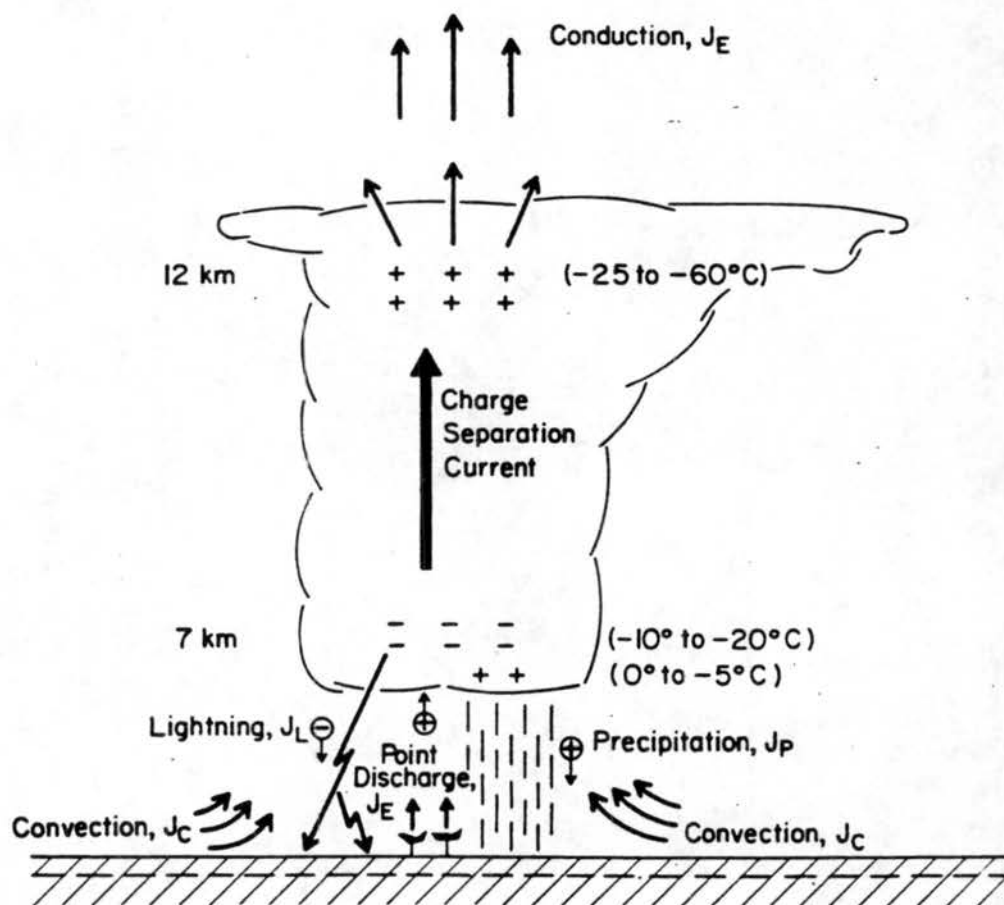


Fig. 1.5. Schematic showing the tripole thunderstorm charge structure. Positive charge is indicated by (+) and negative charge by (-). (From Levin and Tzur, 1986).

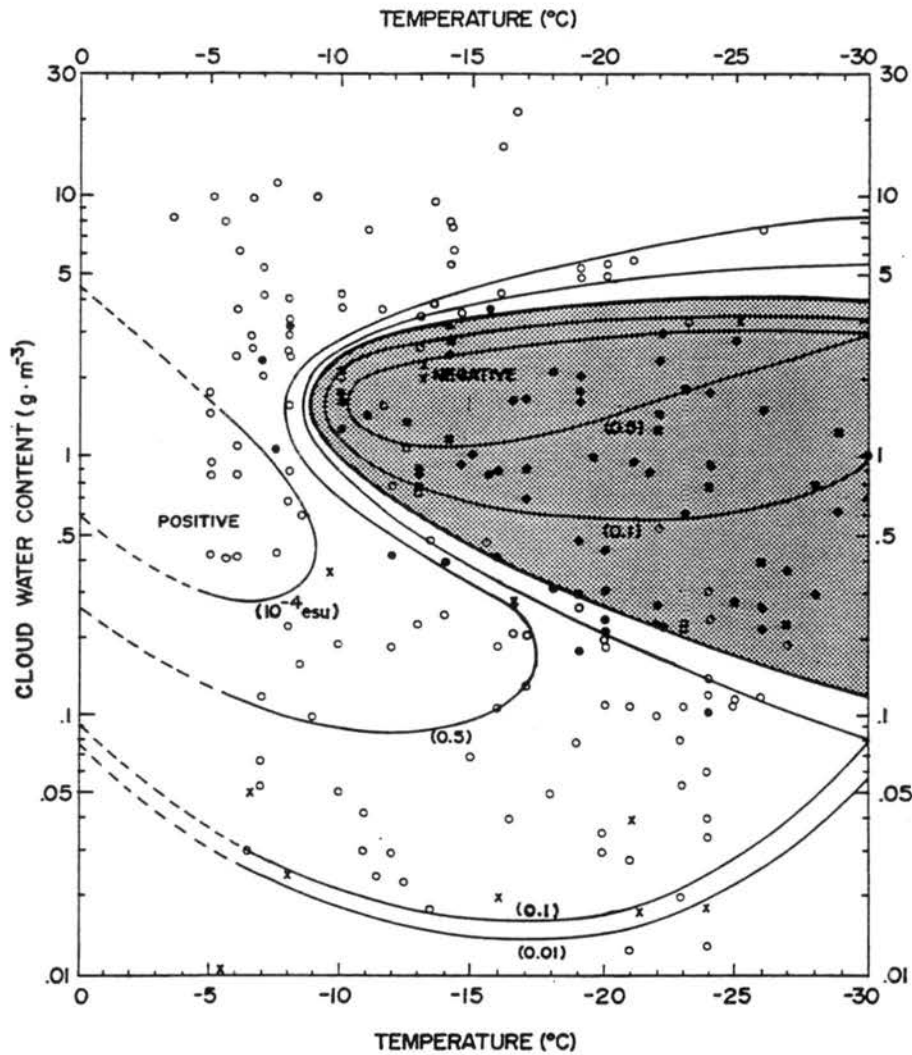


Fig. 1.6. Positive and negative rime charging zones as a function of temperature (°C) and cloud water content (g m^{-3}) from the laboratory experiment of Takahashi (1978). Open circles represent positive charge, solid circles negative charge, and crosses uncharged cases. The electric charge transferred to the rime per ice crystal collision is shown in units of 10^{-4} esu (electrostatic unit), where 10^{-4} esu is 33 fC.

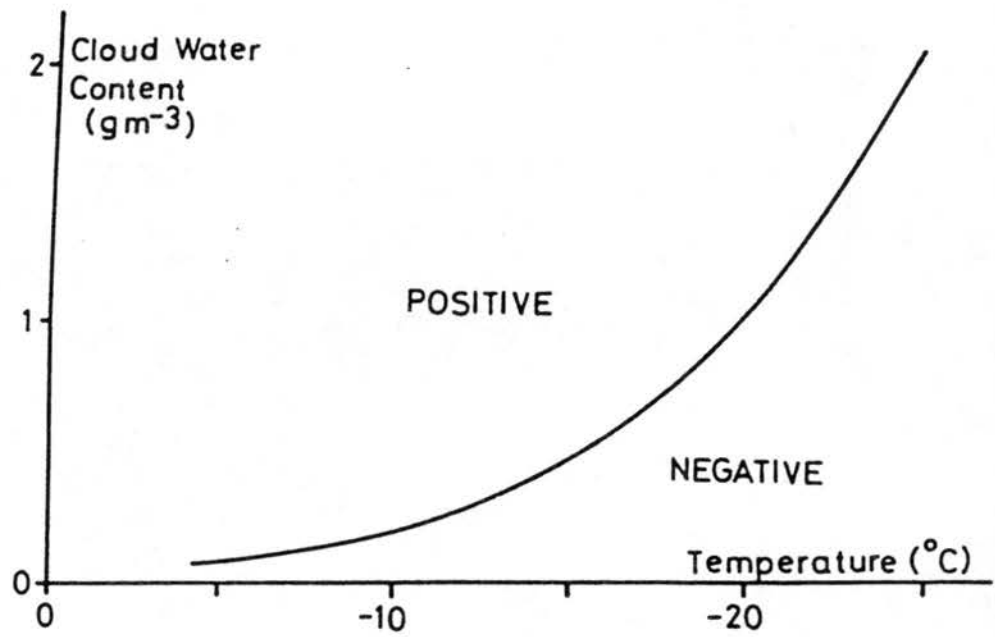


Fig. 1.7. Positive and negative rime charging zones as a function of temperature (°C) and cloud water content (g m^{-3}) from the laboratory study of Jayaratne et al. (1983).

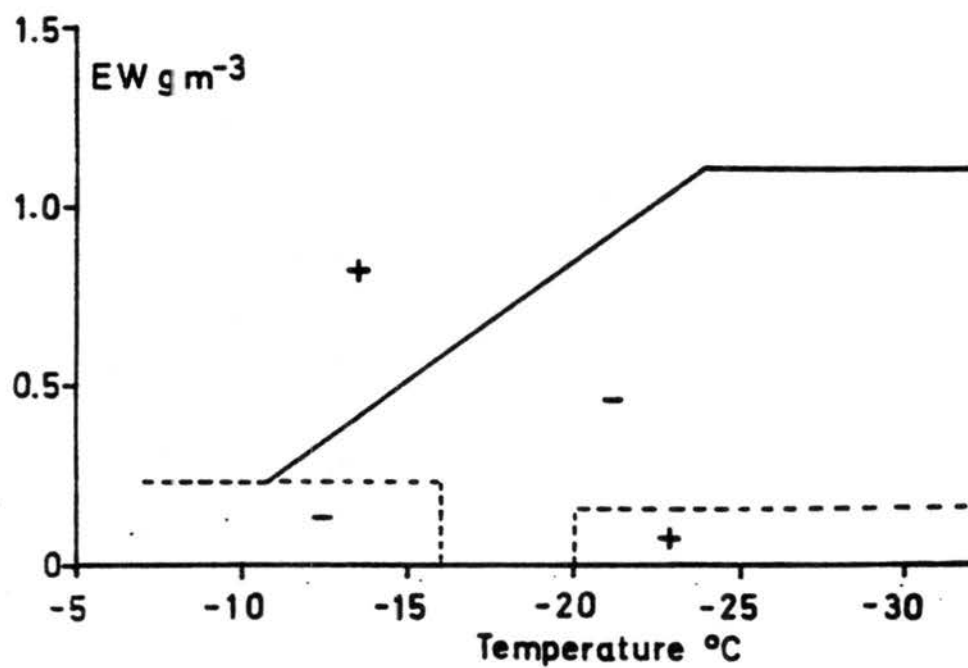


Fig. 1.8. Positive and negative rime charging zones as a function of temperature ($^{\circ}\text{C}$) and effective liquid water content (g m^{-3}) from the laboratory experiment of Saunders et al. (1991).

CHAPTER 2

DATA ANALYSIS

During the Spring of 1991, the National Severe Storms Laboratory (NSSL), in collaboration with several University groups, conducted the COPS91 (Cooperative Oklahoma Profiler Studies) project over central Oklahoma. One of the primary objectives of COPS91 was to investigate the charge structure and electrical development of midlatitude MCSs. Whereas earlier MCS electrification studies (e.g., Schuur et al., 1991; Hunter et al., 1992) typically relied upon single EFM profiles and limited supplemental data, the COPS91 experiment focused on the collection of multiple EFM data sets and simultaneous in situ aircraft measurements of MCS kinematic, microphysical, and thermodynamic structures. The observational component of this study focuses on data obtained in two MCSs observed during the COPS91 project, one symmetric and one asymmetric (observed on 24 May 1991 and 8 May 1991, respectively). In this chapter, we present an overview of the data sources and analysis methods used in this composite study.

2.1 Data sources

2.1.1 Electric field meter data

The primary electrical data source used in our observational analysis are vertical profiles of electric field as measured by balloon-borne electric field meters (EFMs). During COPS91, two NSSL mobile laboratories (hereafter referred to as NSSL1 and NSSL2, Rust, 1989) were deployed to locations in advance of developing MCSs. EFMs were then launched into the MCSs as they passed over the launch sites. The locations of the EFM launch sites with respect to the

state of Oklahoma and the NSSL Cimarron (CIM) and Norman (NRO) radars (discussed later) are depicted in Fig. 2.1. The EFM's used during the COPS91 project (see Fig. 2.2) were modifications of that presented by Winn et al. (1978). In short, the EFM consists of two metallic spheres centered on a 1.5 m fiberglass pole. One of the styrofoam rhomboid vanes, which are located at either end of the pole, contains a small electric motor that rotates the spheres about the horizontal axis at a rate of 2.5 Hz. The instrument measures the equal and opposite charges induced on the spheres by the electric field; from this measurement, the magnitude of the vertical component of the electric field can be deduced (Winn and Byerley, 1975). At the same time, air motion past the rhomboid vanes causes a rotation about the vertical axis at a rate of approximately 0.125 Hz and results in a signal modulation, from which the horizontal component of the electric field can be discriminated (Winn et al., 1978). Electric field data are received by the mobile laboratories, digitized, and continuously recorded. A complete description of EFM data processing is provided by Stolzenburg (1996). According to Marshall et al. (1995b), the EFM provides electric field measurements to an accuracy of 10%.

The EFM is flown on a 1200 g helium filled balloon, at the end of an instrument chain (see Fig. 2.3) that includes a parachute, Vaisala RS80 radiosonde and, occasionally, a particle/charge measuring instrument (e.g., Bateman et al., 1994; Bateman et al., 1995). Both mobile laboratories are equipped with mobile Cross Chain Atmospheric Sounding Systems (Rust et al., 1990) that receive the thermodynamic and wind information from the sonde and computes dewpoint and geopotential height at 10 s intervals. The Vaisala sondes are accurate to within $\pm 0.2^{\circ}\text{C}$ for temperature, ± 0.5 mb for pressure, and $\pm 2\%$ for relative humidity. This results in geopotential height accuracies of 10 m at the surface, 20 m at 8 km MSL, and 50 m at 16 km MSL (Stolzenburg, 1996). In post-processing, the thermodynamic and electric field data sets are interpolated and combined to provide vertical profiles of electric field (e.g., Schuur et al., 1991).

In our analysis, E denotes the magnitude of the total electric field vector $\mathbf{E}$, E_z the magnitude of the vertical component of $\mathbf{E}$, and E_h the magnitude of the horizontal component of $\mathbf{E}$. E_z is taken to be positive when an upward directed force is exerted on a positive test charge. Therefore, in the fair weather atmosphere where the Earth has a negative charge, E_z is negative (i.e., E_z is directed downward) and has a value of approximately -100 V m^{-1} . As described earlier, the EFM is capable of measuring both E_z and E_h . However, E_h is much more difficult to measure accurately than E_z (e.g., Stolzenburg, 1993) since 1) the spin rate of the EFM about the vertical axis is dependent on air motion past the rhomboid vanes, which can be quite variable in regions where the balloon's ascent rate changes dramatically (e.g., near the 0°C isotherm where the balloon frequently accumulates ice), and 2) the EFM's rotation about the vertical axis merely produces a high frequency modulation to the slower changing signal resulting from the primary (motor driven) rotation about the horizontal axis. This makes E_h more difficult to ascertain. When E_h is negligible or poorly sampled, the electric field profile appears smooth; when E_h is significant, the profile appears "noisy". In the cases where E_h is not negligible, a trace of the electric field profile's outer envelope is representative of E and a trace of the inner envelope is representative of E_z (e.g., Winn et al., 1978). In theory, E_h may then be extracted from the profile according to

$$E_h^2 = E^2 - E_z^2 \quad (2.1)$$

The electric field profiles presented in this study typically show horizontal electric field magnitudes that are small when compared to those in the vertical.

The charge density ρ of a region of charge may be computed using Gauss' Law in differential form, which states that ρ is proportional to the divergence of the electric field vector.

That is,

$$\nabla \cdot \mathbf{E} = \frac{\rho}{\epsilon_0} \quad (2.2)$$

where ϵ_0 ($8.86 \times 10^{-12} \text{ F m}^{-1}$) is the permittivity of air. Due to the aforementioned difficulties in measuring E_h , however, and also since this study focuses on electric field measurements in the MCS stratiform clouds, which typically exhibit stratified thermodynamic and microphysical environments, we neglect contributions from E_h and assume that the charge is stratified into “layers” of infinite horizontal extent (i.e., $\partial E_h / \partial h = 0$). Similar assumptions have been made in previous studies of both convective (e.g., Byrne et al., 1987; Marshall and Rust, 1991; Marshall et al., 1995a; Stolzenburg, 1996) and stratiform (i.e., Chauzy et al., 1980; Chauzy et al., 1985; Byrne et al., 1989; Marshall et al., 1989; Schuur et al., 1991; Hunter et al., 1992; Marshall and Rust, 1993; Stolzenburg et al., 1994) clouds. In keeping with these assumptions, we confine the analysis to appreciably deep regions wherein $\partial E_z / \partial z$ is nearly constant. Equation 2.2 can then be approximated as

$$\rho \approx \epsilon_0 \left(\frac{\Delta E_z}{\Delta z} \right) \quad (2.3)$$

Therefore, when the balloon is ascending, the slope of the electric field profile has the same sign as the polarity of charge found within that layer and the magnitude of the slope is proportional to the magnitude of the charge density. Whereas many previous studies (e.g., Schuur et al., 1991; Hunter et al., 1992) have used this analysis method to examine charge regions in detail, the analysis presented in this study follows the more restrictive criteria set forth by Stolzenburg et al. (1994) in that changes in E_z that result in charge layers wherein either 1) $|\rho| < 0.1 \text{ nC m}^{-3}$, 2) $\Delta z < 200 \text{ m}$, or 3) $|\rho \Delta z| < 50 \text{ nC m}^{-2}$, are ignored. An example of an electric field profile and a table showing charge densities derived using this approach are shown in Fig. 2.4 and Table 3.1, respectively. As can be seen by examining Fig. 2.4, which depicts our first symmetric MCS

transition zone electric field profile, E_z is as large as 90 kV m^{-1} whereas E_h (determined using Eqn. 2.1) was typically $<30 \text{ kV m}^{-1}$ (not shown). The remainder of the symmetric and asymmetric MCS EFM data used in our study will be presented in Chapter 3.

The final assumption that we make in the EFM analysis is that the change in E is primarily the result of the balloon's ascent through the cloud. If we assume that the balloon-borne EFM shares the air motion and additionally experiences a vertical displacement w' relative to the air, then the total derivative of E is given by

$$\frac{dE}{dt} = \frac{\partial E}{\partial t} + u \left(\frac{\partial E}{\partial x} \right) + v \left(\frac{\partial E}{\partial y} \right) + (w + w') \left(\frac{\partial E}{\partial z} \right) \quad (2.4)$$

where u , v , and w represent the balloon's displacement due to the air motion and $w'(\partial E/\partial z)$ is the change in E due to balloon's vertical displacement relative to the air. Since the space charge and balloon-borne EFM share the same horizontal motion (i.e., are advected by the mean wind), $u(\partial E/\partial x)$ and $v(\partial E/\partial y)$ are essentially zero. Furthermore, scale analysis (Schuur et al., 1991) has shown that $\partial E/\partial t$ resulting from in situ charging is insignificant when compared to the change in the electric field resulting from the balloon's vertical displacement, which is typically $\sim 5 \text{ m s}^{-1}$. The $\partial E/\partial t$ resulting from lightning is more difficult to address. Though electric field profiles, especially those from convective cores, frequently exhibit significant field changes resulting from lightning discharges, the electric field profiles presented in this study generally do not. Additionally, when small field changes do exist in our data, we restrict our analysis to vertically coherent charge regions over which the *gradient* in E_z remains relatively unchanged. Finally, since several studies (e.g., Houze and Rappaport, 1984; Smull and Houze, 1985; Srivastava et al., 1986; Roux, 1988; Rutledge et al., 1988; Biggerstaff et al., 1993; Cifelli and Rutledge, 1994, and many others) have demonstrated that mesoscale motions are typically $< 1 \text{ m s}^{-1}$, we conclude that $w < w'$. Equation 2.4 can then be approximated as

$$\frac{dE}{dt} \approx w' \left(\frac{\partial E}{\partial z} \right) \quad (2.5)$$

That is, the change in the electric field profile is primarily the result of the balloon-borne EFMs ascent through each layer of charge.

2.1.2 Radar data

During the EFM flights, a National Oceanic and Atmospheric Administration (NOAA) P-3 research aircraft (*N43RF*) executed flight legs that documented the MCSs kinematic and microphysical structures. The primary source of radar data used in this study comes from the P-3's vertically scanning tail-Doppler radar, which is described in detail by Jorgensen et al. (1983) and Parrish (1989). The P-3 tail radar is a X-band (3.2 cm wavelength) radar that is mounted near the tail of the aircraft (see Fig. 2.5a). During the COPS91 project, the P-3 collected data using the fore-aft scanning technique (FAST, e.g., Frush et al., 1986; Hildebrand, 1989; Jorgensen and Smull, 1993; Jorgensen et al., 1996) whereby two scans of data are collected in sequence: one pointing approximately 25° forward, and the other approximately 25° aft of a plane normal to the aircraft track. As the aircraft executes a quasi-linear flight leg using this scanning strategy, a three-dimensional region of intersecting beams is swept out (see Fig. 2.5b) over which a dual-Doppler wind synthesis may later be constructed. When the antenna is rotated at its maximum rotation rate of 10 rpm (60° s⁻¹) at a typical aircraft speed of 120 m s⁻¹, the FAST scanning results in intersecting, adjacent beams at a horizontal resolution of approximately 1.5 km. Occasionally, when a higher density data coverage is desired, the tail-Doppler radar data is collected on only one side of the aircraft, thereby decreasing the time interval between adjacent beams and giving an improved horizontal resolution of approximately 750 m.

Prior to the construction of dual-Doppler wind fields, an aircraft motion of approximately 50 m s^{-1} , which is introduced into the data by the $\sim 120 \text{ m s}^{-1}$ ground speed and $\sim 25^\circ$ antenna pointing angle, is removed from the radial velocity data. At this time, velocity artifacts resulting from improper antenna alignment (as revealed by an examination of ground returns) are also removed. According to Jorgensen (1984), this process results in radial velocity estimates that are accurate to within 1.5 m s^{-1} . The data are then edited to dealias velocity ambiguities and objectively analyzed to grids with a data spacing of 1.5 km in the horizontal and 0.5 km in the vertical. A Shuman (1957) filter is then applied to the U and V winds which, assuming an error variance of 1.5 m s^{-1} in the U and V wind fields, results in a subsequent theoretical vertical velocity resolution of 1 m s^{-1} . Mass integration of divergence profiles using a downward method of integration are adjusted according to O'Brien (1970) in order to assure zero vertical velocities at both ground and cloud top. As will be described in Chapter 3, multiple dual-Doppler wind syntheses in this study are averaged to provide system-relative, slab-averaged composites that depict each MCSs precipitation and kinematic structures. Temporal and spatial averaging of dual-Doppler synthesized wind fields is a common method of resolving mesoscale vertical motions (e.g., Smull and Houze, 1987a, Biggerstaff and Houze, 1991a; Matejka and Schuur, 1991; Jorgensen and Smull, 1993).

Storm motions for each MCS were computed by tracking and averaging the motions of several convective cells along the line. From this, we computed storm motions of 11.7 m s^{-1} from 270° and 11.2 m s^{-1} from 285° for the symmetric and asymmetric MCSs, respectively. These motions were then removed from the synthesized wind fields to yield storm-relative winds. In addition, the storm motions were used to shift the radar observations for each flight leg to a common, system-relative frame of reference (located at the center time of each flight leg) in order to account for system motion during the time of data collection. It should be noted that the storm

motion computed for the asymmetric MCS is significantly different than that used by Jorgensen and Smull (1993), who presented a dual-Doppler analysis of the same system. The motion used in their study, however, was computed with respect to the MCS's mesovortex, which had a much more southerly component of motion than that of the convective line. We chose the convective line as our frame of reference since it allows us to better examine the system-relative flow structure and convective line's influence on the MCS regions sampled by the balloon-borne EFMs.

In addition to data from the P-3 tail-Doppler radar, ground-based radar data were collected by NSSL's 10 cm Cimarron (CIM) and Norman (NRO) Doppler radars during COPS91. Data collection problems that resulted in a failure of NRO to sample storm tops, however, precludes their use in the construction of three-dimensional wind fields. We therefore primarily use the ground-based radar data to 1) provide one-dimensional, line-normal, depictions of system evolution, and 2) for the construction of fine resolution horizontal reflectivity plots over which other data sources (i.e., aircraft flight tracks, lightning, etc.) may be overlaid and analyzed. Compositing National Weather Service WSR-88D radar data are also used to document system structure and evolution.

2.1.3 Microphysical data

In addition to radar data, the P-3 also collected in situ cloud microphysical data within each MCS. Microphysical data collected by the aircraft included two-dimensional particle image data in the cloud and precipitation size ranges by 2D-C (50-1600 μm) and 2D-P (200-6400 μm) optical array probes (OAPs), respectively. Additionally, cloud droplet liquid water contents were measured by a Johnson-Williams (JW) liquid water content meter. Knollenberg (1970, 1976) provides a description of data sampling techniques with two-dimensional OAPs while Jorgensen and Willis (1982) describe the implementation of the Particle Measuring System (PMS) OAPs

on the P-3 aircraft. In short, as the OAPs consist of a two-dimensional array of diodes. As cloud and precipitation particles pass through their field of detection, diodes in the array are occulted and the shadowed state of each diode is written to tape. Later, with a knowledge of the airspeed and data sampling rate, the image slices are reconstructed to provide two-dimensional images of each particle (hereafter referred to as shadow graph imagery). Some information regarding the composition and evolution of each particle may then be determined from the shape and other information such as temperature, liquid water content, and aircraft altitude. A polarization switch is also used to aid in the discrimination between water and ice in mixed-phase regions. The JW liquid water content probe uses a heated platinum wire to sense cloud water drops that are primarily less than 30-40 μm in diameter (Knollenberg, 1972) and has a resolution and accuracy of 0.1 g m^{-3} and $\pm 0.20\%$, respectively (Jorgensen, 1984). A description of the P-3's other instrumentation for the measurement of thermodynamic and kinematic variables is described by Jorgensen (1984). In this study, we did not utilize data from the Forward Scattering Spectrometer Probe (FSSP) to measure cloud droplet spectra since FSSP measurements have been shown to be unreliable in mixed-phase conditions (e.g., Hallett and Gardiner, 1985), which is typical of the environments sampled by the P-3 during COPS91.

In this study, computer software similar to that described by Jorgensen and Willis (1982) is used to process the 2D-C and 2D-P OAP data. The software objectively analyzes the shadow graph imagery by removing artifacts, discriminating between water and ice, sizing images, and creating particle number spectra for both cloud and precipitation-sized hydrometeors over 10 s averaging periods (e.g., Black and Hallett, 1986; Hendricks, 1993; Black and Hallett, 1995). Unfortunately, due to intermittent OAP vibration problems that resulted in the occasional misalignment of diodes, microphysical data were often of insufficient quality to produce time series of hydrometeor concentrations over complete flight legs. The data did, however, allow us

to determine occasional hydrometeor concentrations using this objective technique. These concentrations, most of which were collected in the MCS transition zone region, are used in the modeling component of this study to initialize microphysical profiles immediately behind the convective line. Nevertheless, due to the frequent poor data quality resulting from the OAP vibration (and additionally by intermittent tape drive problems that resulted in occasional 5 to 6 minute gaps in the OAP data record), our analysis of OAP data primarily relies upon a qualitative analysis and comparison of the available shadow graph imagery collected in each MCS. In this qualitative analysis, we follow the identification criteria specified by Black and Hallett (1986) to discriminate hydrometeor type.

Though the majority of the P-3 flight in each MCS was spent on constant altitude passes up and down the transition zone (to facilitate the collection of high quality tail-Doppler radar data), advecting spiral descents (e.g., Lo and Passarelli, 1982; Willis and Heymsfield, 1989) were also conducted in each system at the approximate time and location of EFM flights. In the advecting spiral descent, the aircraft spirals downward in a constant bank angle at the approximate fallspeed of the hydrometeors. Data from the advecting spiral descent in each system are qualitatively analyzed and compared to ascertain any microphysical differences that might account for the MCS's different electrical structures. An analysis comparing the OAP data of the advecting spiral descents in these two systems has also been completed by Hendricks (1993). Additionally, we compare cloud liquid water contents, as determined by the P-3 JW probe, by averaging the JW data by temperature level for each flight and comparing the resultant vertical profiles of cloud water content to the radar derived kinematic structure and OAP shadow graph imagery.

2.1.4 Supplemental data

In addition to the aforementioned, electrical, kinematic, and microphysical measurements, several supplemental data sets are also utilized in this study. Foremost, we use CG lightning data from the National Lightning Detection Network (NLDN), which consists of a network of magnetic direction-finder antennas across the contiguous U.S. that determine the location, strength, polarity, and multiplicity of CG flashes (e.g., Orville et al., 1983; Orville et al., 1991). The detection efficiency of CG flashes sampled by the NLDN (at the time COPS91 was conducted) was estimated to be between 75 and 90% and the accuracy of ground strike locations was considered to be within several kilometers. Other data sets, including surface electrical measurements made by NSSL mobile laboratories (e.g., Rust, 1989), will be discussed later as needed.

2.2 Coulombs law model

To test the appropriateness of the horizontal stratification assumption in the EFM data analysis (see Section 2.1.1), a simple three-dimensional charge model (e.g., Stolzenburg, 1993; Schuur and Rutledge, 1993; Stolzenburg and Marshall, 1994) was developed in which point sources of charge were uniformly distributed throughout the model domain. A simulated electric field profile was then constructed by the ascent of a hypothetical balloon through the steady-state charge structure at an ascent rate of 5 m s^{-1} (which is typical of our EFM flights). Electric fields at each point were then determined by summing the contributions from each point source of charge by using Coulombs Law, i.e.,

$$\mathbf{E} = \frac{1}{4\pi\epsilon_0} \sum_{i,j,k} \frac{q_{i,j,k}}{r^2} \hat{\mathbf{r}} \quad (2.6)$$

where r represents the distance from the balloon to the point source of charge. To assure that the hypothetical balloon's location never coincided with the location of the point charges, all point charges in the model were offset one-half grid space from the locations of the balloon ascent.

Since the Earth is a conducting surface, a charged cloud will induce a charge on the Earth's surface, which must also be taken into consideration when constructing the electric field profile. The induced charge at the Earth's surface in our model was approximated by the method of images (e.g., following Uman, 1987) wherein a hypothetical layer of charge of equal charge density but opposite polarity was placed an equal distance below the Earth's surface. Thus, the model is analogous to a series of embedded capacitors. Theoretically, as a capacitor's plate area approach infinity, end effects may be neglected and the electric field within the plates approaches a constant magnitude. Outside of the capacitor, the electric field is zero. This principle was used to test the numerical model by creating a simple, single charge layer profile and examining the rate at which electric fields approached their theoretically predicted values. As can be seen from Fig. 2.6, which depicts tests in which the model was initialized with a single 500 m deep layer with its base at 2 km and a charge density of 2 nC m^{-3} (typical of our EFM derived magnitudes), rather significant end effects are present for all charge layer widths less than approximately 20 km. Using a 10 km wide disk in the single-layer model, the maximum electric field predicted was 67% of that provided by the infinite approximation; at 20 km, the maximum model predicted electric field was 81% of that provided by the infinite layer approximation. These tests with a single-layer model suggest only a slightly greater sensitivity to disk width than presented by Stolzenburg and Marshall (1994), who tested entire profiles of charge derived from in situ EFM data and concluded that disks as narrow as 6 km in width closely approximated that derived by assuming infinite stratification.

Using a 10 km wide disk, which is typical of that used by Stolzenburg and Marshall (1994), we further tested the contributions of the horizontal electric field by ascending the hypothetical balloon through the model grid of the single-layer charge model at locations slightly offset from center. As can be seen from Fig. 2.7, which shows the results of balloon ascents at locations from 1 to 6 km from the disk center, horizontal electric fields are typically weak ($< 20 \text{ kV m}^{-1}$) and reductions in the vertical electric field small for balloon ascents removed from grid center by as much as 3 km. As the balloon approaches the edge of the disk (or is located outside of the disk as depicted in Fig. 2.7f), however, the reduction in the vertical electric field becomes significant and horizontal electric fields can be in excess of 50 kV m^{-1} . As will be shown in Chapter 3, the horizontal electric fields indicated by our EFM measurements were typically much less than this.

Finally, in a manner similar to that presented by Stolzenburg and Marshall (1994), we tested a complete vertical profile of charge density in our model. In this test we used a charge density profile derived from the electric field data depicted in Fig. 2.4. Due to the fine (25 m) vertical grid spacing used in our model, we used a slightly more detailed charge analysis than that presented in Table 3.1 (derived using the charge layer constraints presented by Stolzenburg et al., 1994) to initialize the model. Results from these model test runs, which tested disk widths from 2 through 10 km (the maximum disk width tested by Stolzenburg and Marshall, 1994), are shown in Fig. 2.8. As can be seen from Fig. 2.8, a disk width of 10 km appears to provide a reasonable approximation to the infinite horizontal stratification assumption. Stolzenburg and Marshall (1994) found that a disk width of 10 km resulted in a reduction in maximum electric fields by approximately 10%; our studies show similar results. As the disk size was decreased from 10 km, similar vertical profiles were modeled, though the intensity of the predicted electric fields were significantly reduced.

From our modeling test runs, we conclude the assumption of infinite horizontal stratification is indeed appropriate for use in this study. Though our results with the single-layer model show a greater sensitivity to “edge effects” than did that of Stolzenburg and Marshall (1994), our tests with complete vertical profiles of charge density yield similar results. That is, disks as small as 10 km provided reasonable approximations to the data derived using the infinite horizontal stratification assumption. This model result combined with the observations (e.g., Marshall and Rust, 1993; Stolzenburg et al., 1994, and the data presented in this study), which consistently show markedly similar MCS electric field profiles at locations separated in distance by as much as 100 km and time by as much as 2 hours, suggests that either 1) the charge is stratified into rather extensive layers (i.e., > 10 km in width), or 2) that the EFM observations in several MCSs by mere coincidence have, on a consistent basis, sampled localized regions of charge that exhibit the same charge structure. We believe the former case is more likely.

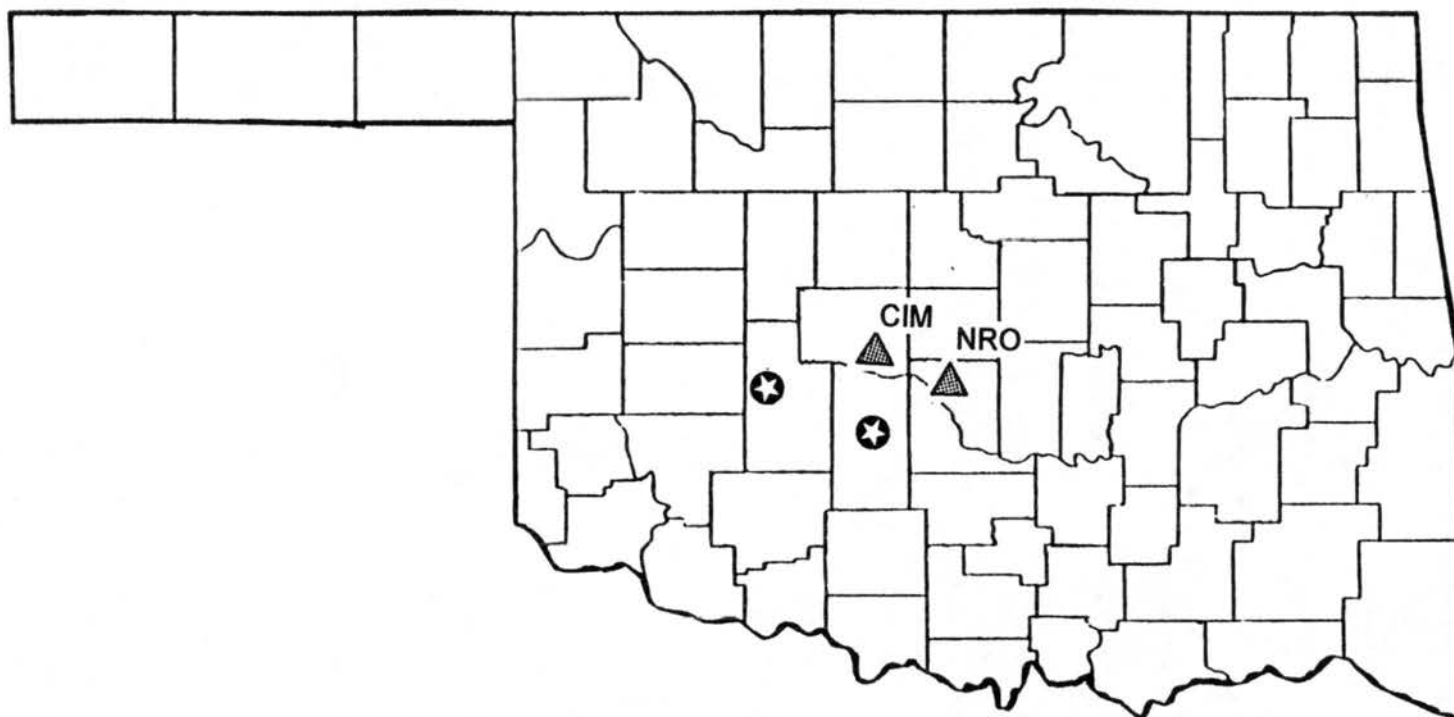


Fig. 2.1. Locations of NSSL Norman (NRO) and Cimarron (CIM) Doppler radars (denoted by shaded triangles) and electric field meter (EFM) launch locations (denoted by stars). Multiple balloon-borne EFMs were launched from both sites into the 24 May 1991 (symmetric) MCS and from the eastern site into the 8 May 1991 (asymmetric) MCS.

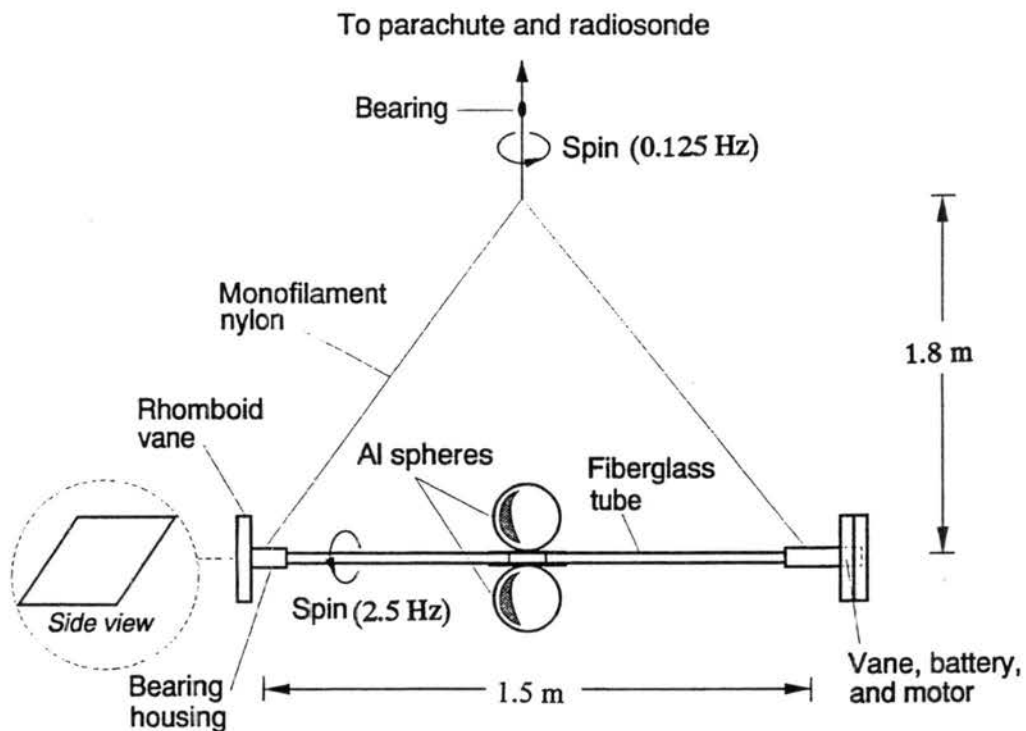


Fig. 2.2. Electric field meter used during COPS91. Equal and opposite charges induced upon the spheres by rotation about the horizontal axes allow for the computation of vertical and horizontal components of the electric field vector, respectively.

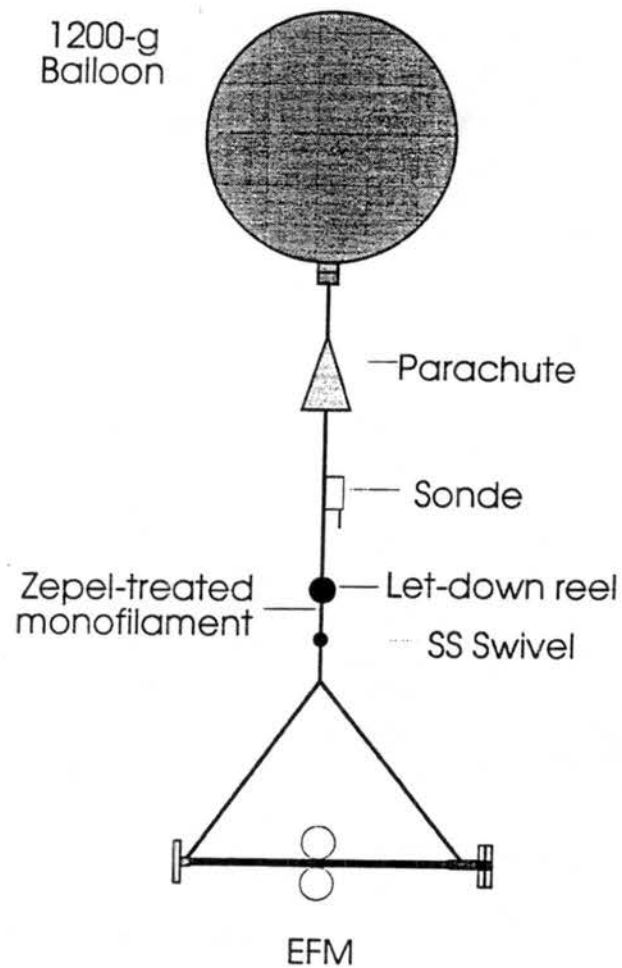


Fig. 2.3. Balloon-borne instrument train consisting of a 1200 g balloon, parachute, sonde, and electric field meter. (From Shepherd et al., 1996).

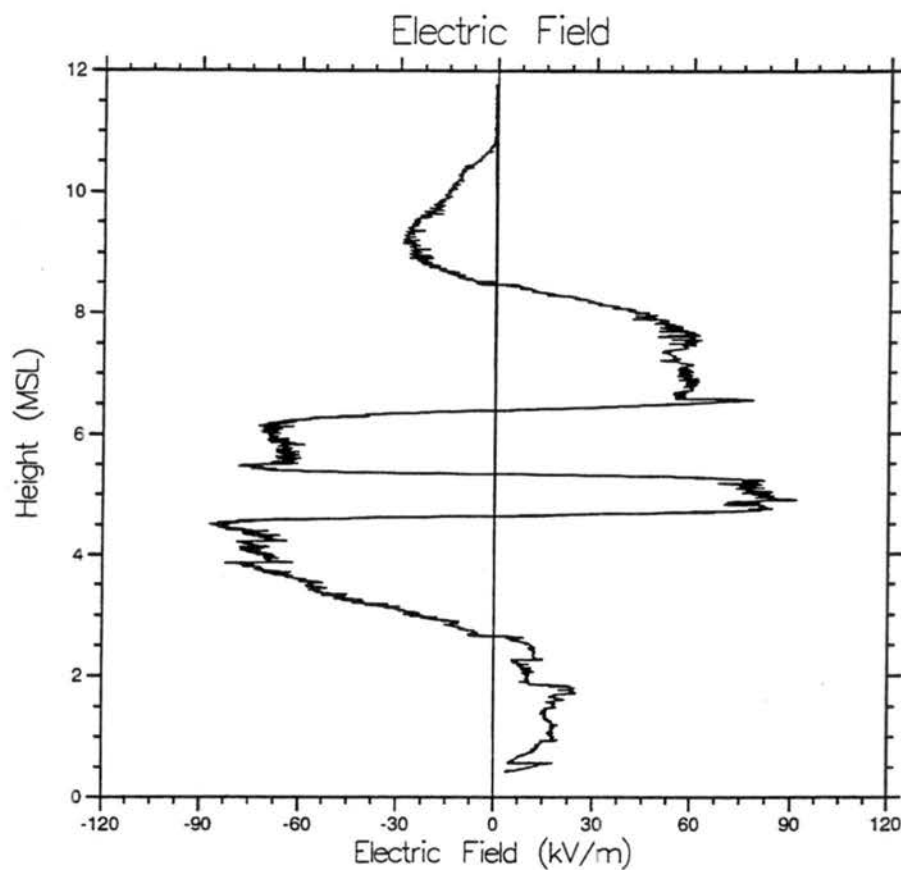


Fig. 2.4. Electric field profile obtained from a balloon-borne electric field meter flight launched near the leading edge of the radar bright band of the 24 May 1991 (symmetric) MCS at 0549 UTC. Charge densities derived from this profile using an approximation to Gauss' Law (see text) are shown in Table 3.1.

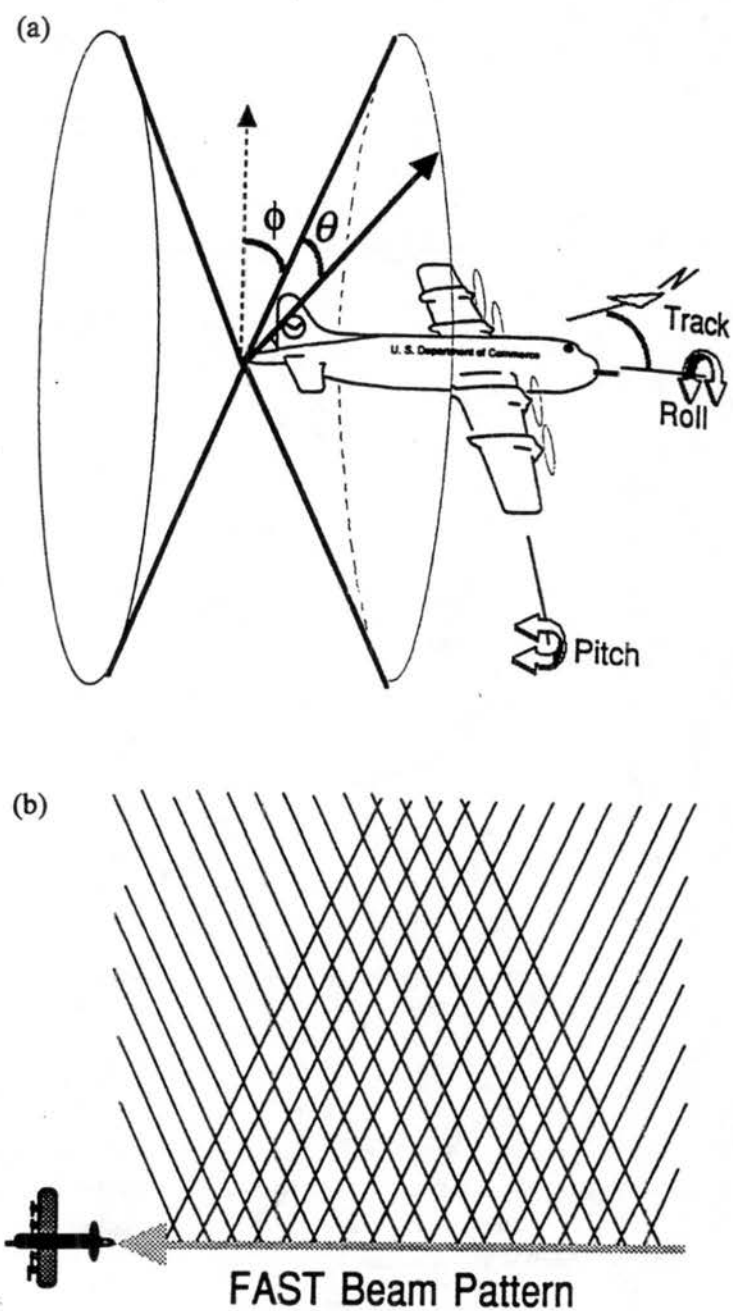


Fig. 2.5. (a) Basic geometry of fore-aft (FAST) scanning with respect to the P-3 aircraft. (b) Two-dimensional area swept out by FAST scanning during a quasi-linear flight leg.

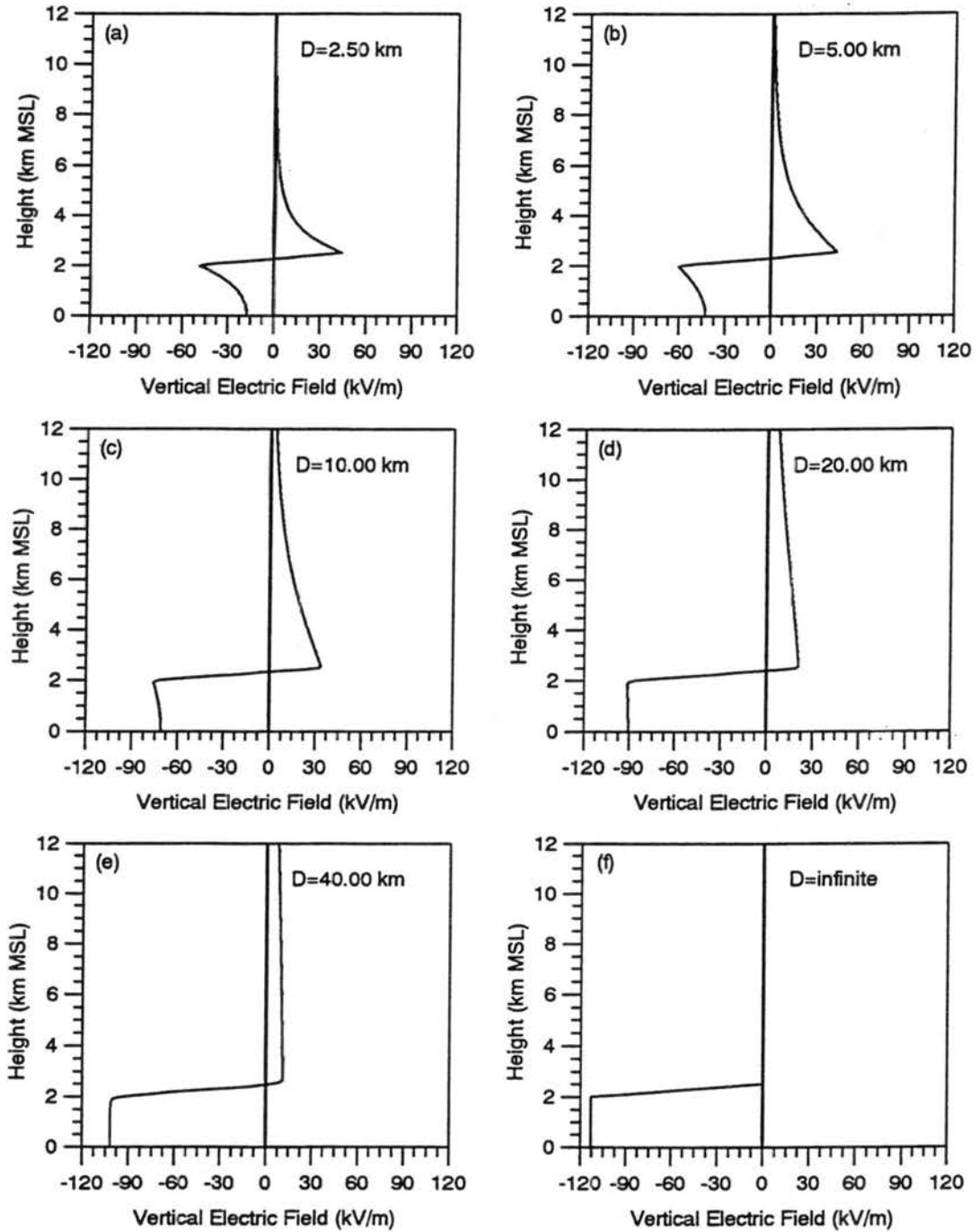


Fig. 2.6. Coulomb's Law model test of EFM ascent through charge disks of diameter (a) 2.5 km, (b) 5.0 km, (c) 10.0 km, (d) 20.0 km, (e) 40.0 km, and (f) $\gg 40$ km (infinite, determined theoretically) in width. The model disk used in this test was located between 2.0 and 2.5 km and had a charge density of 2 nC m^{-3} .

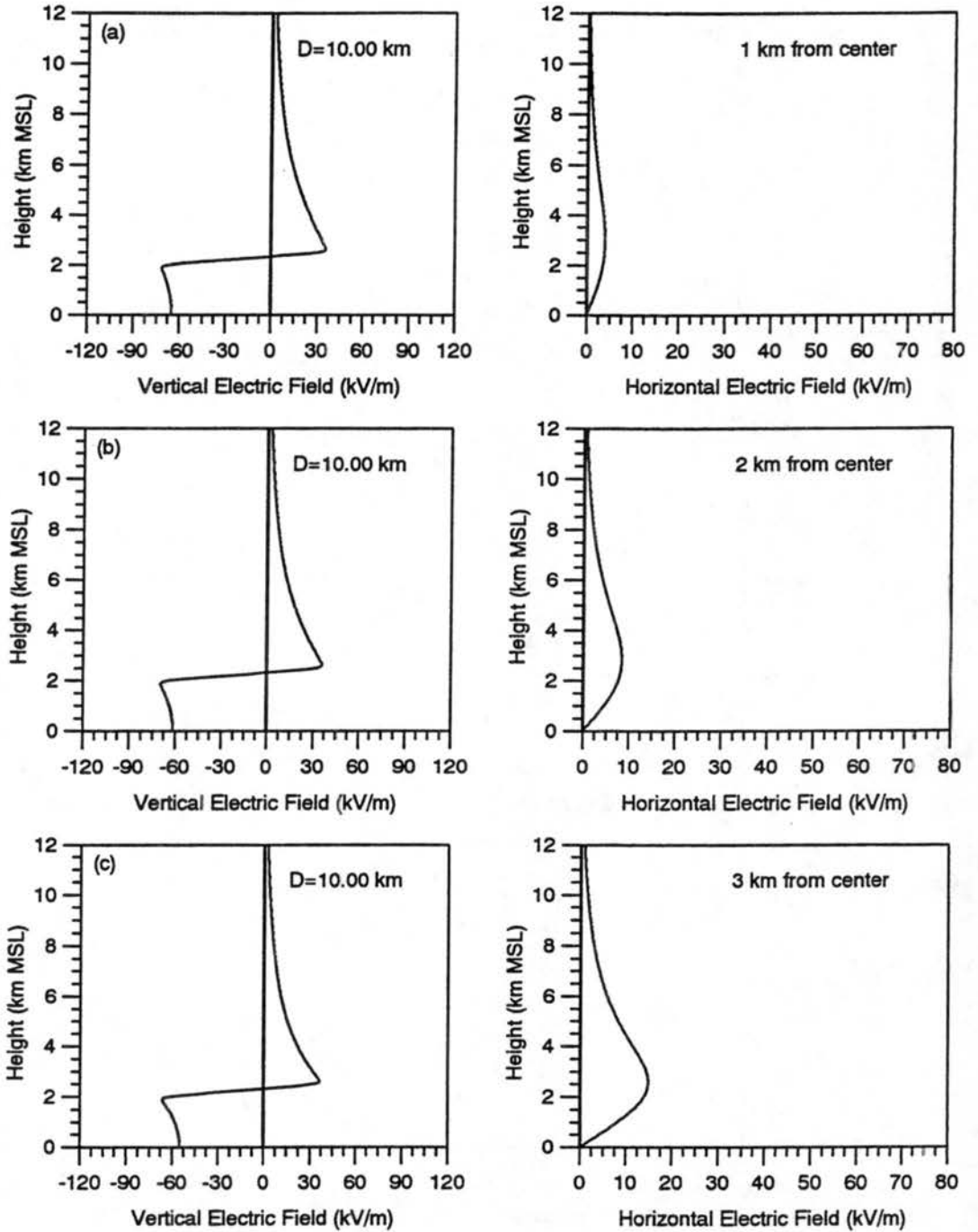


Fig. 2.7. Coulomb's Law model test of vertical and horizontal electric field components resulting from EFM ascent through a 10 km diameter disk of charge at locations (a) 1 km from the center, (b) 2 km from the center, (c) 3 km, (d) 4 km, (e) 5 km, and (f) 6 km (i.e., outside of the disk of charge) from the center of the disk. The model disk used in this test was located between 2.0 and 2.5 km and had a charge density of 2 nC m^{-3} .

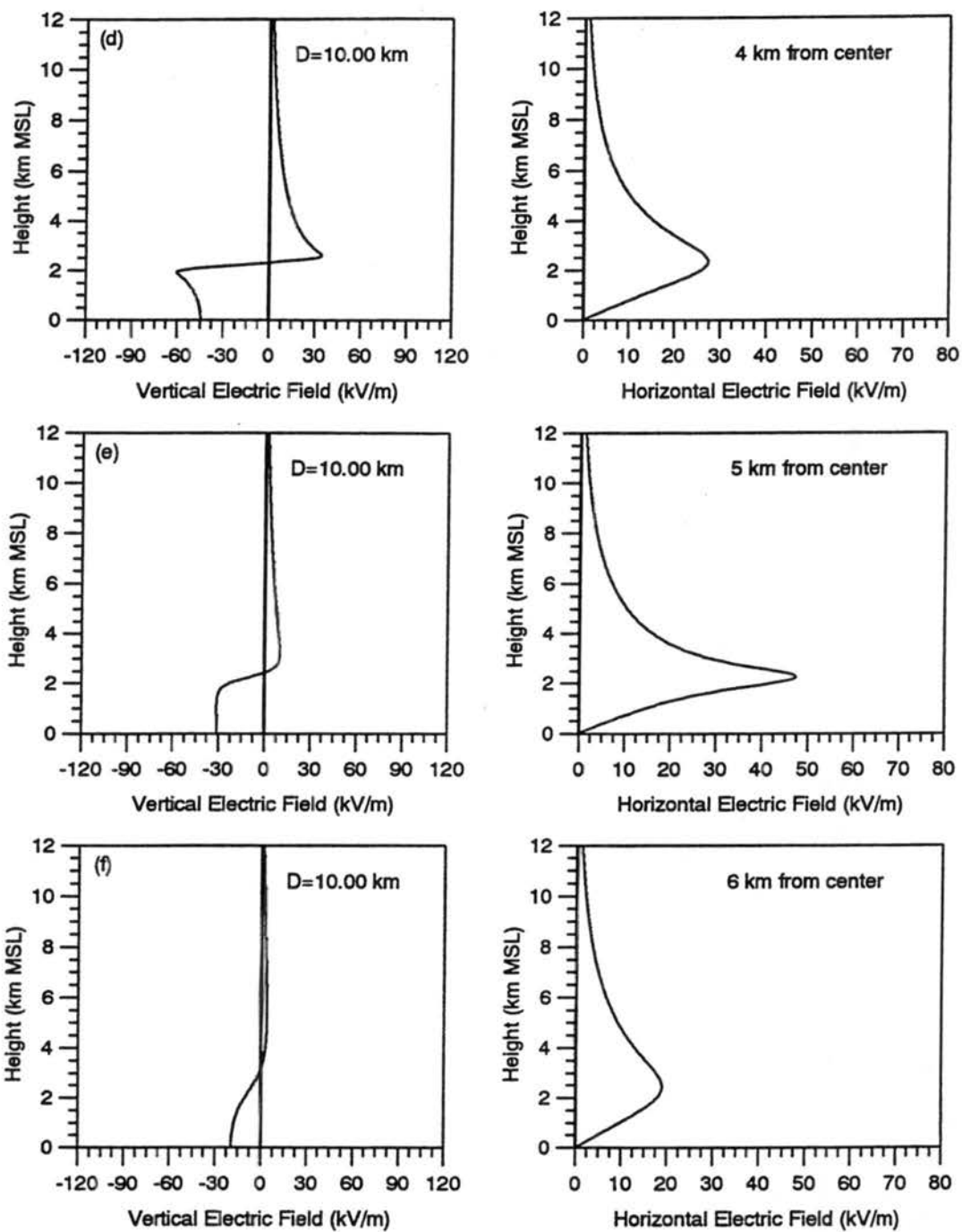


Fig. 2.7. Continued.

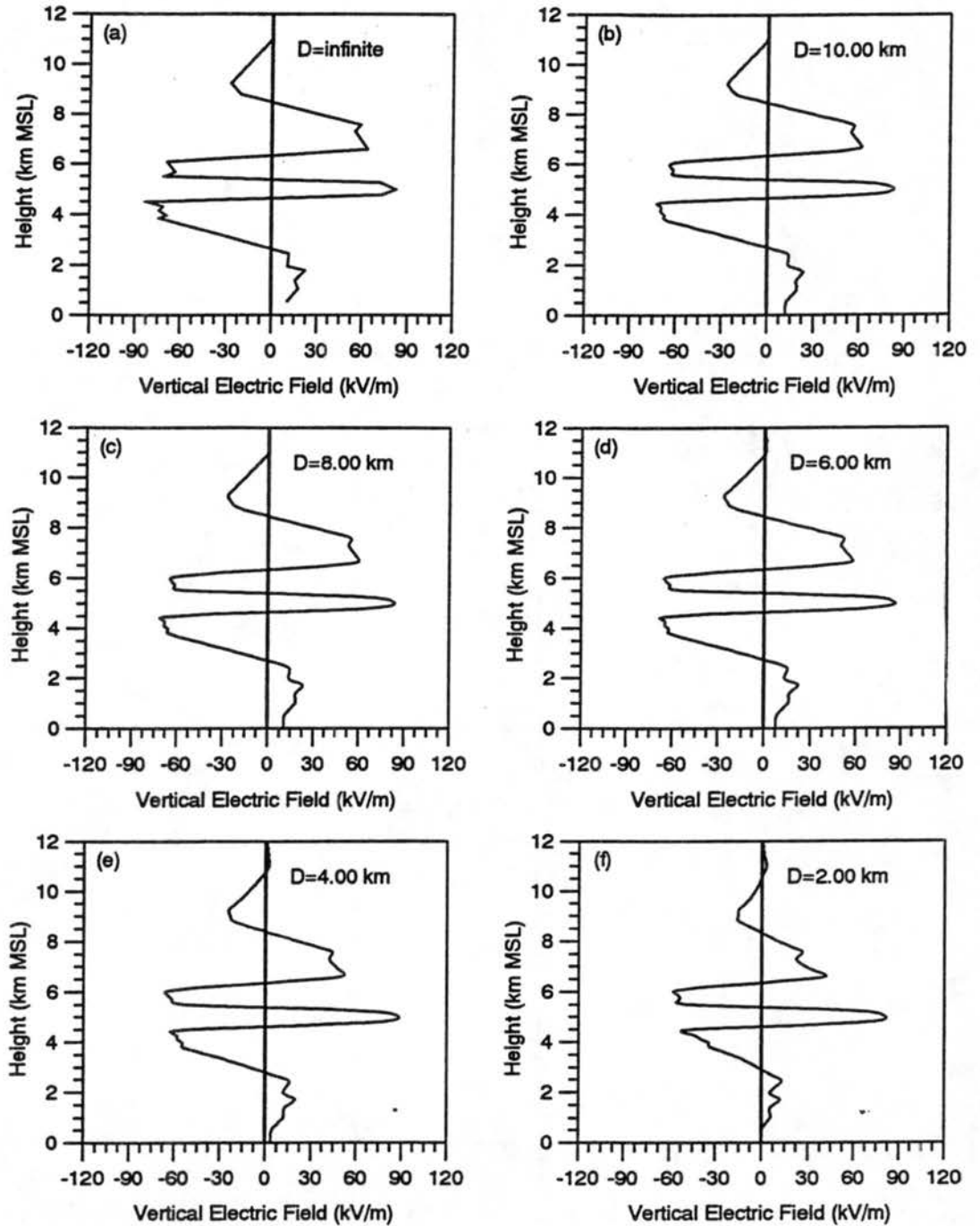


Fig. 2.8. Coulomb's Law model test of electric field profile depicted in Fig. 2.4 using (a) infinite (theoretically derived from Fig. 2.4 using Gauss' Law) (b) 10 km, (c) 8 km, (d) 6 km, (e) 4 km, and (f) 2 km wide disks of charge.

CHAPTER 3

KINEMATIC, MICROPHYSICAL, AND ELECTRICAL OBSERVATIONS

During COPS91, data sets documenting concomitant kinematic, microphysical, and electrical structures were collected in several MCSs. In this study, we focus on data from two MCSs: a symmetric MCS observed on 24 May 1991, and an asymmetric MCS observed on 8 May 1991. Marshall and Rust (1993), in a study that examined and compared the electric field structures of these two systems, concluded that their distinctly different electrical structures might be related to identifiable differences in MCS flow structure, microphysics, or both. In this study we seek to investigate this hypothesis by presenting a detailed analysis of the two MCSs kinematic, microphysical, and electrical structures. In this chapter, we present the data collected in the symmetric and asymmetric MCSs, respectively. Then, in Chapter 4, we compare the combined structures of the two systems and present conceptual models summarizing their identifiable differences. Finally, in the modeling component of the study, the symmetric MCS observations are used for model development and, additionally, to provide an observational basis for understanding the model output.

3.1 The symmetric MCS of 24 May 1991

3.1.1 Environmental conditions and MCS development

At 1200 UTC on 23 May 1991, a well defined, north-south oriented dry line extended across extreme eastern New Mexico. Ahead of the dry line, weak southeasterly surface winds were advecting moisture laden air from the Gulf of Mexico northwestward into central Oklahoma

(where dewpoints were as high as 20°C). Behind the dry line, winds veered to a west-southwesterly direction while dewpoints dropped to approximately 2°C. Meanwhile, above the surface, a strong southerly jet was centered over west Texas at 850 mb; at 500 mb, a weak short-wave trough was approaching from the southwest. These two features combined to destabilize the atmosphere over the Texas panhandle throughout the day as, by mid-afternoon, lifted indices over west Texas reached as high as -10°C. By 1600 UTC, a small region of intense convection had broken out over northwestern Oklahoma. Over the course of the next several hours, this convection became organized into a slow moving, southwest-to-northeast oriented convective line that produced substantial amounts of rainfall (> 14 cm) in northwestern Oklahoma. Meanwhile, by 0000 UTC, a north-south line of strong thunderstorms began to develop along the dry line over extreme eastern New Mexico.

Fig. 3.1 depicts the evolution of the radar reflectivity pattern of this MCS from 0300 through 0800 UTC (which encompasses the times of our aircraft and EFM observations). At 0300 UTC (Fig. 3.1a), the convective line was just beginning to overtake and merge with convection in northwestern Oklahoma. By 0400 UTC (Fig. 3.1b), the two convective regions had completely merged to form a quasi-linear, north-south oriented convective line. Between 0400 and 0500 UTC (Figs. 3.1b and 3.1c), radar observations by NSSL's CIM Doppler radar (presented later) indicated that the convective line had reached its most intense stage as several cells along the line exhibited radar reflectivities in excess of 60 dBZ. During this time, storm reports from western Oklahoma indicated that thunderstorm winds of $> 25 \text{ m s}^{-1}$ and hail up to 2 cm in diameter were common. By 0600 UTC, approximately 15 minutes after the launch of the first balloon-borne EFM, the convective line had weakened slightly; the stratiform precipitation region, on the other hand, had increased in intensity, and exhibited bright band reflectivities in excess of 45 dBZ.

3.1.2 MCS kinematic structure

3.1.2.1 P-3 flight track

At 0047 UTC, the NOAA P-3 left OKC and proceeded westward towards the developing MCS. The flight track and levels for this MCS, which are depicted in Fig. 3.2, can be compared to Fig. 3.1 to determine the underlying radar reflectivity structure associated with each leg. After spending 2.5 hours flying in clear air in advance of the line, during which time high density fore-aft tail-Doppler radar data were collected on a single side of the aircraft, the P-3 ascended to approximately 510 mb (-8°C) and, at approximately 0350 UTC, entered the MCS stratiform precipitation region at the MCS's northern end (see Fig. 3.2). For the next 3 hours (from 0400 through 0700 UTC), the P-3 conducted quasi-linear, north-south oriented flights legs up and down the MCS's transition zone and stratiform precipitation region. The first of these flight legs extended the entire length of the MCS; subsequent flight legs (10 total) were conducted towards the MCSs southern end where NSSL mobile laboratories had been deployed to launch balloon-borne EFM's.

During these flight legs, which were all flown at a near constant altitude of 5400 m, the primary data collection objective was to document the MCS's kinematic structure using fore-aft continuous scanning of the P-3's tail-Doppler radar. Shortly after 0700 UTC, the P-3 concluded its series of transition zone/stratiform region (TZ/SR) flight legs by conducting an advecting spiral descent in the trailing stratiform region (near the vicinity of a balloon-borne EFM launch at 0705 UTC) to study the microphysical structure of this region.

3.1.2.2 P-3 Doppler observations of horizontal flow structure

The observational component of this study seeks to develop conceptual models of the kinematic, microphysical, and electrical structures of each MCS. Since it is beyond the scope of this study to present a detailed analysis of every flight leg used (10 total), we focus on a detailed

look at dual-Doppler synthesized wind fields from a north-to-south flight leg that was conducted from 0512 through 0525 UTC (20 minutes prior to the first balloon-borne EFM launch on 24 May 1991). Since the MCS had reached a quasi-steady structure by this time, this analysis time is representative of the MCS precipitation and kinematic structure over the entire three hour observational period.

Fig. 3.3 depicts the low-level radar reflectivity pattern (composited from NSSL CIM radar data) at 0515 UTC, the location of the dual-Doppler grid box corresponding to the 0512-0520 UTC flight leg (see Fig. 3.2), and CG lightning strike locations in a 15 minute period centered on the dual-Doppler analysis time. Lightning observations are discussed in more detail later. The dual-Doppler, storm-relative winds and reflectivities at 8 levels are shown in Fig. 3.4. All analyses presented here are with respect to MSL where ground level over central Oklahoma is at approximately 400 m MSL.

As can be seen from Fig. 3.4a, storm-relative winds at 0.9 km had a strong ($> 10 \text{ m s}^{-1}$) northerly component in the lower levels. Though a weak rear-to-front component is also noticeable in the winds at this level, it is significantly smaller than seen in other symmetric systems (e.g., Smull and Houze, 1987b). Nevertheless, single-Doppler radar measurements by NSSL's CIM radar (presented later) and dual-Doppler analyses at other times indicate that a rear-inflow jet was indeed present within this system, though it appeared to be a more dominant feature at locations further to the north and at later analysis times. Further evidence, albeit tenuous, of a rear-inflow jet over central Oklahoma is provided by the low-level reflectivity patterns depicted in Fig. 3.1, which indicate the presence of a weak reflectivity notch in the stratiform precipitation (Figs. 3.1c and 3.1d) and, later, a noticeable bowing of the convective line (Figs. 3.1e and 3.1f) during the MCS's dissipating stage (e.g., similar to the evolution of another symmetric MCS as discussed by Smull and Houze, 1985).

At 1.9 km (Fig. 3.4b), storm-relative winds were typically quite weak ($< 5 \text{ m s}^{-1}$) while, at 2.9 and 3.9 km (Figs. 3.4c and 3.4d, respectively), a noticeable transition had taken place to a dominant front-to-rear flow regime (similar to that depicted by the conceptual models of Figs. 1.1 and 1.2). At both of these levels, a significant line-parallel component is also apparent in the observations as the mean wind direction through and to the rear of the convective line was from approximately 135° . Furthermore, the analyses at 2.9 and 3.9 km both indicate a significant deceleration of the winds (by as much as 10 m s^{-1}) through the convective line, which suggests that the southeasterly, low-level momentum was being transported vertically upwards by the convective updrafts (e.g., LeMone, 1983; Smull and Houze, 1987a). This is supported by the analyses at higher levels (Figs. 3.4e through 3.4h), which all indicate the presence of strong (25 to 30 m s^{-1}) southeasterly winds from 4.9 to 7.9 km. At higher levels (not shown), upper-level divergence of the convective updrafts resulted in a slightly more southerly component to the airflow.

Though the analysis presented in Fig. 3.4 only depicts a small segment at the southern end of the convective line, analyses at other times (and particularly the 270 km long flight leg from 0350 to 0425 UTC, see Fig. 3.2) all indicate that this basic horizontal flow structure was rather unvarying over both space and time. Thus, overall, the reflectivity patterns and dual-Doppler (storm-relative) wind fields presented here provide a picture of the 24 May 1991 MCS that exhibits many marked similarities to that depicted for a symmetric MCS by the conceptual model in Figs. 1.1 and 1.2. At lower levels, the mesoscale flow pattern was dominated by a weak rear inflow jet that also had a slight northerly component; at upper levels the system-relative winds were southeasterly, but predominantly front-to-rear.

3.1.2.3 P-3 Doppler composites of vertical motion

In addition to the providing information about the horizontal flow structure, the synthesized wind fields were also used to construct divergence profiles which were integrated vertically (using the procedures discussed in Chapter 2) to obtain vertical velocities. Because of the many inherent errors in airborne Doppler radar observations (e.g., Hildebrand, 1989; Jorgensen et al., 1996) and, furthermore, uncertainties in determining the upper boundary condition (e.g., Shuman, 1957; Biggerstaff et al., 1991a; Jorgensen et al., 1996), it is often impossible to resolve mesoscale vertical motions accurately to within $10\text{-}50\text{ cm s}^{-1}$ (typical of mesoscale motions) using this technique. Therefore, most studies (e.g., Smull and Houze, 1987a; Matejka and Schuur, 1991; Biggerstaff et al., 1993) rely upon either spatial and/or temporal averages of dual-Doppler grids to accurately resolve mesoscale vertical motions.

In this study, we average vertical velocity data from 10 dual-Doppler grids to create a line-normal, system-relative composite. Thus, the resulting composite represents an average over both space and time (930 km of aircraft data collected over a 3 hour period) and is therefore most similar to the averaging technique used by Biggerstaff and Houze (1993). Several previous studies (e.g., Smull and Houze, 1987a, Matejka and Schuur, 1991) have presented line-normal composites of MCS vertical motion that were computed by simply averaging dual-Doppler grids in a north-south fashion. Since convective lines, however, are not always completely linear nor necessarily aligned with the dual-Doppler grid (as is often the case in our study, see Fig. 3.4), this method has the possibility of introducing significant errors into the analysis by aliasing higher vertical motions from more rearward positioned convective cells into the mesoscale vertical motion profile. Therefore, in our analysis, we composite the data with respect to two system-relative frames of reference: 1) the leading edge of the transition zone, and 2) the leading edge of the trailing stratiform precipitation region.

To create these system-relative composites, each dual-Doppler grid was subjectively analyzed to determine the east-west grid point (at each north-south location along the line) at which these two precipitation regimes were located. The leading edge of the transition zone and stratiform precipitation region were analyzed to be collocated with the largest reflectivity gradients at 1) the trailing edge of the convective line, and 2) leading edge of the radar bright band, respectively. A north-south array was then created for each analysis grid which listed these respective grid locations. Using this array, slab-averaged composites were then constructed by realigning and averaging all line-normal cross-sections into common frames of reference that reflected kinematic conditions that were *centered* at each of these grid locations (i.e., transition zone and stratiform). This procedure allows us to unambiguously address the vertical velocity structure of the transition zone and trailing stratiform precipitation region, thus minimizing aliasing of convective vertical motions. All dual-Doppler composites constructed in this manner were then averaged over time to develop the final, three hour space-time composite.

In the process of creating the composited profiles, several steps were taken to assure the quality of the data. Foremost, since the P-3's tail-Doppler radar is a vertically scanning radar, poor estimates of the horizontal winds are often made at high pointing angles, thereby resulting in a cone of data both above and below the aircraft wherein horizontal wind estimates are poor. This problem is frequently compounded at high pointing angles by contributions from falling hydrometeors. By examining dual-Doppler analyses at several times, it was determined that the contribution from these two effects was unacceptable at locations within 12.5 km of the aircraft track. Since subsequent flight legs were flown at a variety of distances behind the convective line (ranging from 10 to 40 km), however, this data gap does not appear in our space-time composite.

The results of this compositing technique are shown in Figs. 3.5 through 3.8. Fig. 3.5 depicts the line-averaged reflectivity and vertical motion profile, plotted with respect to the

leading edge of the transition zone, for a flight leg at 0408 UTC (at which time the convective line was at its most mature stage, cf. Fig. 3.1). Though all flight legs were composited in this manner, we show this one since it best reflects the vertical motion pattern that is exhibited in the corresponding space-time composite (Fig. 3.6). As can be seen from Fig. 3.5, a 10 km wide region of deep subsidence is found approximately 15 to 20 km behind the convective line. In agreement with the results of Biggerstaff and Houze (1993, hereafter referred to as BH93), this downdraft region appears to consist of two distinct parts: 1) an upper-level downdraft, which BH93 proposed was mechanically forced, and 2) a low-level downdraft, which BH93 speculated was primarily the result of microphysical processes (i.e., sublimation, evaporation, and melting). Trailing the deep subsidence region is the mesoscale updraft which, in the line-averaged composite, *precedes* the leading edge of the stratiform precipitation region bright band by approximately 10 km and exhibits upward vertical motions of $> 25 \text{ cm s}^{-1}$.

The same basic mesoscale vertical motions structure, albeit with weaker mean vertical motions, is shown by the space-time composite in Fig. 3.6. Since this represents a spatial *and* temporal average, certain evolutionary features are evidently lost in this analysis. For example, single-Doppler analysis of the MCS by the NSSL CIM radar indicated a slight rearward tilting of the MCS convection with time in a manner similar to that described by the modeling studies of Rotunno et al. (1988) and Fovell and Ogura (1988) and observationally by Rasmussen and Rutledge (1993). At earlier times (i.e., during the 0408 UTC flight leg, cf. Fig. 3.1b), the convection was more aligned vertically and intense. If the upper-level transition zone downdraft is indeed mechanically forced by either gravity waves or upper-level convergence (i.e., Biggerstaff and Houze, 1993), we expect that the compensating, upper-level subsidence in the transition zone would also be greatest when the convective line is at its most intense stage (i.e., at ~0408 UTC in our analysis). However, at the later times after the convective line had acquired a

more rearward tilt and was less intense, we expect that the compensating upper-level subsidence would also be less vigorous. These evolutionary features, if they indeed exist, are smoothed in our space-time composite.

A similar view of the mesoscale vertical motions is provided by Figs. 3.7 and 3.8, which depict these same analyses (0408 UTC flight leg and space-time composites, analogous to Figs. 3.5 and 3.6, respectively), only plotted with respect to a reference frame that is centered on the leading edge of the trailing stratiform region. Fig. 3.7, which depicts the mean reflectivity and velocity structures derived for the 0408 UTC flight leg, indicates that the mesoscale updraft exhibited a maximum velocity (in the composite) of approximately 50 cm s^{-1} at a height of 7 km (-15°C), situated near the leading edge of the stratiform precipitation region. The stratiform region updraft appeared to slope upwards towards the rear of the system, being situated above mesoscale descent. It is likely that this sloping mesoscale downdraft region is the result of a combination of the evaporation of precipitation (e.g., Brown, 1979) and the entrainment of dry mid-level air towards the rear of the system by a developing, descending rear-inflow jet (e.g., Smull and Houze, 1987b).

Though the corresponding space-time composite (Fig. 3.8) indicates much weaker vertical motions, it also suggests (in agreement with Figs. 3.5 through 3.7) that the mesoscale updraft was strongest and deepest near the leading edge of the stratiform precipitation region. This finding is similar to that of Matejka and Schuur (1991), who suggested that the position, width, and intensity of the stratiform region rainband could be related to the trajectories of ice particles that are advected from the convective line followed by passage through this leading stratiform region updraft (which represents a modification of the seeder-feeder process discussed by Rutledge and Houze, 1987). In subsequent sections, we will discuss possible implications of this structure on MCS microphysics and electrification.

3.1.2.4 Ground-based Doppler observations

During aircraft operations, single-Doppler data collection by NSSL's Cimarron (CIM, located in central Oklahoma, see Fig. 2.1) radar documented the evolution of the MCS's precipitation and kinematic structures. A brief overview of these observations are presented to help understand the mechanisms that produced the structure seen in the composited vertical velocity. Reflectivity and radial velocity data from CIM sector scans were interpolated, at 15 minute intervals, to grids with a 1 km resolution in the horizontal and 0.5 km in the vertical. A storm motion of 11.7 m s^{-1} was then subtracted from the radial velocities to yield storm-relative winds. Since the MCS motion was from 270° , east-west cross-sections that pass through CIM provide an unambiguous depiction of the two-dimensional (i.e., east-west), system-relative flow structure. Furthermore, by comparing the location of CIM to the radar reflectivity patterns in Fig. 3.1, it is apparent that CIM was not only located due east of the reflectivity notch (Figs. 3.2a and 3.2b), but also near the apex of a distinct bow that appeared in the convective line from 0600 through 0800 UTC (both indicators of rear-inflow jet development, e.g., Smull and Houze, 1985). Hence, the CIM radar was well-situated to sample the rear inflow jet.

Though the aircraft data indicated that the upper-level, transition zone downdraft was most intense at approximately 0400 UTC, the rather long range from CIM prevented high resolution coverage of the MCS's transition zone and stratiform region. We present data from a CIM cross-section at 0540 UTC (5 minutes prior to the first balloon-borne EFM launch on 24 May 1991), which depicts radar reflectivity, storm-relative radial velocities, and the composited vertical velocity structure (from the previous section) overlaid on contours of radar reflectivity (Figs. 3.9a-c, respectively). In addition, the dark shading in Fig. 3.9c depicts two-dimensional (i.e., line-normal) convergence in excess of $1 \times 10^{-3} \text{ s}^{-1}$. Though these convergence calculations neglect any contributions from the line-parallel airflow component, analyses of several symmetric

MCS's by Rasmussen and Rutledge (1993) suggest that variations in the line-parallel flow are typically much smaller than variations in the line-normal flow. As a result, they concluded that "forcing", in a line-parallel sense, is typically an order of magnitude smaller than in the line-normal direction. This is not always the case for asymmetric systems. Since our dual-Doppler observations similarly indicate that the line-parallel component of airflow in this system varied relatively little from one end to the next, we suggest the conclusions of Rasmussen and Rutledge (1993) are valid here.

The two-dimensional cross-sections presented in Fig. 3.9 reveal a mesoscale structure consisting of two primary flow regimes: 1) an ascending front-to-rear flow that accelerates through the convective line, bridges the transition zone at approximately 15 m s^{-1} , and then decelerates to approximately 5 m s^{-1} in the trailing stratiform region, and 2) a rear-inflow jet that achieves maximum rear-to-front velocities of $\leq 15 \text{ m s}^{-1}$ in the low-levels of the transition zone. Two primary convergence maximum appear in the CIM data (Fig. 3.9): one immediately behind the convective line at cloud top, and a second near the leading edge of the stratiform precipitation region, immediately above the 0°C isotherm. In comparing these regions to the horizontal flow structure in Fig. 3.9b, it is apparent that the first of these two regions, in agreement with the findings of Biggerstaff and Houze (1993), is likely the result of strong convective outflow. The second appears to be collocated with the strong deceleration of the front-to-rear flow described earlier. This is a feature that was not resolved in the Biggerstaff and Houze (1993) analysis and appears to offer at least some qualitative support for the mesoscale updraft structure resolved in the composite analysis constructed from the airborne Doppler data wherein the stratiform updraft in the symmetric MCS was deepest and most intense near the leading edge of the stratiform precipitation region.

3.1.3 MCS microphysical structure

3.1.3.1 Optical array probe data

Two-dimensional shadow graph imagery from each of the TZ/SR flight legs, all of which were conducted at -8°C , consistently revealed a wide variety of particle habits, including lightly rimed columns, needles, and graupel with sizes as large as 2 mm. Aggregates were also occasionally observed but were typically much smaller than those observed in the stratiform precipitation region. This finding is similar to that of Yeh et al. (1991), who suggested that the lower reflectivity associated with the transition zone was likely the result of mean hydrometeor sizes that were approximately half of those found in the trailing stratiform precipitation region (even though the particle *concentrations* in the transition zone were much larger). Given the OAP instrumentation problems during the COPS91 project (described in Chapter 2), it is difficult to similarly address this issue. Nevertheless, using the software and artifact elimination procedures described in Chapter 2, short segments of data from some flight legs were analyzed to reveal hydrometeor concentrations that were as high as 40 l^{-1} for the precipitation-sized (200-6400 μm) particles and 300 l^{-1} for the cloud-sized (50-800 μm) particles at locations immediately behind the convective line. These concentrations are used later in this study to initialize the convective line's mid-level microphysical profile in the modeling component of this study.

We focus on observations from the P-3's advecting spiral descent, which was conducted at the time and approximate location of the second (0705 UTC) EFM launch on 24 May 1991. Unfortunately, due to tape drive problems, microphysical data were not recorded over much of the descent and only 5 minutes of high quality data were available for analysis. A summary of 2D-C and 2D-P shadow graph imagery collected at -8.5°C (0714-0716 UTC) in the advecting spiral descent are presented in Fig. 3.10. The most striking feature in both the 2D-C and 2D-P shadow graph imagery is the presence of significant dendritic structures, with sizes that ranged

from < 1 mm to 3 mm, at -8.5°C. It is well known that dendritic growth occurs under water saturated conditions (e.g., Magono and Lee, 1966; Ryan et al., 1976; Keller and Hallett, 1982, and others). Since these images were collected at -8.5°C, which was immediately below one of the more significant charge transitions (i.e., the change in the charge density polarity) in our symmetric MCS electric field profiles (presented later), this data suggests that water saturated conditions were present through a deep layer above the -8.5°C level.

The shadow graph imagery further indicate the presence of aggregates of dendrites. This is not surprising since Passarelli and Srivastava (1978) have indicated that horizontal oscillations of falling hydrometeors and the subsequent interlocking of dendritic arms can result in collection efficiencies as high as 140% in the -12 to -15°C layer (and possibly at warmer temperatures). Other studies have indicated that aggregation efficiency increases in the presence of high electric fields (e.g., Crowther and Saunders, 1973). The possible relationship between this observed microphysical structure, the previously documented symmetric MCS kinematic structure, and electrical observations will be summarized by a conceptual model in Chapter 4.

3.1.3.2 Cloud water

During the optical array probe imaging of hydrometeor spectra, cloud liquid water contents (LWCs) were simultaneously measured by the P-3 JW probe. In general, LWCs during most TZ/SR flight legs ranged from 0.1 g m⁻³ to 0.5 g m⁻³ with averages over most flight legs between 0.2 and 0.25 g m⁻³. Though averages computed over most these flight legs were typically > 0.2 g m⁻³, observations of localized pockets of enhanced LWCs as high as 0.5 g m⁻³ suggests the presence of embedded convective elements within the mean mesoscale updraft, which is in agreement with the structure indicated by the dual-Doppler observations presented earlier in this study. Additionally, data from the advecting spiral descent were averaged over the -16 to -8°C

layer, which encompassed one of the primary charge transitions in the symmetric MCS electric field profiles and also is immediately above the level where OAP imagery indicated the presence of significant dendritic particles. Cloud liquid water averaged 0.23 g m^{-3} over this layer. This is consistent with the presence of dendritic crystals, which require water saturated conditions.

3.1.3.3 Particle trajectories

In an attempt to understand the microphysical and electrical influences the convective line may exert on the stratiform region, dual-Doppler synthesized winds (at several analysis times) were used to compute ice particle trajectories. Hydrometeor fallspeeds were derived by averaging output from a series of extended velocity azimuth display analyses (EVAD, Srivastava et al., 1986) conducted on CIM data.

Using these fallspeeds, hydrometeor trajectories were computed by tracking the horizontal position of ice particles as they were ejected from the convective line at 8.5 km AGL (approximately 2 km above one of the primary charge transition levels in the electric field profiles). The results from this analysis are depicted, in a conceptualized manner, by Fig. 3.11, which shows the locations at which the ice particles would pass through the 0°C isotherm (denoted by the arrow head of each depicted trajectory) overlaid on a low-level reflectivity plot of CIM data at 0500 UTC. The line AA' through Fig. 3.11 delineates the location of a line-normal conceptual model, presented later.

As can be seen from Fig. 3.11, all hydrometeor trajectories exhibited a rather simple structure, essentially southeast to northwest (from 135°) in agreement with the mean flow exhibited by the dual-Doppler analysis for the 3.9-7.9 km layer (Figs. 3.4e-h). While this result is expected given the previously discussed kinematic structure, when compared against a similar

analysis for the asymmetric system, it reveals important differences that might offer some explanation for the different electrical structures and/or microphysical charging regimes.

3.1.4 MCS electrical structure

3.1.4.1 EFM observations

While P-3 kinematic and microphysical data were being collected, two NSSL mobile laboratories (hereafter referred to as NSSL1 and NSSL2) were deployed to launch balloon-borne EFM's into the advancing MCS. The primary objective of this mission was to collect multiple EFM profiles that would address both the temporal and spatial variability of MCS charge structure. Three electric field profiles were obtained in this system. Figs. 3.12 and 3.13 and Tables 3.1 and 3.2 depict the system-relative launch locations, electric field profiles, and analyzed charge densities (using the procedures described in Chapter 2) for the 0549 and 0705 EFM flights, respectively. Although these two soundings appear to have been launched at locations near the leading edge of the stratiform precipitation region, we hereafter refer to them as the symmetric MCS transition zone EFM profiles in order to distinguish them from the electric field profiles that were clearly obtained in a region of heavy stratiform precipitation. System-relative flight trajectories for these two balloon-borne EFM profiles, though obtained approximately 75 minutes apart, indicate that they passed through analogous regions of the MCS stratiform cloud. The system-relative launch location and electric field profile for the 0712 UTC EFM flight are presented in Fig. 3.14. Since horizontal electric fields in our 0712 UTC electric field profile are significant (discussed later), thereby violating the primary analysis assumption (i.e., that of horizontal stratification), we do not present a corresponding analysis of charge densities for that profile.

Analysis of electric fields to obtain charge densities (using Eqn. 2.3) is a subjective process. Depending on how one chooses the upper and lower boundaries in defining a charge layer, it is

possible to obtain charge densities that are quite different than those presented here. Previous studies (i.e., Marshall and Rust, 1993) have examined charge densities derived from these electric field profiles. Since the goal of this study is to understand potential relationships between MCS stratiform kinematic, microphysical, and electrical structures, we are most interested in examining the levels at which these charge transitions take place; thus, in cases where the actual level of charge transition is ambiguous because of either 1) the presence of high horizontal electric fields, or 2) an intervening region of charge wherein the electric field magnitude changes very little (signifying a near-zero charge density layer), we do not deduce charge densities. With these caveats in mind, an examination of Tables 3.1 and 3.2 suggests that stratiform charge densities in the 24 May 1991 MCS were on the order of 5 nC m^{-3} , a magnitude typical of convective cores (i.e., Marshall and Rust, 1993). We now examine, in more detail, the location at which the depicted profiles were collected and the possible relationship of the charge transition levels to the documented MCS kinematic and microphysical features.

The 0549 and 0705 UTC EFMs were both launched into the transition zone, albeit at locations that were relatively close to the stratiform region bright band (cf. Figs., 3.12, 3.13). Fig. 3.15 depicts line-normal cross-sections of radar reflectivity through the locations where the three EFMs were launched (indicated by the arrows). In a system-relative sense, both of the transition zone EFMs were initially carried towards the convective line in the low-level, rear-to-front flow (cf. Fig. 3.9). Upon ascending above approximately 2 km, however, they were quickly carried rearward over the most intense portion of the radar bright band by strong, front-to-rear flow (cf. Fig. 3.9). When comparing these trajectories to the composited vertical velocity structure (Figs. 3.5 through 3.8), it appears that both of these EFMs likely passed through the deepest and most intense portion of the stratiform updraft, which was located near the leading edge of the stratiform precipitation region. The 0712 UTC EFM, on the other hand, was

launched near the rear edge of the intense stratiform precipitation (Figs. 3.14 and 3.15). Given the much weaker system-relative, line-normal winds at this rearward location, however, the EFM appears to have always remained to the west of the most intense stratiform precipitation region. Furthermore, when comparing its trajectory to the composited vertical velocity structure (Figs. 3.5 through 3.8) it appears that the 0712 UTC EFM likely ascended through a much deeper region of mesoscale descent, than did the two transition zone EFM soundings.

Fig. 3.16 depicts the mean vertical component of the electric field from each profile. In comparing the EFM profiles (Fig. 3.16), it is interesting to note the marked similarities in the vertical charge structures. All three soundings indicated essentially five layers of charge located near and above the 0°C isotherm, though the 0712 UTC profile (dotted line in Fig. 3.16) shows considerably less agreement with the other two profiles at the lower levels (i.e., from 0 to -10°C). The similarities between the profiles are most evident when comparing the 0549 and 0705 UTC profiles (solid and dashed lines in Fig. 3.16). Though launched 76 minutes apart and at locations separated in distance by 51.9 km (but at approximately the same location with respect to the convective line), the profiles indicate a steady-state charge structure, at least over this observational period. In both cases, the dominant charge polarity below the 0°C isotherm was negative. A region of positive charge was situated above this region in both soundings. This near-freezing level charge transition is a rather common feature in stratiform region EFM soundings. Shepherd et al. (1996) recently examined charge transitions near the melting level of several MCSs and concluded that melting charging (i.e., Drake, 1968) was a likely candidate to explain charge layers in this region. It is interesting to note, however, that the positive charge layer in the 0549 UTC EFM sounding lies entirely *above* the 0°C isotherm. This is a feature that was also commonly found in the electric field profiles examined by Shepherd et al. (1996), leading them to speculate on several different mechanisms that might act in unison with the

melting mechanism to produce a sharp charge transition at temperatures immediately above the 0°C isotherm. The 0549 and 0705 UTC EFM soundings also show remarkable similarity in the middle levels, particularly at temperatures between -5 and -15°C.

As noted earlier, the 0712 UTC sounding, which was launched near the rear edge of the heavy stratiform precipitation, exhibited many significant differences at temperatures warmer than -10°C compared to the 0549 and 0705 UTC profiles. Yet there were many similarities at colder temperatures. Most notably, the close proximity of the 0712 UTC launch site to the rear edge of the radar bright band may have been responsible for the enhanced horizontal fields seen in the 0712 UTC profile, particularly at and just below the 0°C isotherm (see Fig. 3.14). Due to the presence of high horizontal electric fields in this sounding, we do not present derived charge densities for this electric field profile. Nevertheless, it is still intuitive to compare charge transition layers, as delineated by polarity changes in the electric field profile's slope, which appear to be collocated with similar transitions (particularly at approximately -12°C) in the two transition zone electric field profiles.

Additional insight into differences between the 0712 UTC profile and the 0549 and 0705 UTC profiles might be gained by re-examining the results from the Coulomb's Law model (presented in Chapter 2). Model sensitivity studies that tested the results of a hypothetical balloon ascent at a location that was either 1) near the edge of a layer of charge, or 2) just outside of a layer of charge, (cf. Fig. 2.7e and 2.7f) yielded results that bear marked similarities to the 0712 UTC profile. That is, vertical electric field transitions occurred essentially at the same levels, only with greatly diminished (approximately 30 kV m⁻¹, down from 100 kV m⁻¹ for a balloon that goes through a charge layer center) vertical electric field magnitudes and significantly greater (i.e., 50 kV m⁻¹) horizontal electric field magnitudes. Given the fact that the 0712 UTC electric field profile was obtained towards the edge of the stratiform precipitation

region and, additionally, since the vertical profile of the electric field profile bears marked similarities, at many levels, to the transition zone EFM profiles (only with much larger horizontal component), it is possible that the 0712 UTC balloon sampled the same regions of charge, only at a location that was far removed from the center.

Interestingly, the vertical charge structure observed in the transition zone and stratiform cloud of the 24 May 1991 MCS exhibits many marked similarities to the transition zone electric field profile presented by Schuur et al. (1991), and two stratiform electric field profiles presented by Hunter et al. (1992). Marshall and Rust (1993) examined these similarities, by shifting the various profiles upwards and downwards with respect to height, and determined that a common electrical structure exists in symmetric MCSs. We will present a similar analysis in this study, but will focus on the temperature level of charge transitions and a detailed comparison of those levels to MCS kinematic and microphysical structure. As will be seen when comparing these results to the profiles obtained within the 8 May 1991 MCS, which had a much simpler vertical charge structure, positive charge near the radar bright band level seems to be a fairly consistent feature in these soundings. The primary difference in the symmetric and asymmetric profiles, on the other hand, seems to lie in the stratiform cloud's upper regions, primarily at temperature levels between -5 and -15°C . These differences will also be addressed with respect to the system's respective kinematic and microphysical structures.

3.1.4.2 Lightning observations

We present here a brief overview of CG flash patterns on 24 May 1991 in order to provide a framework upon which the EFM observations might be better interpreted. As depicted by Fig. 3.17, which shows 30 minutes of lightning data overlaid on a low-level radar reflectivity composite at 0512 UTC (30 minutes before the first EFM launch), a distinct bipolar lightning distribution (e.g., Orville et al., 1988; Rutledge et al., 1988) existed during the time the EFM

data were being collected. In particular, the positive CG flashes associated with the trailing region of stratiform precipitation were preferentially located towards the leading edge of the stratiform precipitation region during this time period. A secondary maximum in positive CG flashes was located at the northern end of the system with a third associated with the development of a weak convective cell ahead of the line at a location ($X \approx -40, Y \approx -50$) km from CIM.

The evolution of the CG flash pattern over a 4 hour period leading up to the EFM data collection times (0549, 0705, and 0712 UTC), overlaid on low-level composites of radar reflectivity, is depicted in Fig. 3.18. These plots, which only show 15 minutes of lightning data centered on the radar analysis time, indicate a bipolar pattern from early on in the system's lifetime. In particular, CG lightning activity (both negative in the convective line and positive in the stratiform) increased dramatically by 0400 UTC, at which time the radar observations indicated the convective line had reached its most intense stage (cf. Fig. 3.1). Throughout this period positive/negative CG flash percentages were typically between 5 and 6%.

3.2 The asymmetric MCS of 8 May 1991

3.2.1 Environmental conditions and MCS development

At 1200 UTC 7 May 1991, a weak trough extended from a surface low over south-central New Mexico into eastern Colorado. Over the next 12 hours, the surface low deepened rapidly and moved eastward towards the Texas panhandle, resulting in an intensification of south-southeasterly winds over central Oklahoma and west Texas and a subsequent increase in dewpoints to 15°C (an increase by as much as 6°C over 12 hours earlier). Warm advection at 850 mb and the approach of a vigorous short-wave trough at 500 mb combined to destabilize the atmosphere over eastern New Mexico and the west Texas panhandle. As the surface low moved

eastward, convection was initiated along the corresponding dry line/trough by 2300 UTC in the central Texas panhandle. Eventually the convection became organized into a bowed line of convection similar to that described by Fujita (1978). From 1200 UTC 7 May 1991 to 1200 UTC 8 May 1991, the 500 mb trough underwent rapid deepening as it moved northeastward, apparently in response to the development of a mid-level mesoscale vortex over central Oklahoma (see Fig. 3 of Jorgensen and Smull, 1993).

Fig. 3.19 depicts the evolution of the radar reflectivity pattern of this MCS from 0200 through 0700 UTC (which encompasses the times of our aircraft and EFM observations). From early on (e.g., Fig. 3.19a), the MCS convection exhibited a pronounced “bowed” structure. As the line moved eastward (Figs. 3.19a through 3.19d), it expanded both to the north and south and eventually formed line that was nearly north-south oriented in central Oklahoma and east-west oriented in northern Texas (with the apex of the convective line’s bow tracking along the Red River). By 0500 UTC (Fig. 3.19d), the Texas portion of the convective line had plunged rapidly southeastward; the Oklahoma portion of the line, on the other hand, moved in a more easterly direction (11.2 m s^{-1} from 285°) and eventually merged with the remnants of a quasi-stationary region of disorganized convection that had persisted over north-central Oklahoma for several hours.

Though Figs. 3.19d through 3.19f provide a radar depiction that suggests an MCS with an “open-wave” structure (similar to that described by Blanchard, 1990 and Fortune et al., 1992), it should be noted that the actual location of the MCS’s mesovortex, at this time, was in west-central Oklahoma near the hook-like appendage in the radar reflectivity (see Figs. 3.19e and 3.19f), not at the intersection of the two convective lines. It is the precipitation pattern associated with the more westerly of these two precipitation regions that constitutes the asymmetric MCS of interest. By 0700 UTC (Fig. 3.19f), the approximate time of our first

stratiform EFM flight, the convective line over central Oklahoma had completely dissipated; the stratiform precipitation region (delineated by the intense radar bright band) exhibited radar reflectivities of > 40 dBZ, and had reached its maximum areal coverage.

3.2.2 MCS kinematic structure

3.2.2.1 P-3 flight track

At 0445 UTC, the P-3 took off from OKC, climbed to 750 mb (6°C), and proceeded westward towards the developing mesovortex. The flight track and levels for this MCS, which are depicted in Fig. 3.20, can be compared to Fig. 3.19 to determine the underlying radar reflectivity structure associated with each leg. Upon reaching a location to the west of the convective line, the P-3 made an abrupt turn towards the south-southeast and then conducted a 25 minute flight leg (from 0510 through 0535 UTC) to the rear of the convective line, during which time the MCS's kinematic structure was documented by fore-aft continuous scanning the P-3's tail-Doppler radar. After completing this first transition zone flight leg and returning to the north, the P-3 climbed to 510 mb (-12°C) and conducted a second south-southeasterly flight leg (from 0600 through 0630 UTC) to the rear of the convective line. These TZ/SR flight legs documented the MCS kinematic structure in a manner similar to that described above for the symmetric MCS.

At 0630 UTC, the P-3 completed the second flight leg and returned to the north where, from 0655 through 0725 UTC, it conducted an advecting spiral descent in the trailing stratiform region (near the vicinity of a balloon-borne EFM launch at 0659 UTC) that documented the MCS stratiform cloud's vertical microphysical structure at temperature levels between -20 and 15°C . After completing the spiral descent, the P-3 conducted a series of east-west flight legs

through the mesovortex region at a variety of flight levels. At 1100 UTC, the P-3 completed its final mesovortex flight leg and returned to OKC.

3.2.2.2 P-3 Doppler observations of horizontal flow structure

Unlike the symmetric MCS where a series of 10 TZ/SR flight legs were available for analysis, only two are available for this flight. We now present a detailed look at dual-Doppler synthesized wind fields from the second of these flight legs, which was conducted from 0600 through 0628 UTC (30 minutes prior to the first stratiform, balloon-borne EFM launch on 8 May 1991). An analysis of the mesovortex structure, which was sampled by the P-3 aircraft from 0730-1100 UTC, was discussed by Jorgensen and Smull (1993). Here we focus on the precipitation and kinematic structure to the rear of the convective line (as with the symmetric system) in an attempt to better understand the environment encountered by the balloon-borne EFM's. Using data from both flight legs, we then construct system-relative, line-normal reflectivity and vertical velocity composites using the procedures described previously in Section 3.1.2.3.

Fig. 3.21 depicts the low-level radar reflectivity pattern (composited from NSSL CIM radar data) at 0615 UTC, the location of the dual-Doppler grid box corresponding to the 0610-0628 UTC flight leg (see Fig. 3.20), and CG lightning strike locations in a 15 minute period centered on the dual-Doppler analysis time. The dual-Doppler, storm-relative winds and reflectivities at 8 levels are shown by Fig. 3.22. As can be seen from Fig. 3.22a, storm-relative winds at 0.9 km were from the east at between 5 to 15 m s⁻¹. Above this level (1.9 and 2.9 km) a noticeable transition in wind direction took place, eventually resulting in south-southeasterly (i.e., line-parallel) directed flow at locations immediately behind the convective line and towards the southern end of the domain. At the northern end of the line, winds remained predominantly

easterly in apparent response to the mesovortex that was now situated over west-central Oklahoma.

Also at the southern end of the analysis domain (but near the apex in the convective line's bow, cf. Fig. 3.19e), winds on the western edge of the analysis grid (at 1.9 and 2.9 km) exhibited a distinct, albeit weak ($< 5 \text{ m s}^{-1}$), westerly component. In a modeling study, Weisman (1993) found that the development of "bookend vortices" in bow-echo systems often served to focus rear-to-front flow near the apex of a bowed convective line. Though strong system-relative, rear-to-front flow was not observed at any level in this analysis, ground-relative westerly winds as high as 20 m s^{-1} were observed at 2.9 km in an analysis of the earlier flight leg (0509-0533 UTC). Nevertheless, high surface winds, which are typical with these types of systems (e.g., Johns and Hirt, 1987), were not reported anywhere in Oklahoma during the squall line's passage; the maximum westerly wind measured by the NSSL mobile laboratories was only 15 m s^{-1} .

At higher levels, from 3.9 through 7.9 km (Figs. 3.22d through 3.22h), the system-relative winds were approximately 20 m s^{-1} and from directions that ranged from 135° at 3.9 km to 155° at 7.9 km. These wind speeds and directions are similar to the upper-level winds documented for the symmetric system (cf. Figs. 3.4d through 3.4h). However, due to the more southeast-to-northwest orientation of the asymmetric MCS's convective line, the winds exhibited a much stronger line-parallel component than that in the symmetric MCS.

3.2.2.3 P-3 Doppler composites of vertical motion

Using the two TZ/SR flight legs from 0509-0533 UTC and 0600-0628 UTC, system-relative composites of vertical velocity were computed in the same manner as described earlier for the symmetric MCS. However, since the radar data in this system were collected 1) at a time during which the convective line was dissipating, and 2) at distances that were, in general, further from the convective line, these data do not provide as complete a documentation of the transition

zone's vertical velocity structure as was possible with the symmetric system (e.g., Figs. 3.5 through 3.8). Given these limitations, we focus on a space-time composite constructed with respect to the leading edge of the stratiform precipitation region.

Fig. 3.23 depicts the line-averaged reflectivity and vertical motion profiles with respect to the leading edge of the trailing stratiform precipitation region. The 15 km gap in the vertical velocity data in Fig. 3.23 represents data that was removed from the analysis due to poor quality wind estimates at high radar pointing angles (described earlier). Given the much broader coverage of the symmetric MCS observations, these data gaps did not appear in earlier analyses. Nevertheless, as revealed by an examination of Fig. 3.23, the consistency of the vertical velocity observations on either edge of the data gap suggests that the same basic structure was also present in the data void region.

The most striking feature in the vertical velocity composite is the depth and strength of the mesoscale downdraft at levels between 0 and -15°C . It might be recalled that the analysis of the symmetric MCS consistently revealed a rather deep updraft near the leading edge of the stratiform bright band region. In this system, however, a mesoscale *downdraft* appears to be a more common feature within this layer (with averaged magnitudes as high as 30 cm s^{-1}). Below the radar bright band, mesoscale descent was as large as 60 cm s^{-1} . The transition zone vertical velocity structure generally consisted of weak upward motion near cloud top with a deep region of weak descent below. While there are no indications of a mechanically forced, upper-level downdraft in this analysis, these data were collected at a time during which the convective line was dissipating. Hence it is possible that such a vertical velocity structure existed earlier when the line was more intense. At the right hand side of the composite grid, upward vertical motion from the dissipating convective line is significantly less intense, and centered at a much lower level, than was the case in the symmetric MCS.

3.2.3 MCS microphysical structure

3.2.3.1 Optical array probe data

In order to compare the microphysical structure of the symmetric MCS to that of the asymmetric MCS, we now present optical array probe data from the advecting spiral descent that was conducted from approximately 0655-0725 UTC on 8 May 1991. Again, OAP instrumentation problems resulted in significant data gaps. Nevertheless, we present a sample of the data that were available from this advecting spiral descent and, in particular, focus on observations made at -8.5°C in order to conduct a comparison to the same temperature level of the symmetric MCS. Recall that this temperature level, in the symmetric MCS, was located immediately below one of the primary charge transitions in the symmetric MCS electric field meter profiles. In the asymmetric MCS, this temperature level was located approximately in the middle of a deep region of net negative polarity charge (presented later) wherein no significant charge transitions were observed. Therefore, it is hoped that any identifiable differences in these two microphysical profiles might lend additional insight into possible electrification differences.

The shadow graph imagery at -8.5°C (0706 UTC) are presented in Fig. 3.24. The most striking difference between this data and that presented earlier (for the symmetric MCS) is the lack of dendritic growth at -8.5°C . Whereas the shadow graph imagery from the symmetric MCS spiral descent indicated the presence of dendrites with sizes as large as 3 mm, these data do not indicate the presence of any dendritic growth. Rather, most hydrometeors in the shadow graph imagery are small, irregularly-shaped ice particles; others appear to be fragments and needles. Some images in the 2D-C data appeared to be small (i.e., < 0.5 mm) graupel, though it is unlikely that significant riming could have occurred locally given the weak stratiform updrafts in this region. The lack of dendritic growth in this region is suggestive of ice saturated conditions, not water saturated, throughout a deep layer above the -8.5°C level. This speculation

is supported by the thermodynamic soundings on 8 May 1991, which both indicated slightly subsaturated conditions with respect to water through a deep layer.

3.2.3.2 Cloud water

In general, LWCs measured in the asymmetric MCS were noticeably smaller than indicated in the symmetric MCS. Though occasional pockets of $LWC > 0.2 \text{ g m}^{-3}$ were found, several flight legs showed no cloud water at all (or LWCs that were below the minimum detectable threshold of the instrument). In order to compare the LWCs in the vicinity of the EFM observations with those found in the symmetric MCS, LWCs measured by the P-3 during its advecting spiral descent were averaged over the -8 to -16°C layer. This is the layer immediately above the level wherein we compare the shadow graph imagery of the two MCS and also where a significant charge transition was found in the symmetric MCS EFM profiles. Averaged LWCs over this layer indicated the presence of at least some cloud water, although with contents that were approximately half of that observed in the symmetric system. Nevertheless, the presence of some cloud water in this region suggests that water saturated conditions were encountered over at least a portion of the spiral descent, though those observations may have been made as far as 20 to 30 km from the location of the balloon-borne EFM. In any case, ice particle habits do not support the widespread presence of water saturated conditions.

3.2.3.3 Particle trajectories

Dual-Doppler synthesized winds constructed from several P-3 flight legs were combined with ground-based dual-Doppler radar coverage by NSSL's CIM and NRO radars to construct trajectories for ice particles detrained from the convective line at 8.5 km AGL (i.e., as was done with the symmetric MCS). In this analysis, hydrometeor fallspeed profiles similar to those used in the symmetric MCS analysis were used. The results from this analysis are depicted, in a

conceptualized manner, by Fig. 3.25, which shows the locations at which the ice particles would pass through the 0°C isotherm overlaid on a low-level reflectivity plot of CIM data at 0615 UTC. The line BB' through Fig. 3.25 delineates the location of a line-normal conceptual model, to be discussed in Chapter 4. Unlike the trajectories indicated for the symmetric MCS (Fig. 3.11), ice particle trajectories computed for the asymmetric MCS exhibited a stronger line-parallel component. When composite vertical velocity structures are compared for the symmetric and asymmetric MCSs (Figs. 3.6 and 3.23, respectively), along with differences in line-parallel flow, ice particles detrained from the convective line of the asymmetric system had as much as three times more "residence time" in the transition zone downdraft than in the symmetric system. The implications of these differences will be discussed in more detail later.

3.2.4 MCS electrical structure

3.2.4.1 EFM observations

On 8 May 1991, the NSSL1 mobile laboratory was deployed to a location to the southwest of NSSL to launch a series of balloon-borne EFM's into the advancing, asymmetric MCS. Two stratiform region electric field meter profiles were obtained in this system. Figs. 3.26 and 3.27 and Tables 3.3 and 3.4 depict the system-relative launch locations, electric field profiles, and analyzed charge densities (using the procedures described in Chapter 2) for the 0659 and 0755 EFM flights, respectively. The first EFM (0659 UTC) was launched at a location toward the leading edge of the stratiform precipitation region. Caught up in the dominant southeasterly, system-relative flow (see Fig. 3.22), the EFM was quickly carried to the northwest and eventually passed through the most intense region of the stratiform region bright band. The second EFM (0755 UTC) was also launched in the stratiform precipitation region, though the intensity and areal coverage of the bright band had decreased dramatically by this time. As noted

earlier, both of these EFM profiles were obtained at a time during which the MCS was rapidly dissipating, in contrast to the profiles obtained in the symmetric MCS, which were collected when the convective line was just beginning to enter its dissipative stage. The possible implications of a dynamically active convective line on the concurrent MCS stratiform region kinematic, microphysical, and electrical structures will be addressed in Chapter 4.

Like with the symmetric MCS, the profiles collected in the asymmetric MCS exhibited marked similarities to each other. Both soundings had essentially three layers of charge located near and above the 0°C isotherm. The most striking difference in the electrical structures between the symmetric and asymmetric systems appears to be the presence of an additional two layers of charge in the -5 to -15°C layer of the symmetric system. Like the symmetric MCS EFM soundings, however, both asymmetric MCS electric field profiles exhibited layers of significant positive charge at the 0°C isotherm. Charge densities analyzed within this layer were between approximately 1.5 and 2.5 nC m⁻³ and electric field magnitudes were typically between 60 and 90 kV m⁻¹. Above this intense lower layer, however, charge densities were quite weak (< -0.5 nC m⁻³). Additionally, both asymmetric MCS electric field profiles exhibited rather deep (~1 km) layers immediately above the lower region of positive charge over which horizontal electric fields were significant and, as a result, charge densities were not computed. This is in sharp contrast to the mid-levels of the symmetric MCS stratiform cloud, where horizontal electric fields were typically much smaller (< 30 kV m⁻¹ as determined using Eqn. 2.1).

When comparing the asymmetric MCS EFM profiles to the composite kinematic structure, it is apparent that the two EFMs presented here likely ascended through a much deeper region of mesoscale descent (cf. Fig. 3.23) than did the two transition zone soundings in the symmetric MCS, which appear to have ascended through the deepest and most intense region of the stratiform updraft. Additionally, the cloud top heights in the asymmetric MCS electric field

profiles are significantly lower (as delineated by where the vertical component of the electric field goes to nearly zero) than those of the symmetric MCS. This is likely due to the fact that the MCS was dissipating at the time these electric field data were collected. The possible relationships between mesoscale updraft structure, MCS microphysical development, and levels of electric field polarity transitions will be addressed in Chapter 4.

3.2.4.2 Lightning observations

The evolution of the CG flash pattern of the asymmetric system over a 4 hour period leading up to the EFM data collection times (0659 and 0755 UTC), overlaid on low-level composites of radar reflectivity, is depicted in Fig. 3.28. These plots, which show 15 minutes of lightning data centered on the radar analysis time, can be compared to Fig. 3.18 to reveal differences in the lightning structures between the symmetric and asymmetric systems. As noted earlier, the convective line was dissipating rapidly at the time the EFM observations were made in the asymmetric system. This evolution is also reflected well in the lightning data as a significant decrease in negative CG lightning activity occurs between 0500 and 0600 UTC (by which time the convective line had all but disappeared). During this 4 hour period, positive/negative CG flash ratios were as high as 40%, which is significantly higher than that exhibited by the symmetric MCS. Additionally, the positive CG flashes associated with the asymmetric MCS exhibited much higher (by almost a factor of 2) peak currents. Petersen and Rutledge (1992), in a study of tropical MCSs over northern Australia, found that positive peak currents of CG flashes in the stratiform region were much larger than those from the convective line. They argued that this indicated that the CG flashes in MCS stratiform regions were likely tapping a large reservoir of charge, afforded by horizontally extensive stratiform regions.

Table 3.1: Charge analysis of the 0549 UTC, 24 May 1991 (symmetric) MCS electric field profile.

Z_{bot} (km)	Z_{top} (km)	T_{bot} (°C)	T_{top} (°C)	E_{bot} (kV/m)	E_{top} (kV/m)	ρ (nC/m ³)
2.440	4.476	7.5	-1.3	11.56	-81.64	-0.40
4.476	4.750	-1.3	-1.9	-81.64	72.96	4.99
5.223	5.473	-4.3	-5.8	72.36	-70.34	-5.05
6.048	6.558	-8.8	-11.8	-68.62	63.73	2.30
6.558	9.220	-11.8	-28.5	63.73	-27.13	-0.30
9.220	10.949	-28.5	-43.1	-27.13	-0.80	0.13

Table 3.2: Charge analysis of the 0705 UTC, 24 May 1991 (symmetric) MCS electric field profile.

Z_{bot} (km)	Z_{top} (km)	T_{bot} (°C)	T_{top} (°C)	E_{bot} (kV/m)	E_{top} (kV/m)	ρ (nC/m ³)
2.486	4.062	10.6	0.5	21.49	-104.68	-1.37*
4.062	4.562	0.5	-1.3	-104.68	94.63	3.53
4.562	5.396	-1.3	-6.2	94.63	-69.44	-1.74
5.396	6.364	-6.2	-11.6	-69.44	63.30	1.21
6.364	7.846	-11.6	-21.0	63.30	-15.47	-0.47
7.846	8.571	-21.0	-26.1	-15.47	2.49	0.22

* Representative charge density above and below high horizontal electric field region at approximately 5°C.

Table 3.3: Charge analysis of the 0659 UTC, 8 May 1991 (asymmetric) MCS electric field profile.

Z_{bot} (km)	Z_{top} (km)	T_{bot} (°C)	T_{top} (°C)	E_{bot} (kV/m)	E_{top} (kV/m)	ρ (nC/m ³)
1.737	3.500	10.6	0.6	4.10	-76.99	-0.41
3.500	4.022	0.6	-1.3	-76.99	66.11	2.43
5.079	6.346	-7.2	-15.2	47.03	-23.03	-0.49
6.346	7.791	-15.2	-25.1	-23.24	-3.03	0.12

Table 3.4: Charge analysis of the 0755 UTC, 8 May 1991 (asymmetric) MCS electric field profile.

Z_{bot} (km)	Z_{top} (km)	T_{bot} (°C)	T_{top} (°C)	E_{bot} (kV/m)	E_{top} (kV/m)	ρ (nC/m ³)
0.504	3.212	12.0	2.3	12.72	-60.63	-0.24
3.212	3.935	2.3	-1.2	-60.63	60.75	1.49
4.757	5.874	-6.0	-12.3	55.27	-10.00	-0.52

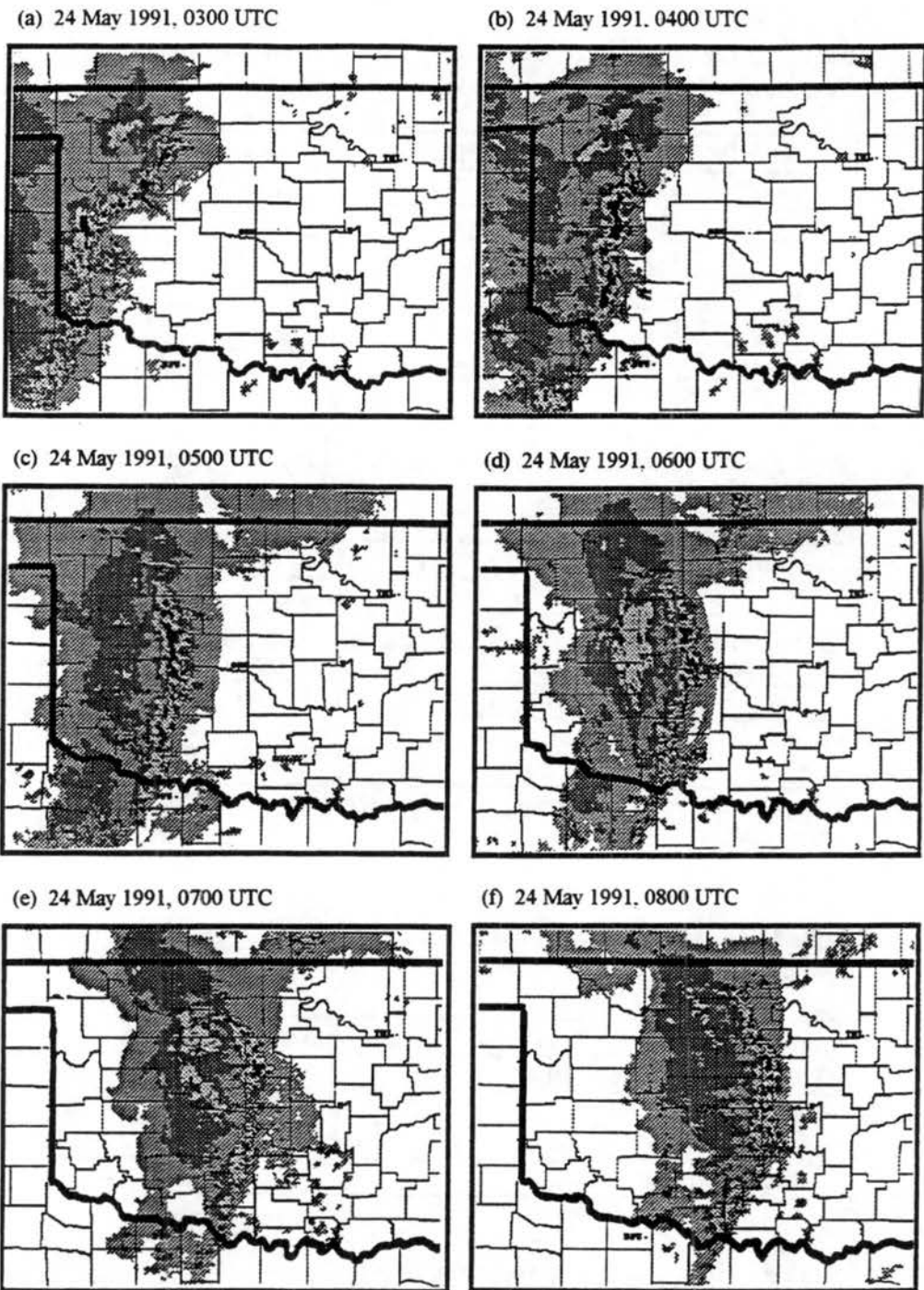


Fig. 3.1. Composite low-level radar reflectivity patterns of the 24 May 1991 (symmetric) MCS at (a) 0300, (b) 0400, (c) 0500, (d) 0600, (e) 0700, and (f) 0800 UTC. The four shading intensities indicate radar reflectivity ranges of 18-30 dBZ, 30-41 dBZ, 41-46 dBZ, and > 46 dBZ, respectively. Radar data from WSI.

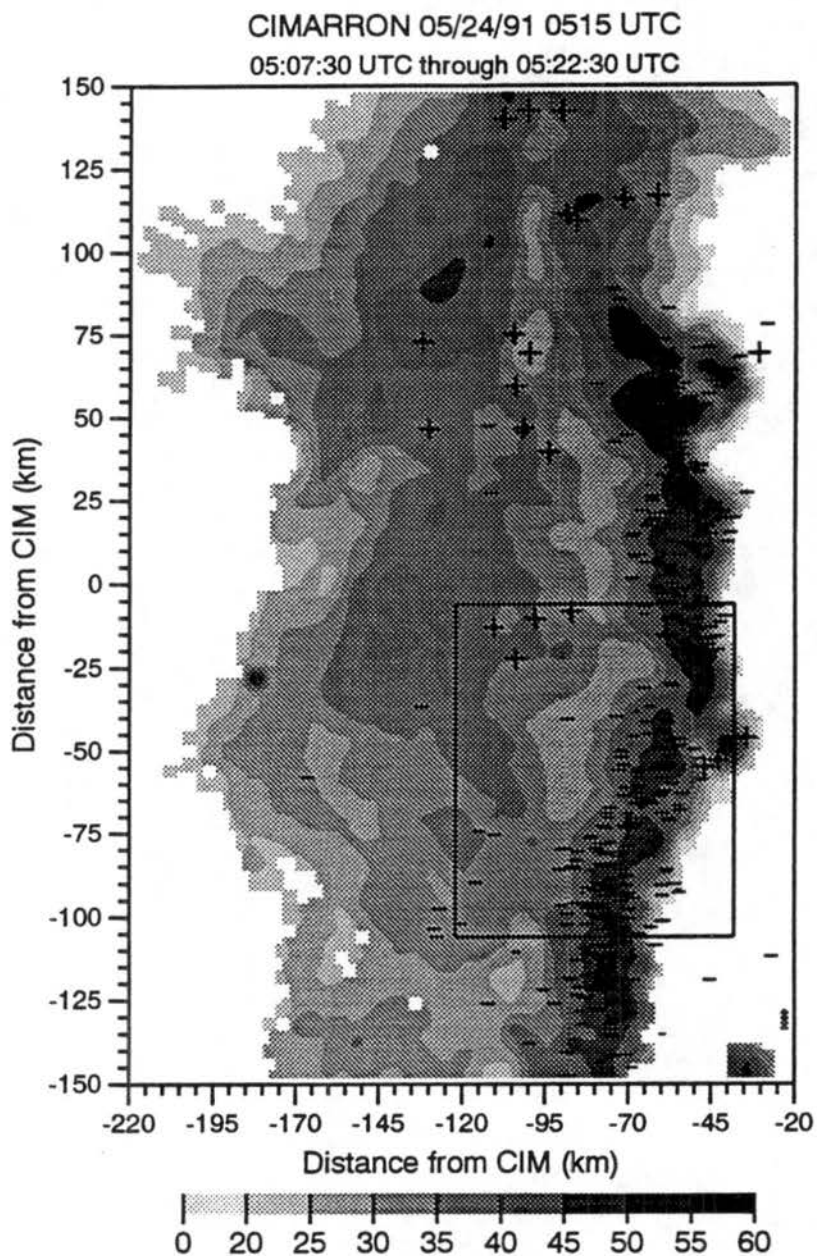


Fig. 3.3. Low-level radar reflectivity pattern (from CIM) at 0515 UTC, 24 May 1991 and CG lightning strike locations from 050730-052230 UTC, 24 May 1991. The grid box overlaid on the low-level radar reflectivity pattern depicts the system-relative location of dual-Doppler analysis presented in Fig. 3.4.

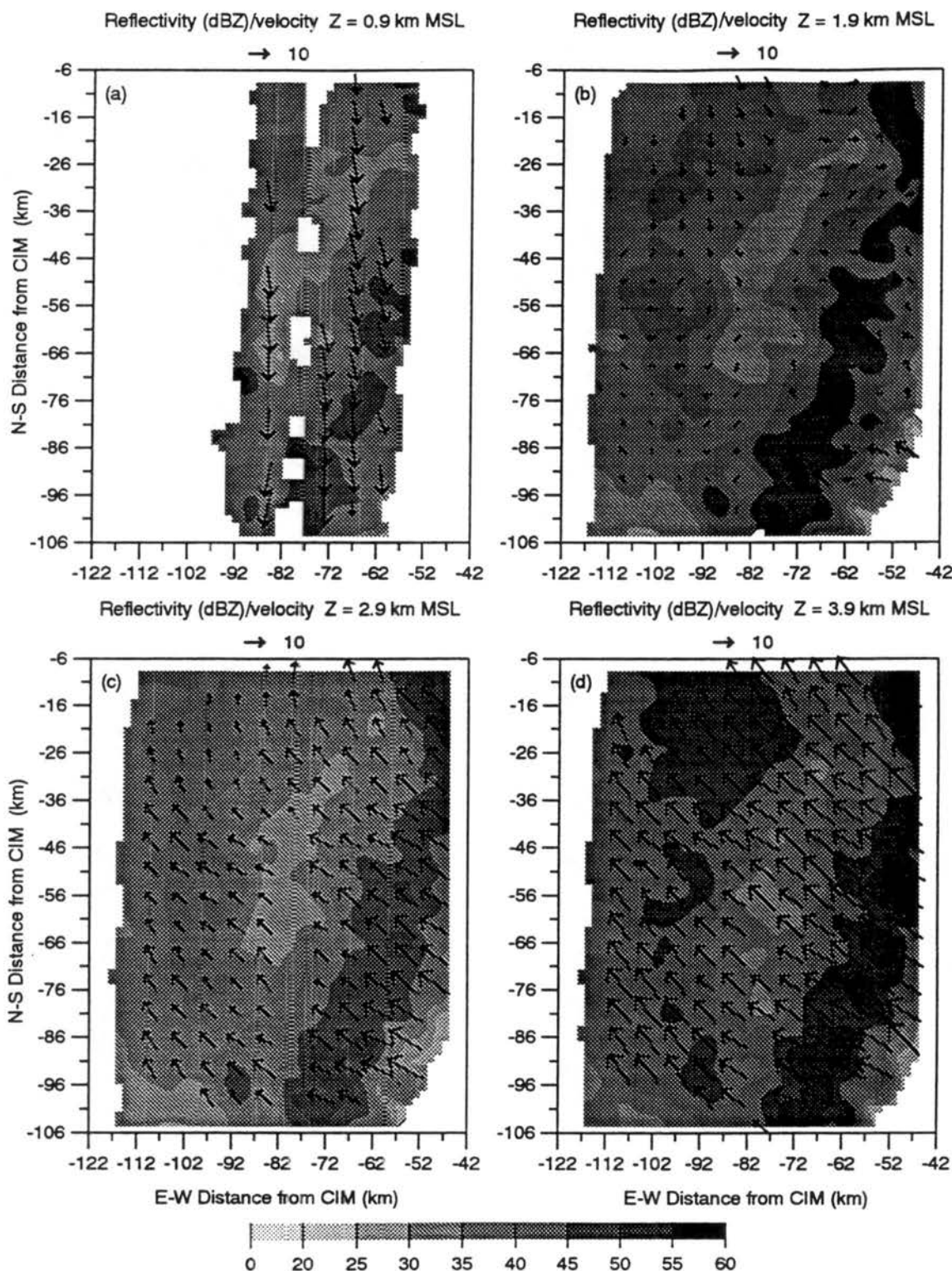


Fig. 3.4. Dual-Doppler radar analysis at (a) 0.9, (b) 1.9, (c) 2.9, (d) 3.9, (e) 4.9, (f) 5.9, (g) 6.9, and (h) 7.9 km MSL. Analysis is constructed from 0512-0520 UTC, 24 May 1991 P-3 flight leg (cf. Figs. 3.2a, 3.3). Vectors (plotted every third grid point) represent system-relative horizontal flow in m s^{-1} . The scale vector at the top of each plot represents a wind speed of 10 m s^{-1} .

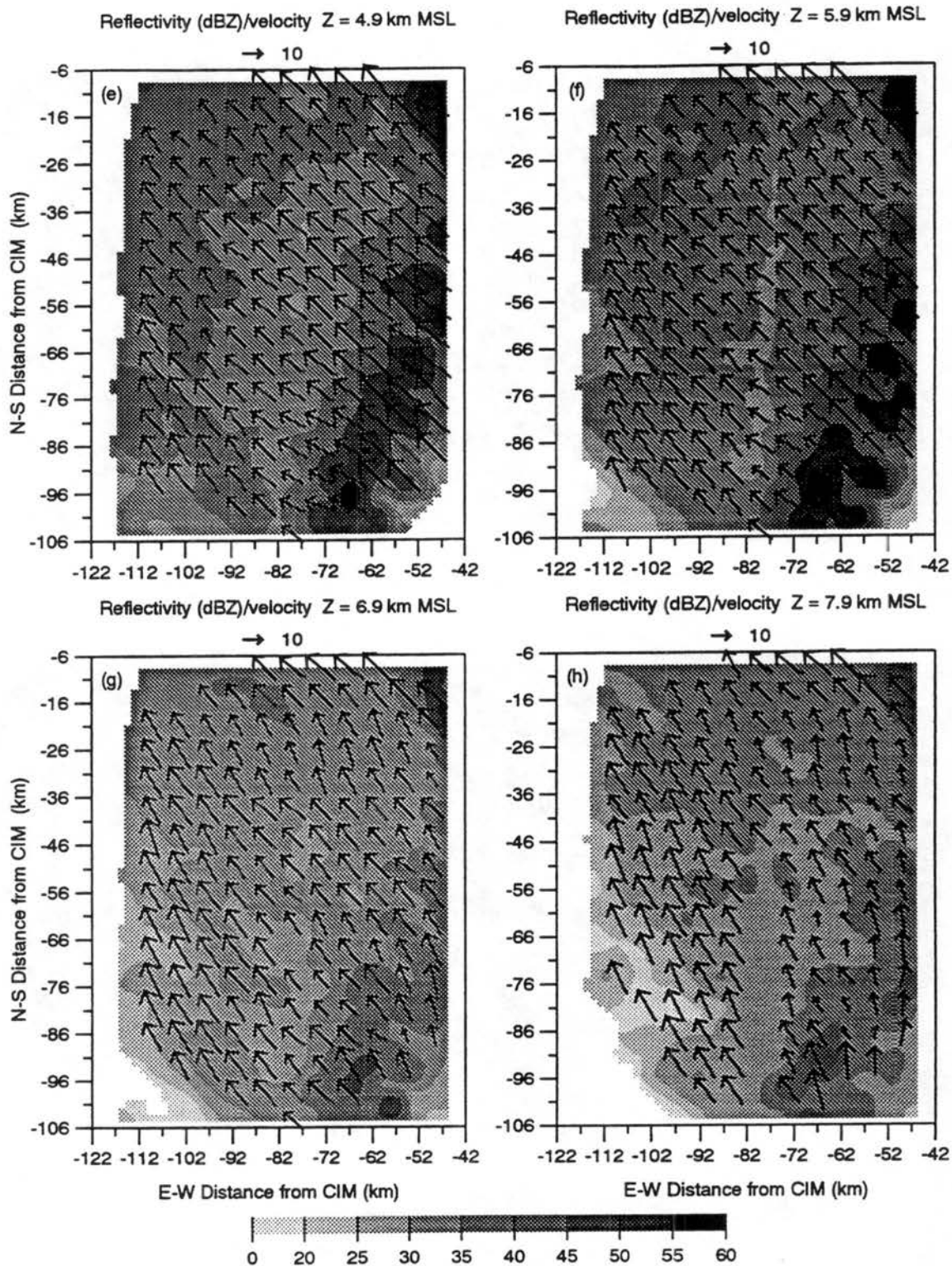


Fig. 3.4. Continued.

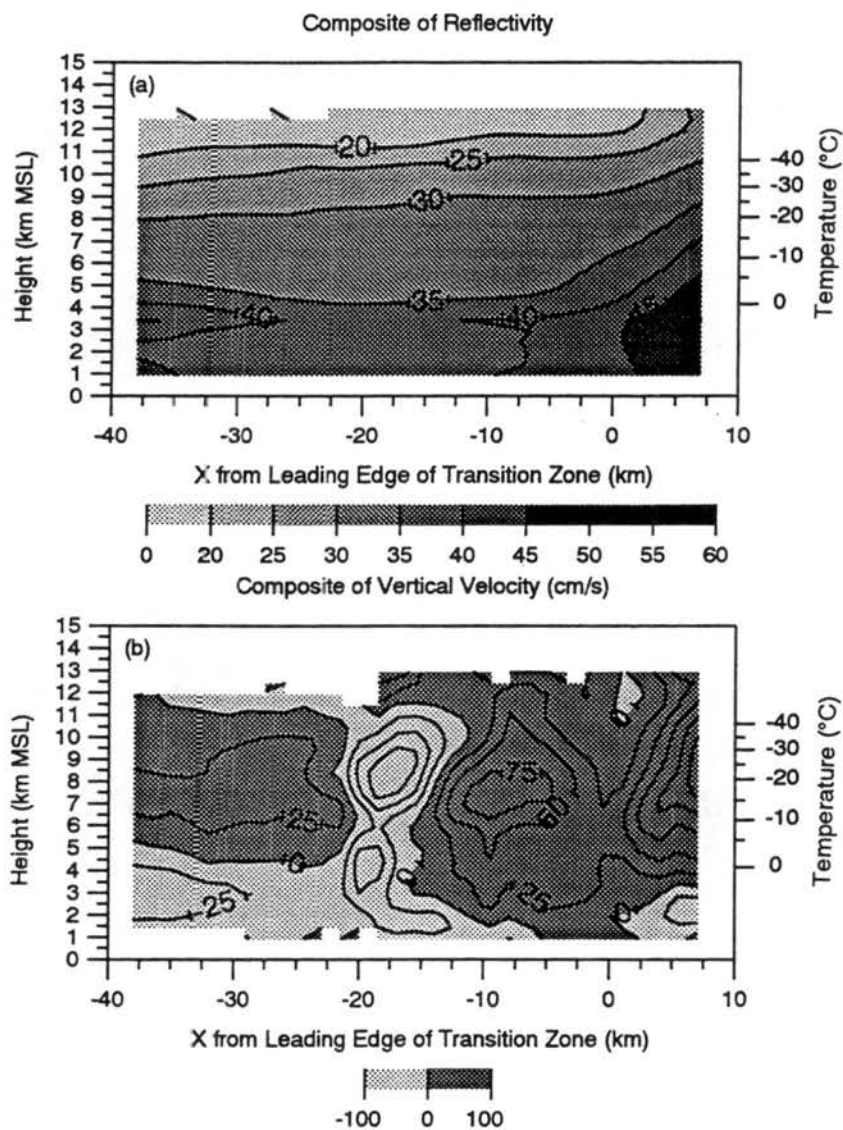


Fig. 3.5. Composite line-normal profile of (a) radar reflectivity (dBZ), and (b) vertical velocity (cm s^{-1}) constructed from the 0408 UTC, 24 May 1991 P-3 flight leg. In (b), dark shading indicates updraft while light shading indicates downdraft, with vertical velocity contoured every 25 cm s^{-1} . Plotted with respect to the leading edge of the transition zone.

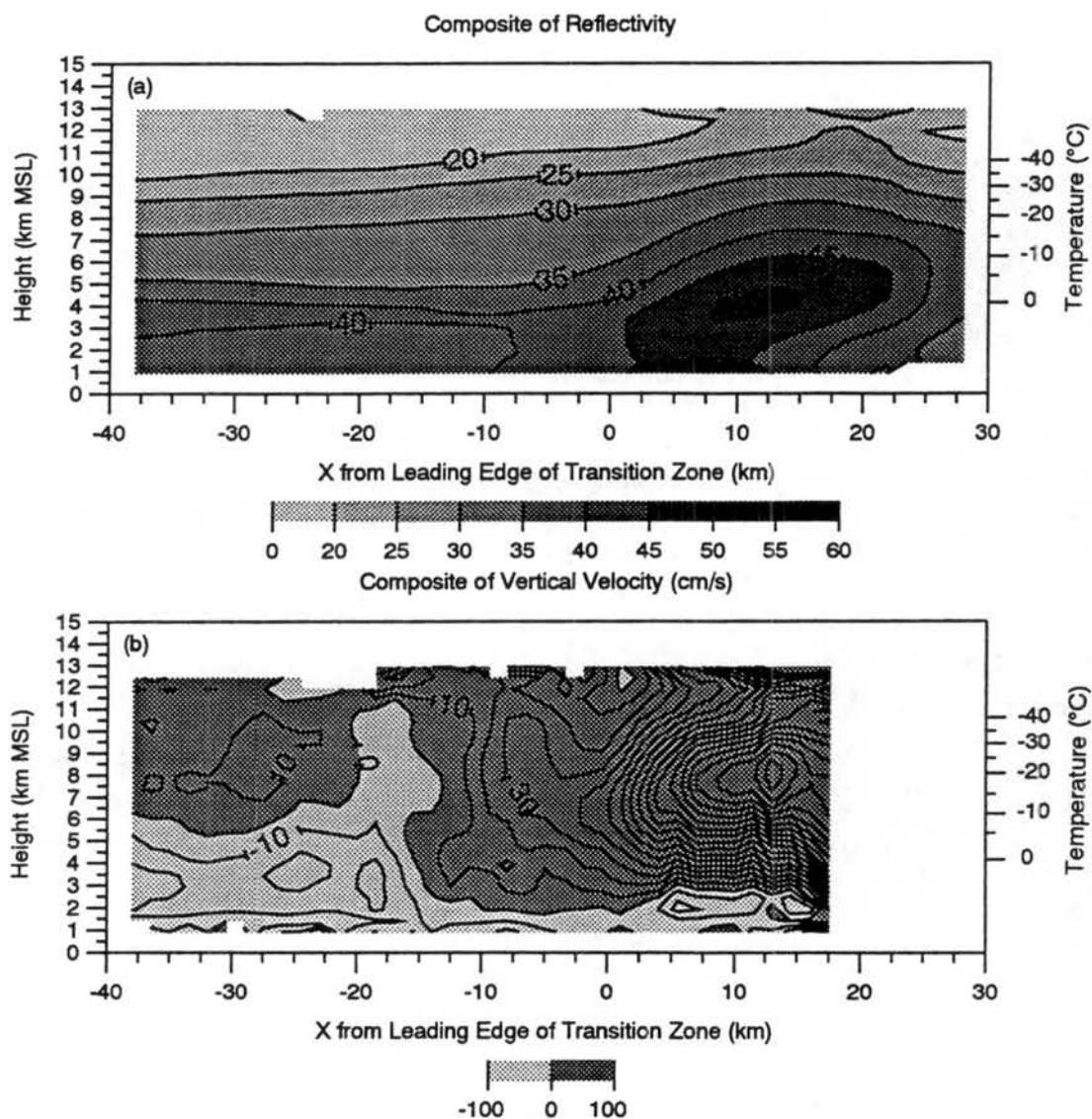


Fig. 3.6. Composite line-normal profile of (a) radar reflectivity (dBZ), and (b) vertical velocity (cm s^{-1}) constructed from all 24 May 1991 P-3 transition zone/stratiform region flight legs. In (b), dark shading indicates updraft while light shading indicates downdraft, with vertical velocity contoured every 10 cm s^{-1} . Plotted with respect to the leading edge of the transition zone.

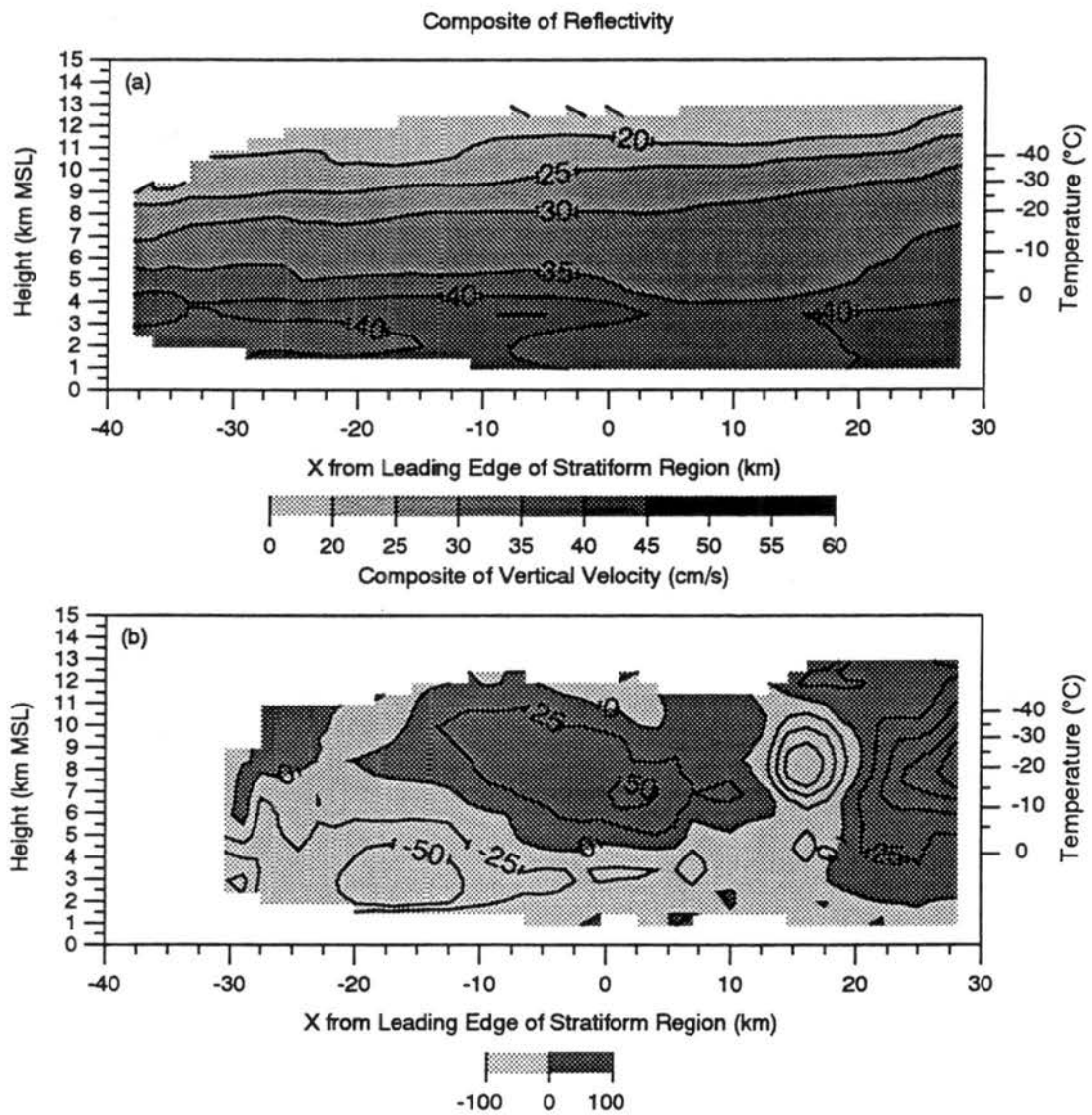


Fig. 3.7. Same as Fig. 3.5 except plotted with respect to the leading edge of the stratiform precipitation region.

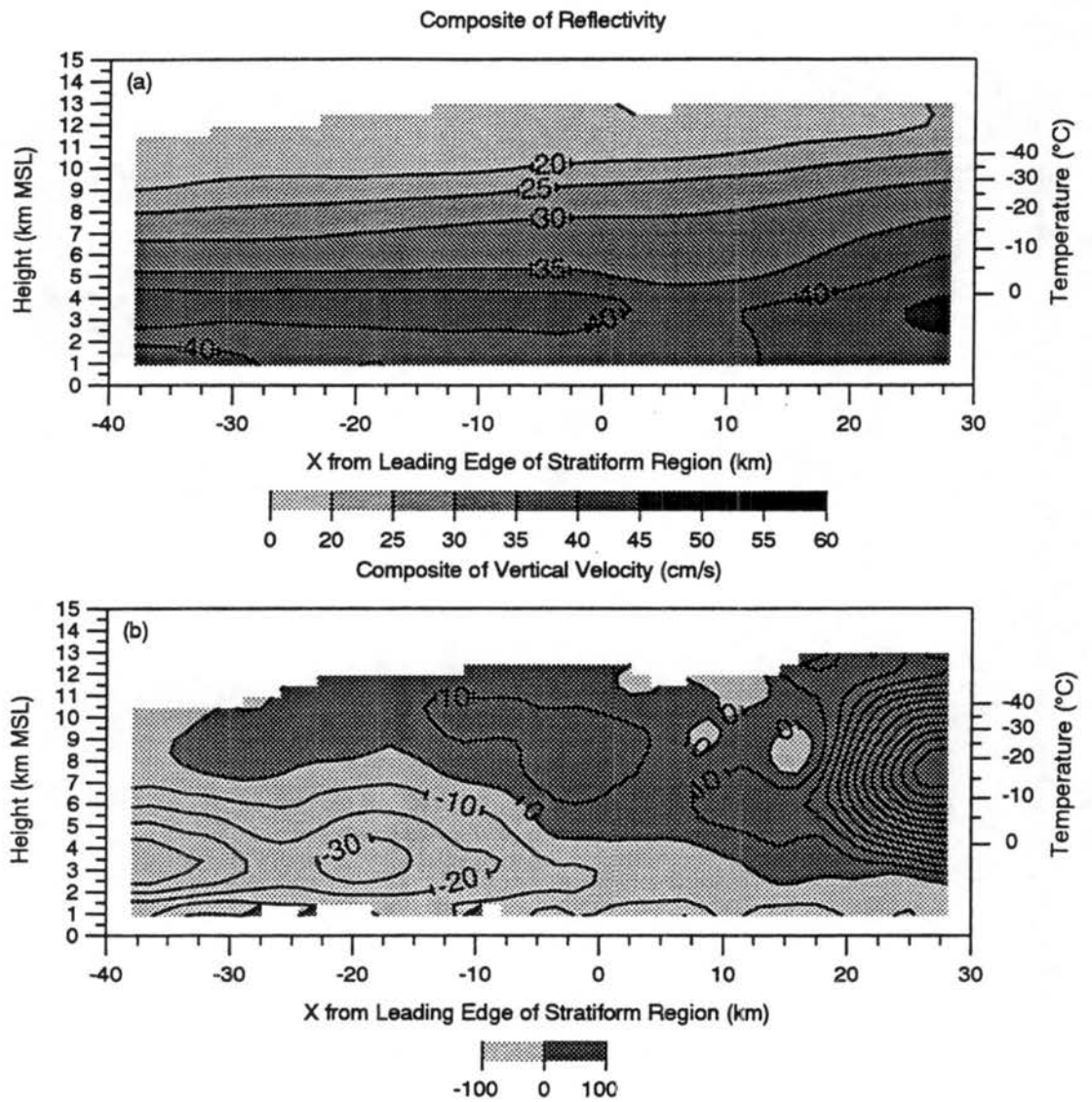


Fig. 3.8. Same as Fig. 3.6 except plotted with respect to the leading edge of the stratiform precipitation region.

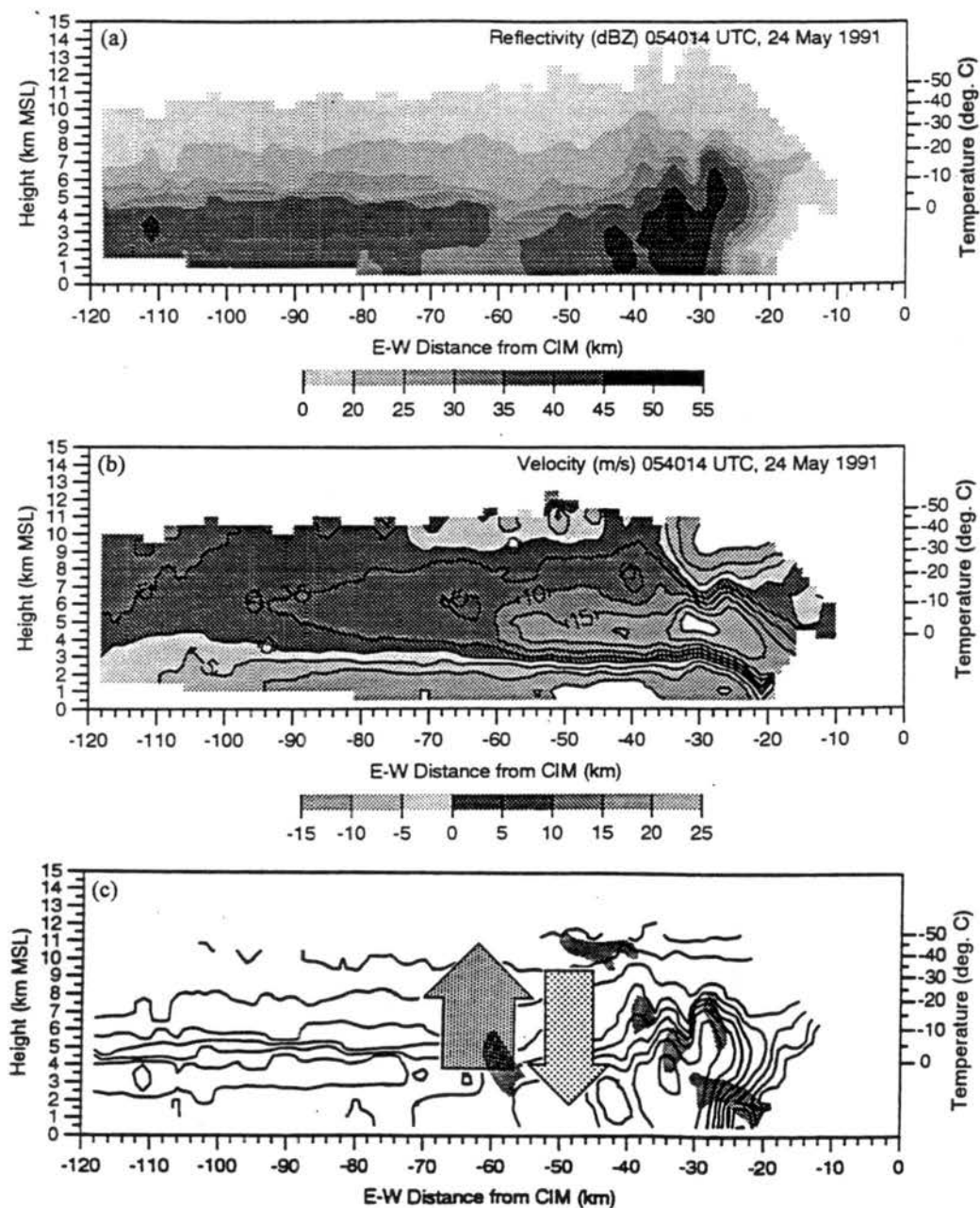


Fig. 3.9. Single-Doppler vertical cross-section of 24 May 1991 (symmetric) MCS (a) reflectivity (dBZ), and (b) system-relative radial velocity (m s^{-1}) constructed along a 270° radial (i.e., normal to the convective line) at 0540 UTC. Panel (c) presents a conceptualization of the mesoscale vertical velocity profile (arrows) overlaid on reflectivity contours that correspond to the shaded regions in part (a). Light shading in (c) delineates regions where two-dimensional (i.e., line-normal) convergence exceeded $1 \times 10^{-3} \text{ s}^{-1}$.

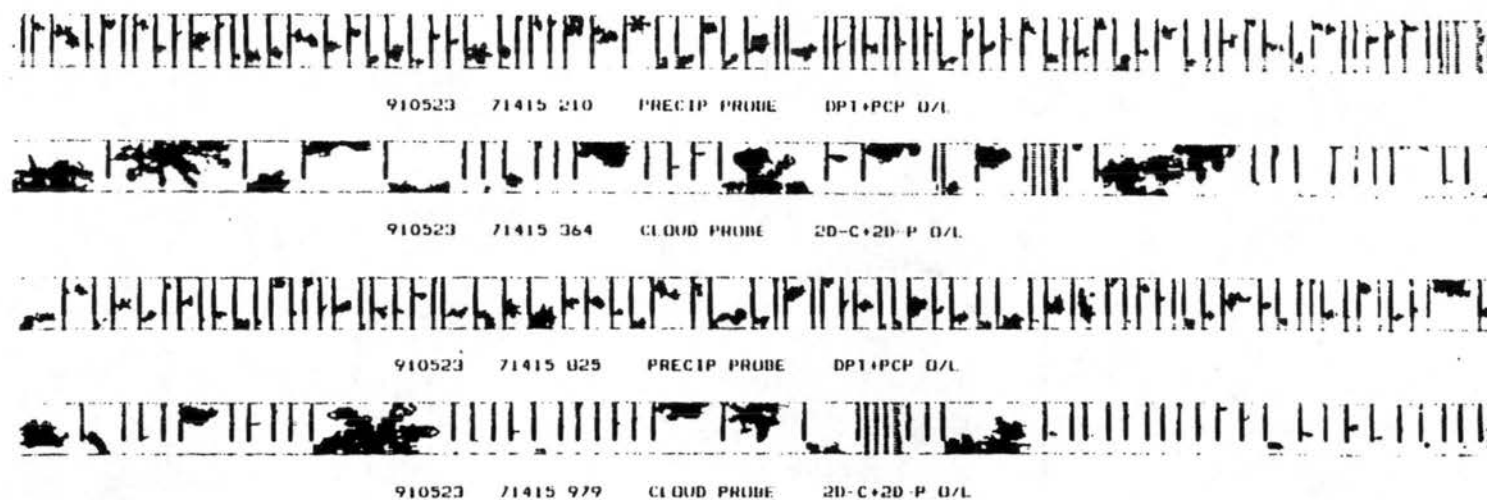


Fig. 3.10. NOAA P-3 shadow graph imagery obtained at a temperature of -8.5°C in 24 May 1991 (symmetric) MCS.

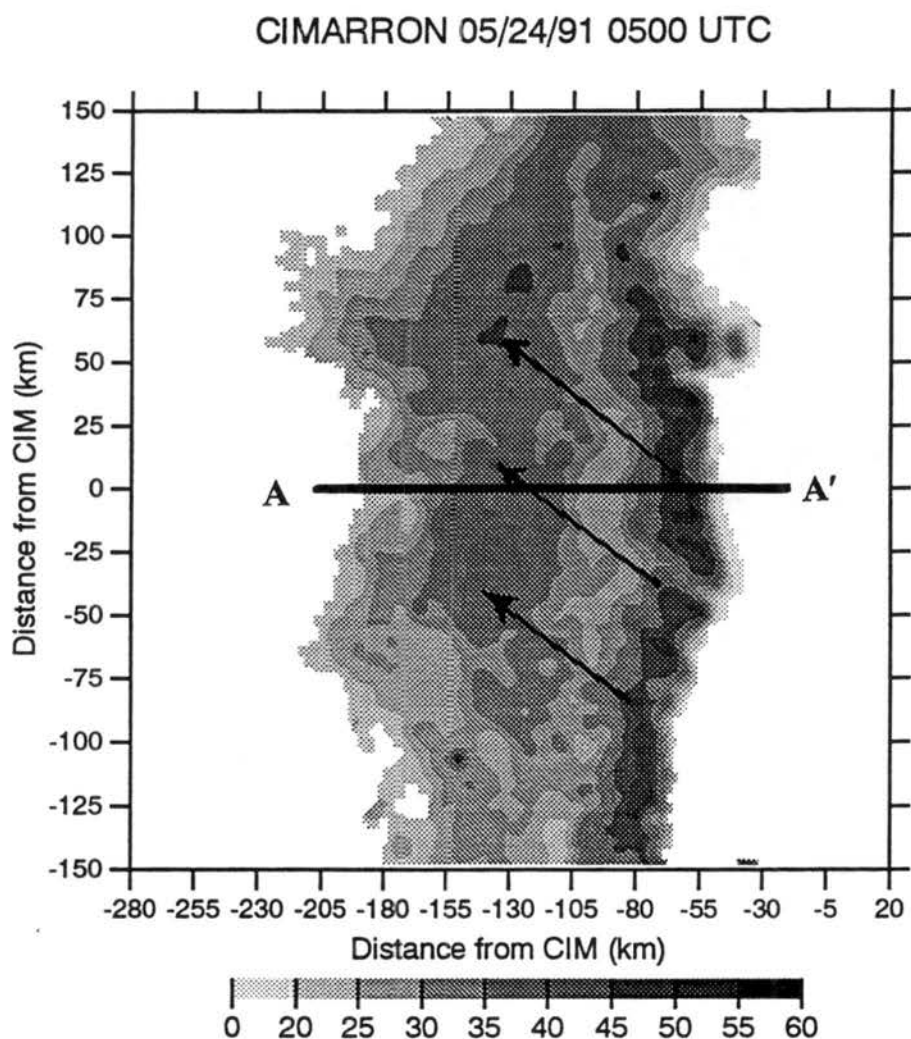


Fig. 3.11. Schematic of ice particle trajectories in 24 May 1991 (symmetric) MCS overlaid on low-level (CIM) radar reflectivity at 0500 UTC. Arrows indicate approximate path taken by ice particles ejected from convective line at 8.5 km AGL to the location where they reach the melting level. Line AA' represents approximate location of conceptual model presented in Fig. 4.5.

(a) 24 May 1991, 0549 UTC

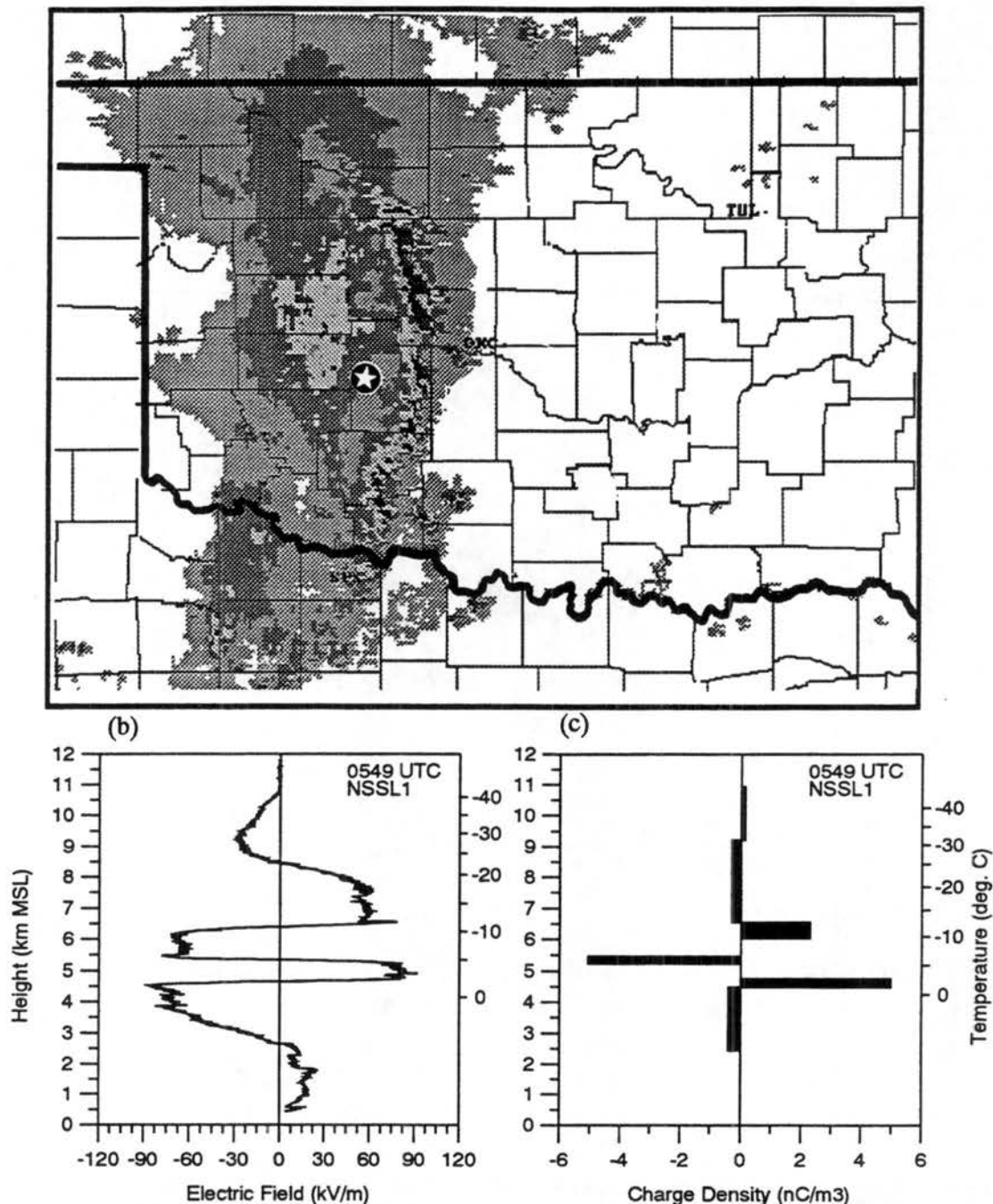


Fig. 3.12. (a) Location of 0549 UTC 24 May 1991 electric field meter launch (indicated by star) overlaid on composite of low-level radar reflectivity patterns of the 24 May 1991 (symmetric) MCS at 0545 UTC. (b) Electric field profile from the 0549 UTC electric field meter flight. (c) Charge densities (nC m^{-3}) derived from the 0549 UTC electric field meter profile using Gauss' Law (Eqn. 2.3). The four shading intensities in (a) indicate radar reflectivity ranges of 18-30 dBZ, 30-41 dBZ, 41-46 dBZ, and > 46 dBZ, respectively. Radar data from WSI.

(a) 24 May 1991, 0705 UTC

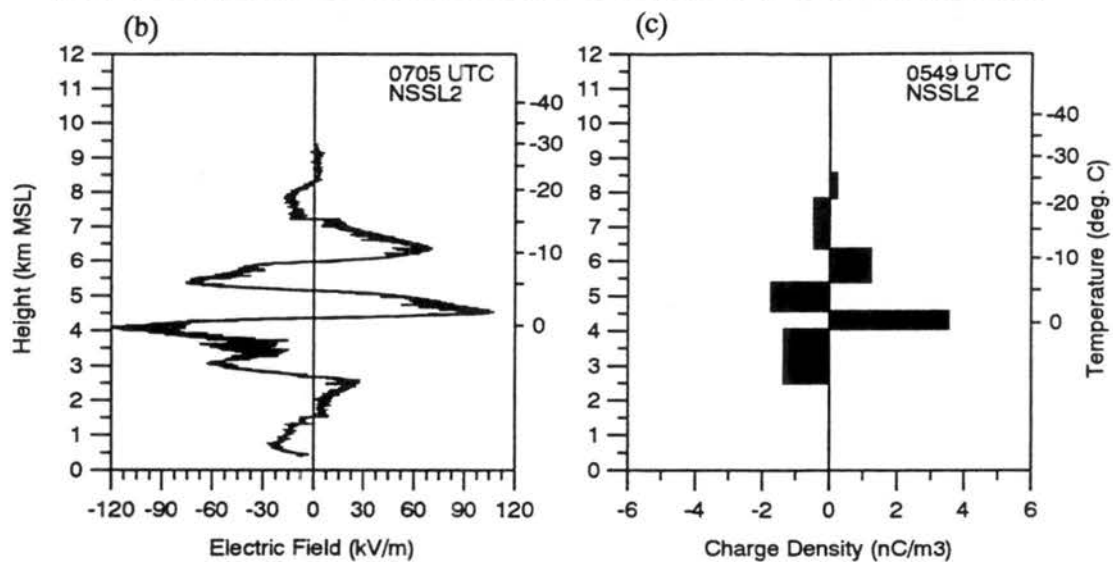
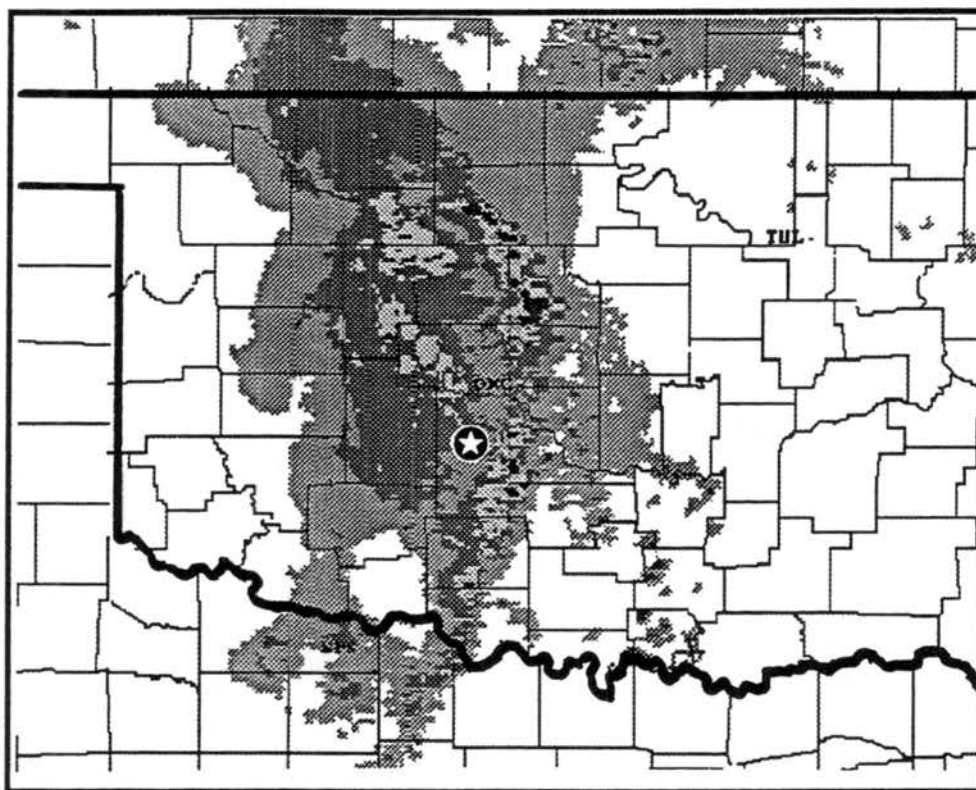


Fig. 3.13. Same as Fig. 3.12 except for 0705 UTC 24 May 1991 electric field meter launch/flight.

(a) 24 May 1991, 0712 UTC

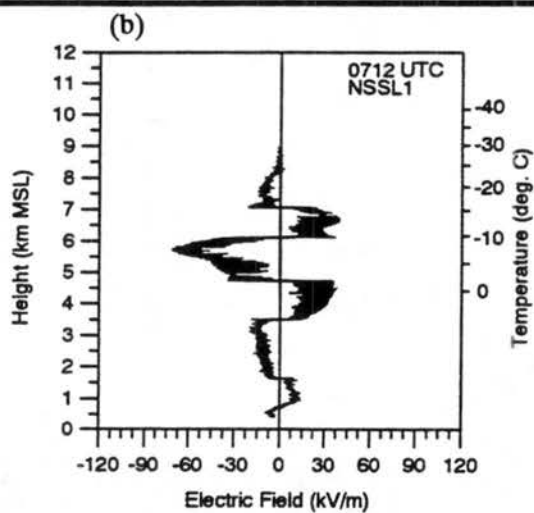
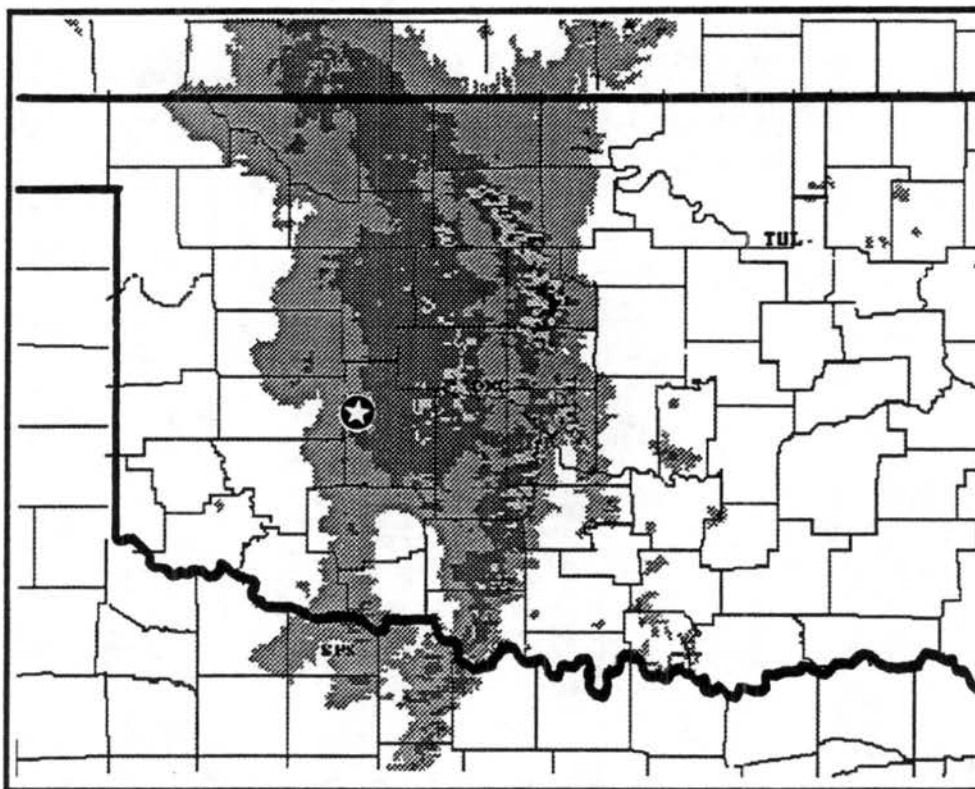


Fig. 3.14. Same as Fig. 3.12 except for 0712 UTC 24 May 1991 electric field meter launch/flight. Due to the presence of large horizontal electric fields, charge densities are not computed for this electric field profile.

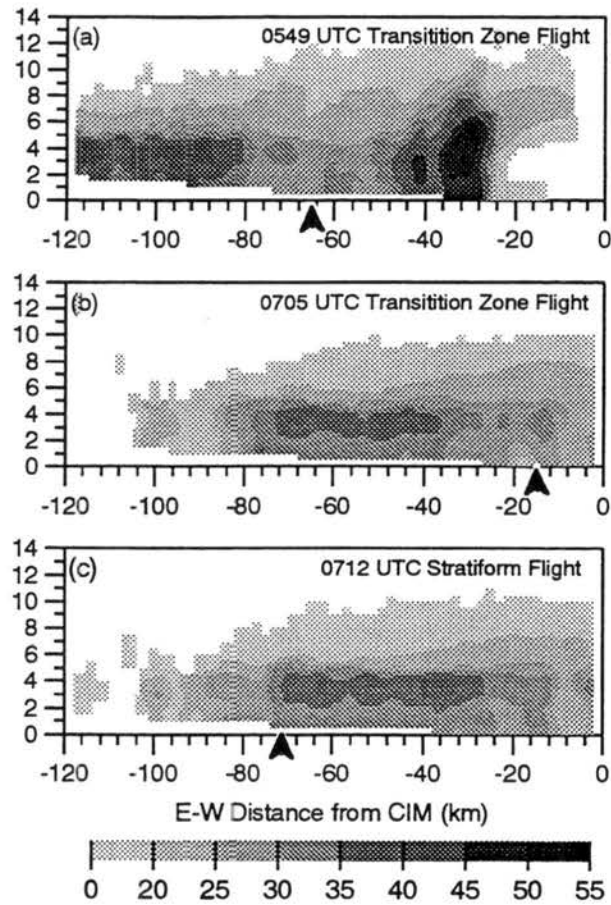


Fig. 3.15. Vertical cross-sections of 24 May 1991 (symmetric) MCS radar reflectivity for (a) 0549, (b) 0705, and (c) 0712 electric field meter launches. Arrows denote approximate, line-relative locations of the electric field meter launches. Launch positions depicted in figure have been adjusted to account for a storm movement of 11.7 m s^{-1} from 270° from the nearest radar analysis time available.

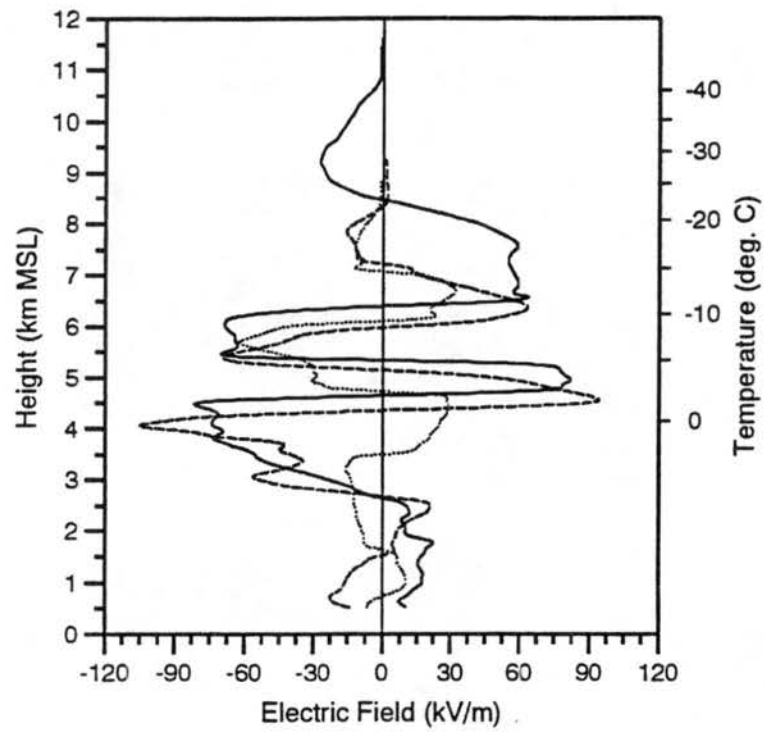


Fig. 3.16. Composite depiction of 0549 (solid line), 0705 (dashed line), and 0712 (dotted line) UTC 24 May 1991 (symmetric) MCS electric field meter profiles.

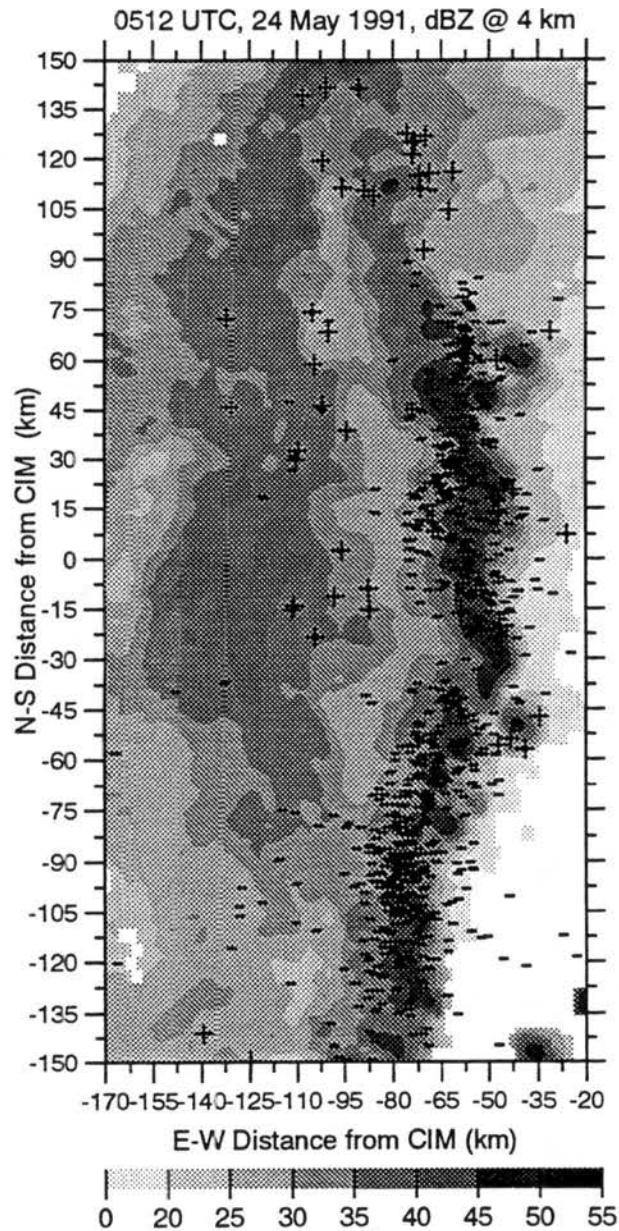


Fig. 3.17. Cloud-to-ground (CG) lightning flashes and bright band level radar reflectivity at 0512 UTC. Negative CG flashes are represented by a (-) while positive CG flashes are represented by a (+). Lightning depicted in the figure represents 30 minutes of data centered on the radar analysis time.

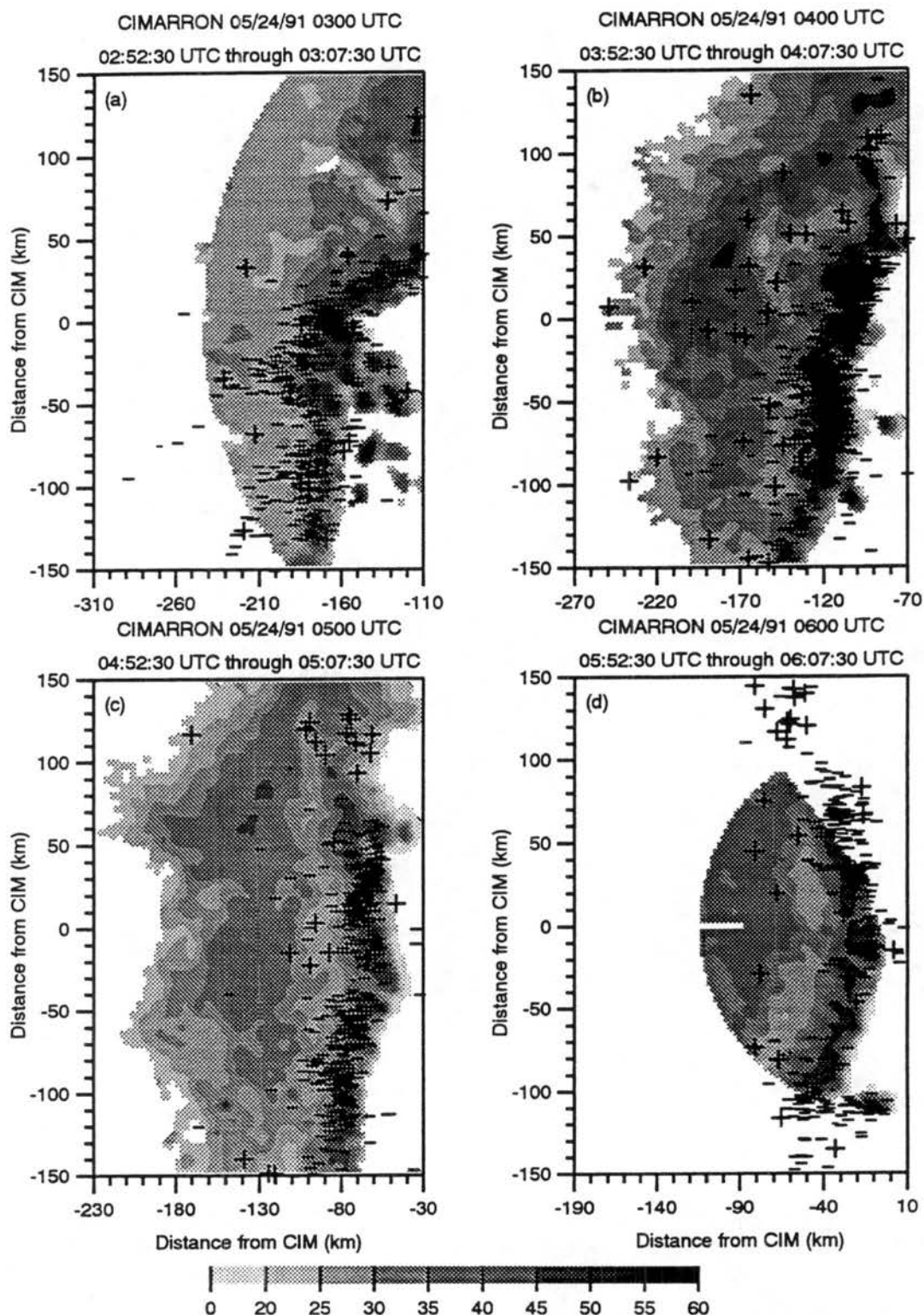


Fig. 3.18. Cloud-to-ground (CG) lightning flashes and low-level radar reflectivity at (a) 0300, (b) 0400, (c) 0500, and (d) 0600 UTC 24 May 1991 (symmetric MCS). Negative CG flashes are represented by a (-) while positive CG flashes are represented by a (+). Lightning depicted in each figure represents 15 minutes of data (times listed in figure header) centered on each radar analysis time.

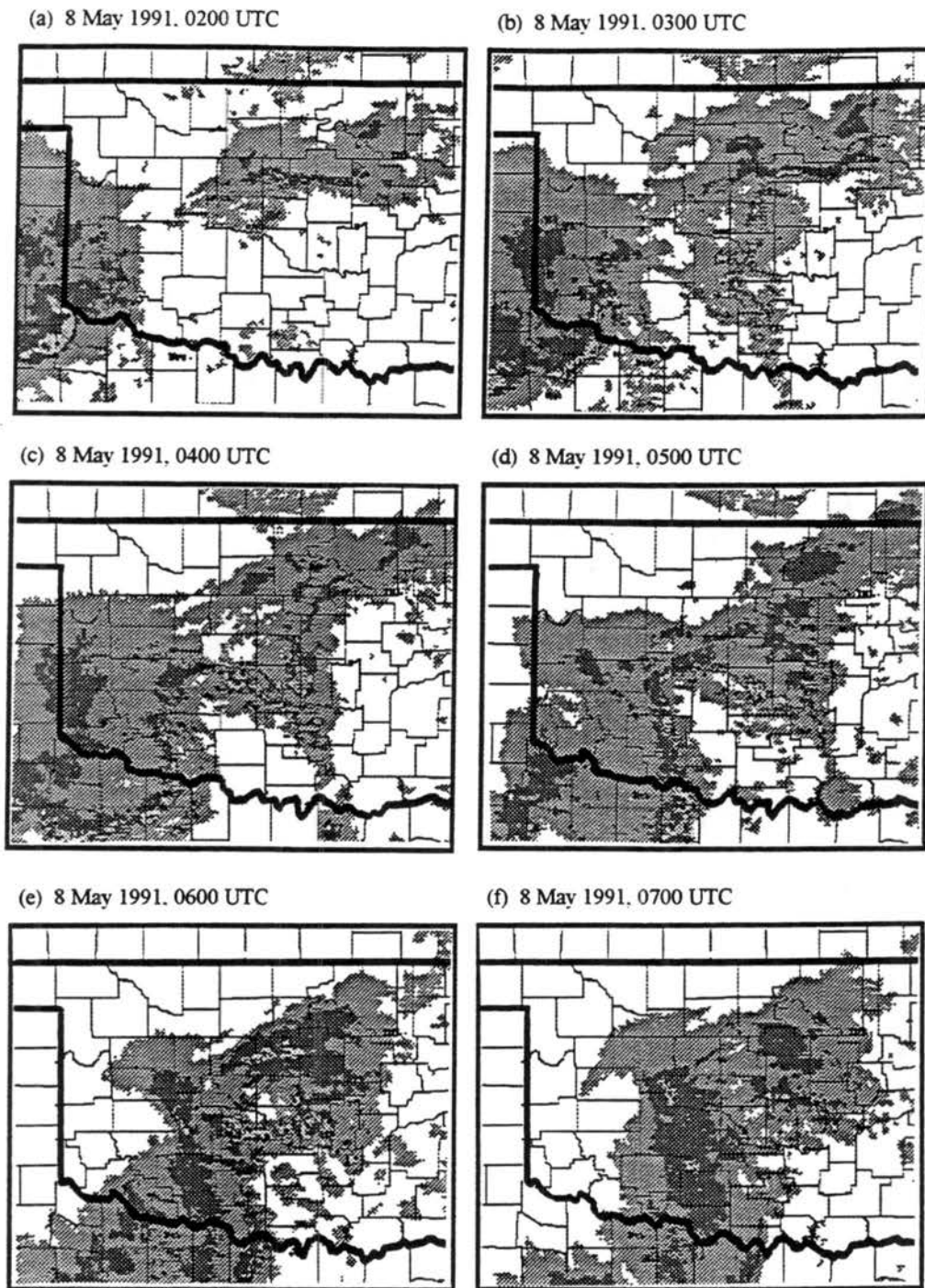


Fig. 3.19. Composite low-level radar reflectivity patterns of the 8 May 1991 (asymmetric) MCS at (a) 0200, (b) 0300, (c) 0400, (d) 0500, (e) 0600, and (f) 0700 UTC. The four shading intensities indicate radar reflectivity ranges of 18-30 dBZ, 30-41 dBZ, 41-46 dBZ, and > 46 dBZ, respectively. Radar data from WSI.

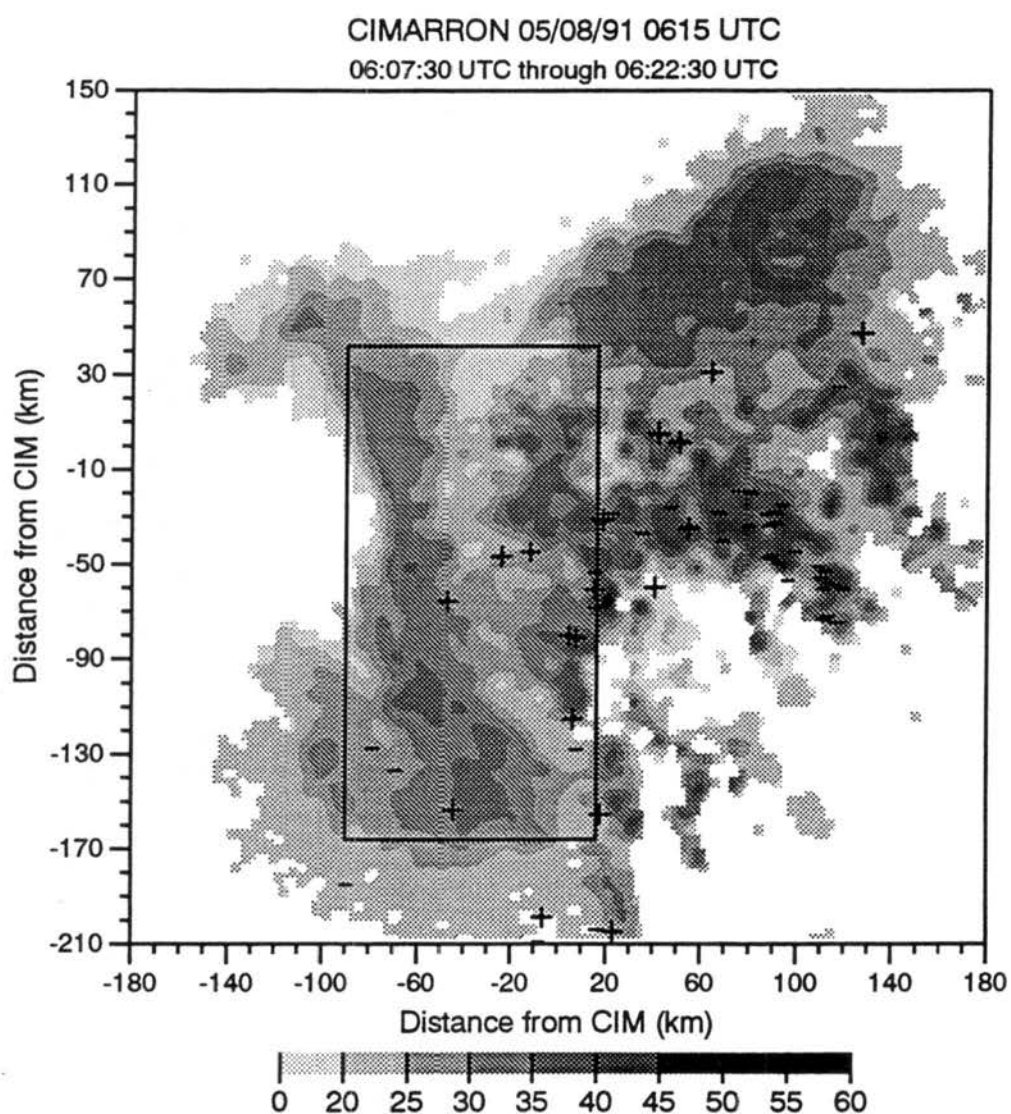


Fig. 3.21. Low-level radar reflectivity pattern (from CIM) at 0615 UTC, 8 May 1991 and CG lightning strike locations from 060730-062230 UTC, 8 May 1991. The grid box overlaid on the low-level radar reflectivity pattern depicts the system-relative location of dual-Doppler analysis presented in Fig. 3.22.

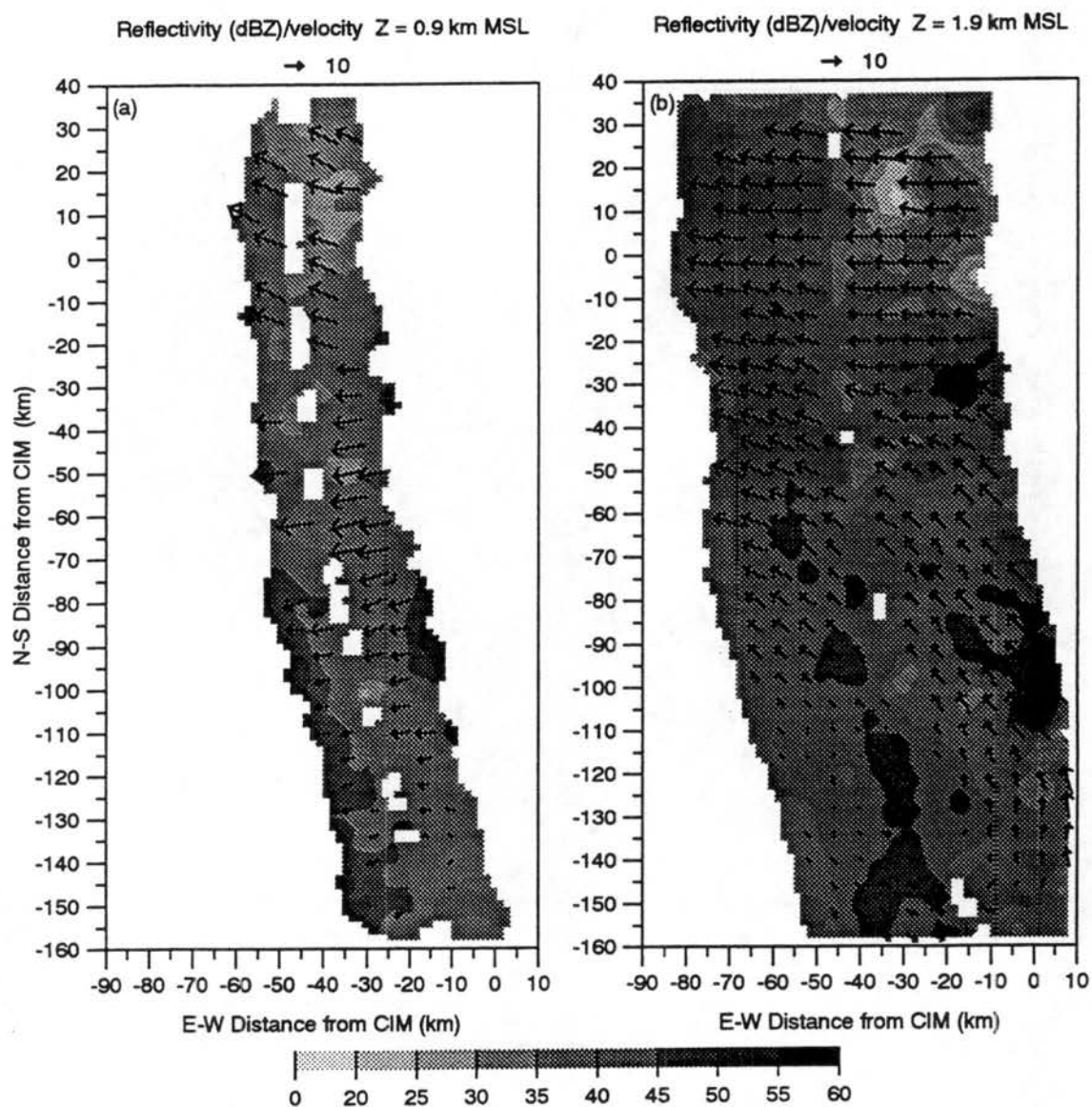


Fig. 3.22. Dual-Doppler radar analysis at (a) 0.9, (b) 1.9, (c) 2.9, (d) 3.9, (e) 4.9, (f) 5.9, (g) 6.9, and (h) 7.9 km MSL. Analysis is constructed from 0600-0628 UTC, 8 May 1991 P-3 flight leg (cf. Figs. 3.20a, 3.21). Vectors (plotted every third grid point) represent system-relative horizontal flow in m s^{-1} . The scale vector at the top of each plot represents a wind speed of 10 m s^{-1} .

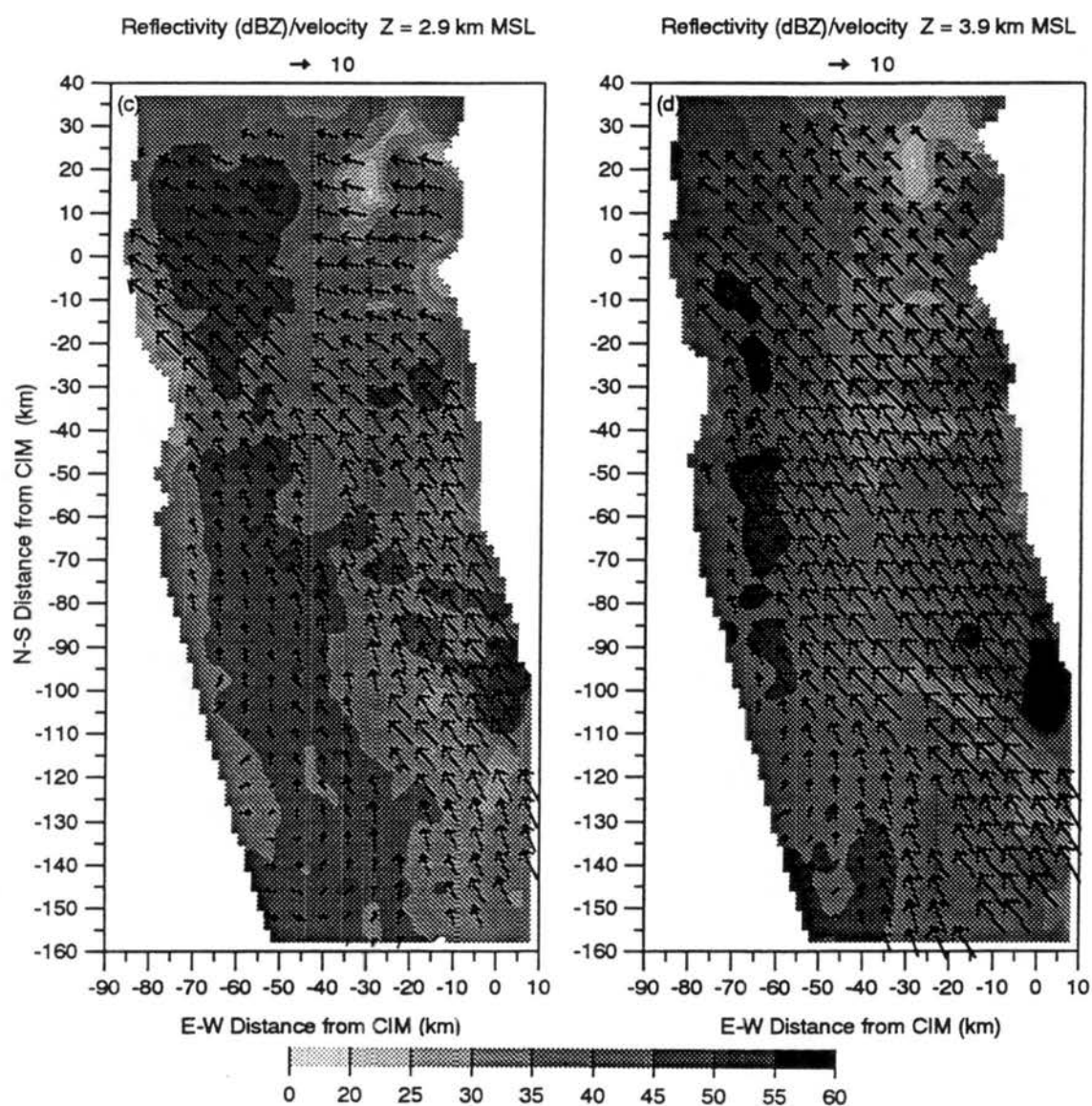


Fig. 3.22. Continued.

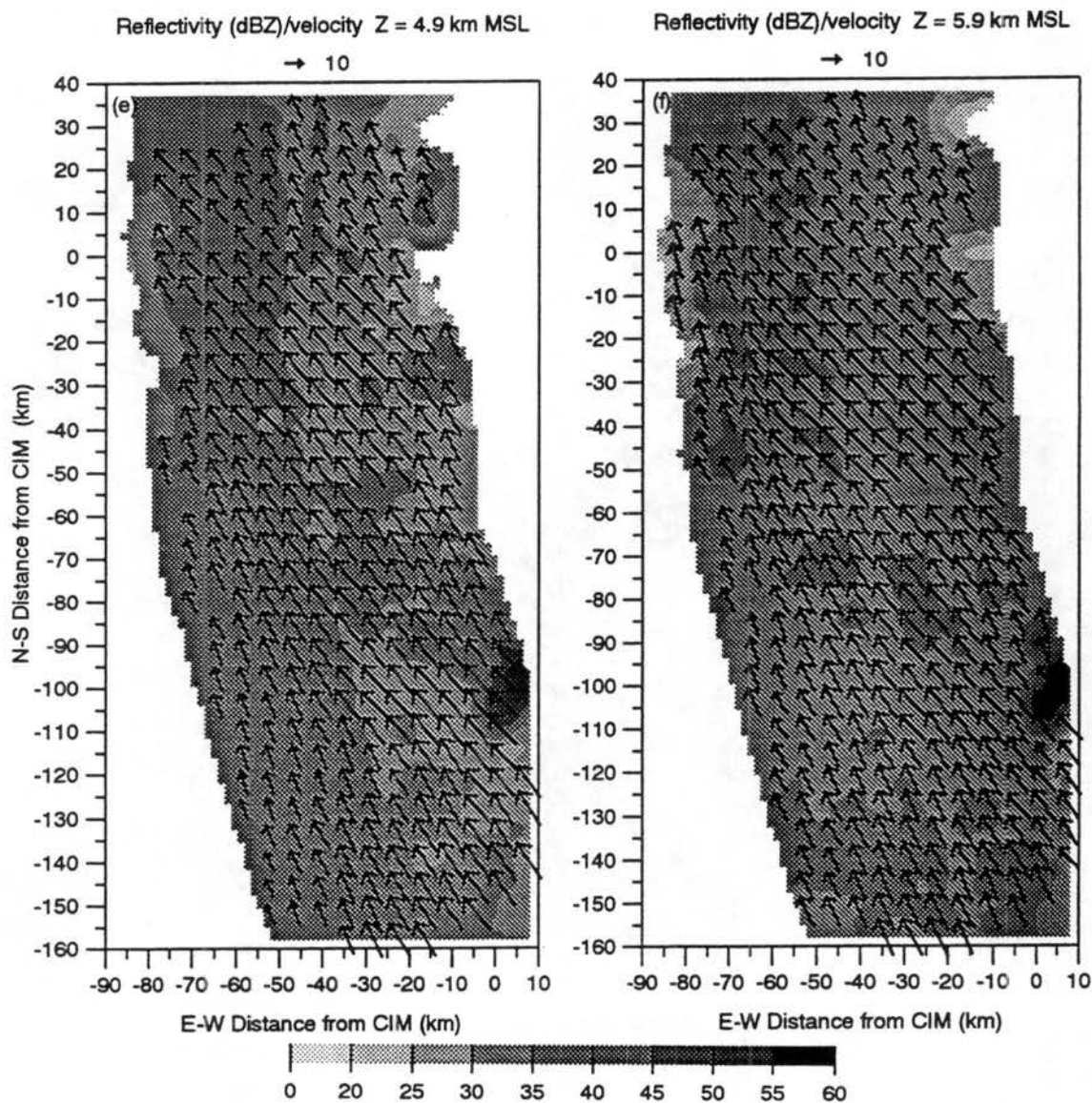


Fig. 3.22. Continued.

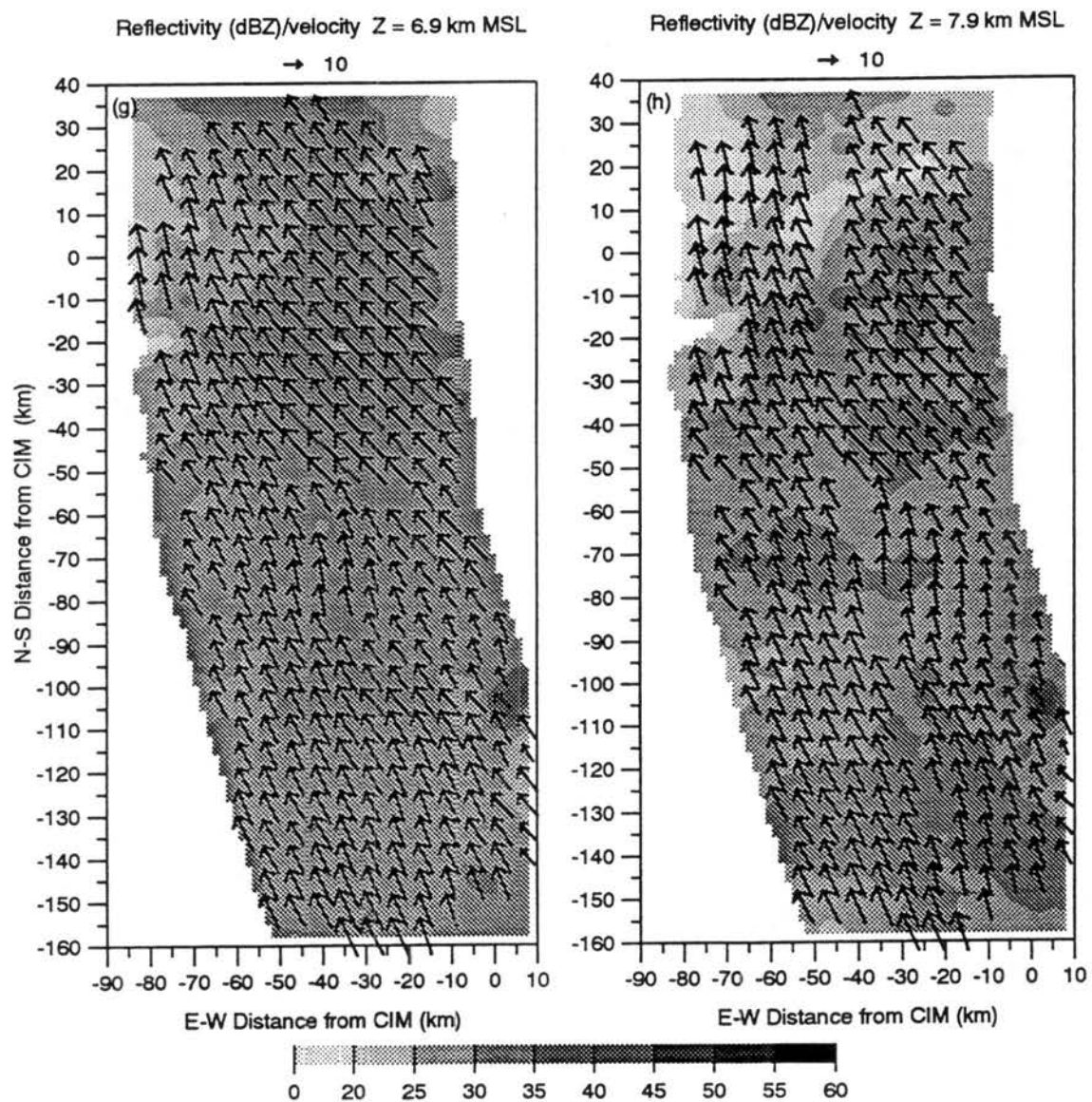


Fig. 3.22. Continued.

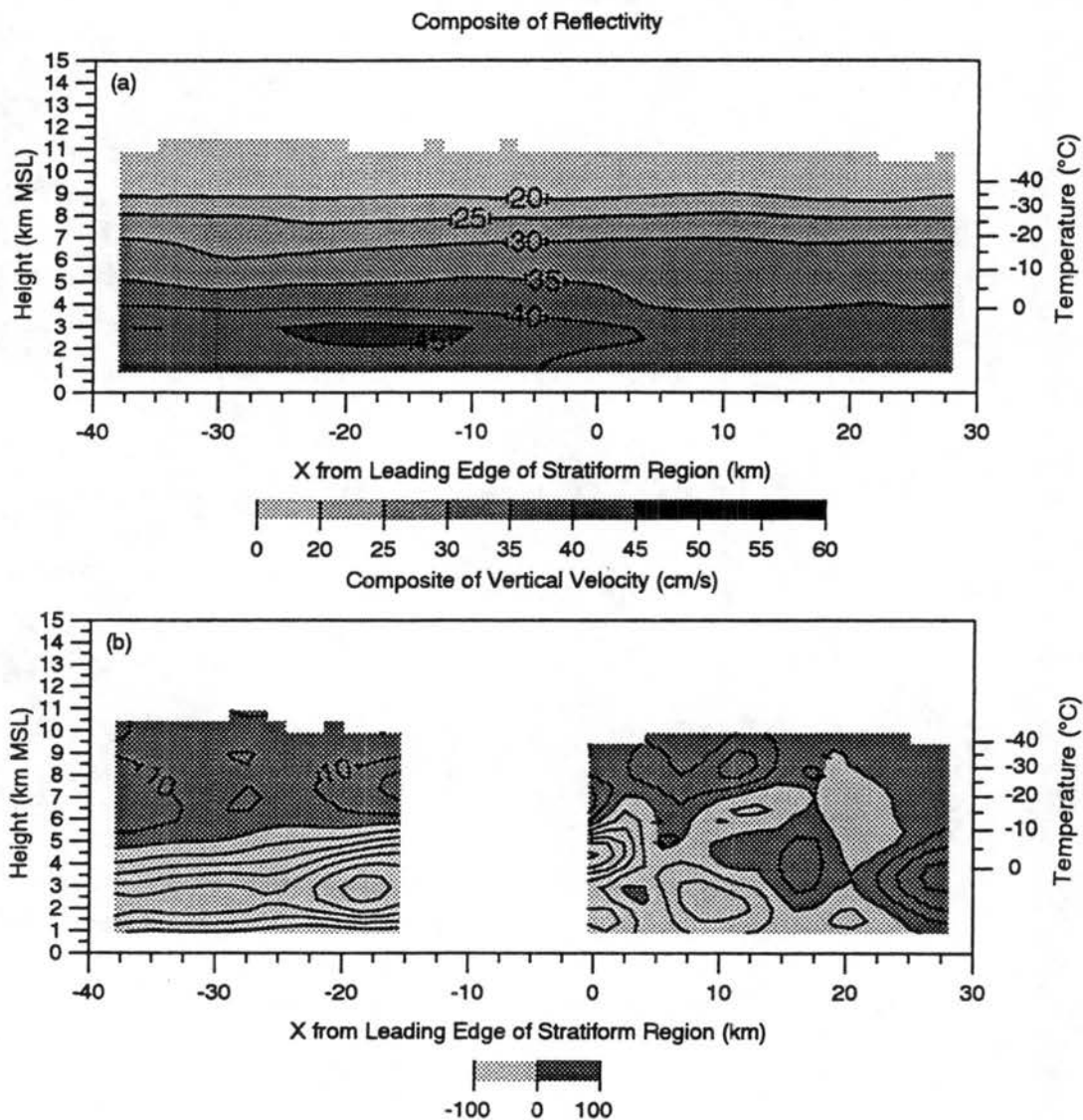


Fig. 3.23. Composite line-normal profile of (a) radar reflectivity (dBZ), and (b) vertical velocity (cm s^{-1}) constructed from all 8 May 1991 P-3 transition zone/stratiform region flight legs. In (b), dark shading indicates updraft while light shading indicates downdraft, with vertical velocity contoured every 10 cm s^{-1} . Plotted with respect to the leading edge of the stratiform precipitation region.



910508. 70603. 4 CLOUD PROBE 2D-C+2D-P O/L



910508. 70603. 158 PRECIP PROBE DPT+PCP O/L

Fig. 3.24. NOAA P-3 shadow graph imagery obtained at a temperature of -8.5°C in 8 May 1991 (asymmetric) MCS (cf. Fig. 3.10 for symmetric MCS).

CIMARRON 05/08/91 0615 UTC

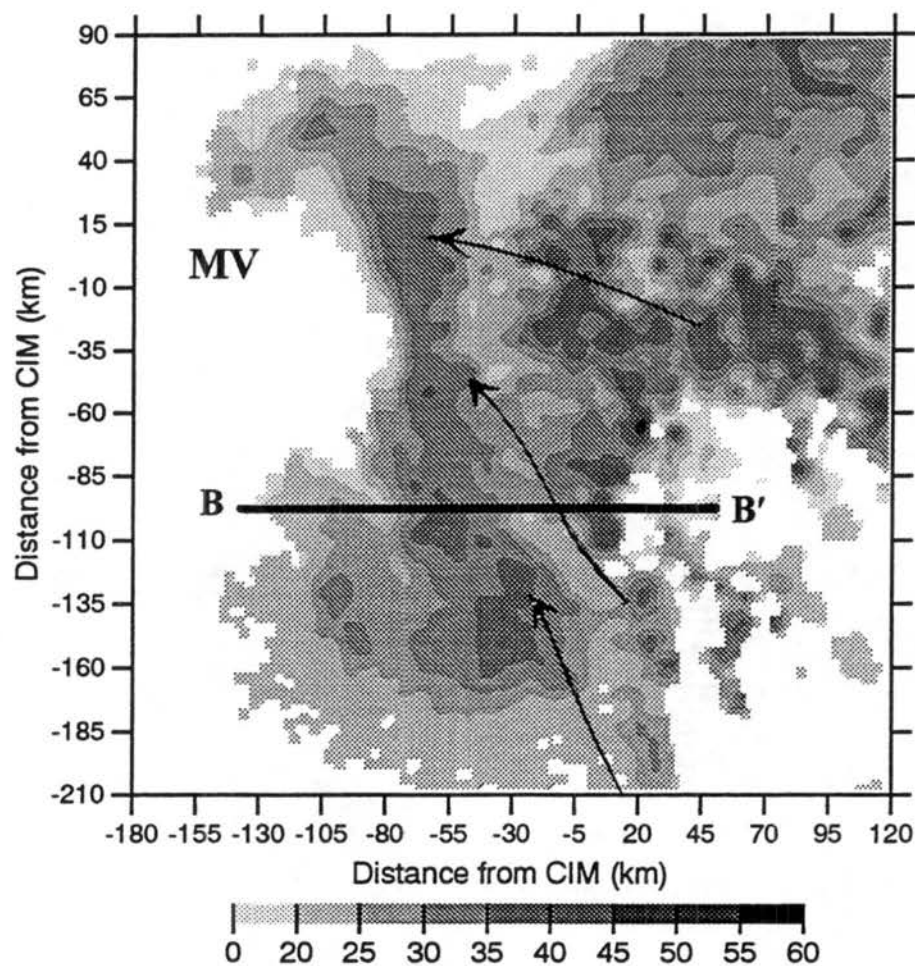


Fig. 3.25. Schematic of ice particle trajectories in 8 May 1991 (asymmetric) MCS overlaid on low-level (CIM) radar reflectivity at 0615 UTC. Arrows indicate approximate path taken by ice particles ejected from convective line at 8.5 km AGL to the location where they reach the melting level. Line BB' represents approximate location of conceptual model presented in Fig. 4.6. MV denotes approximate location of the asymmetric MCS's mesovortex center at this time.

(a) 8 May 1991, 0659 UTC

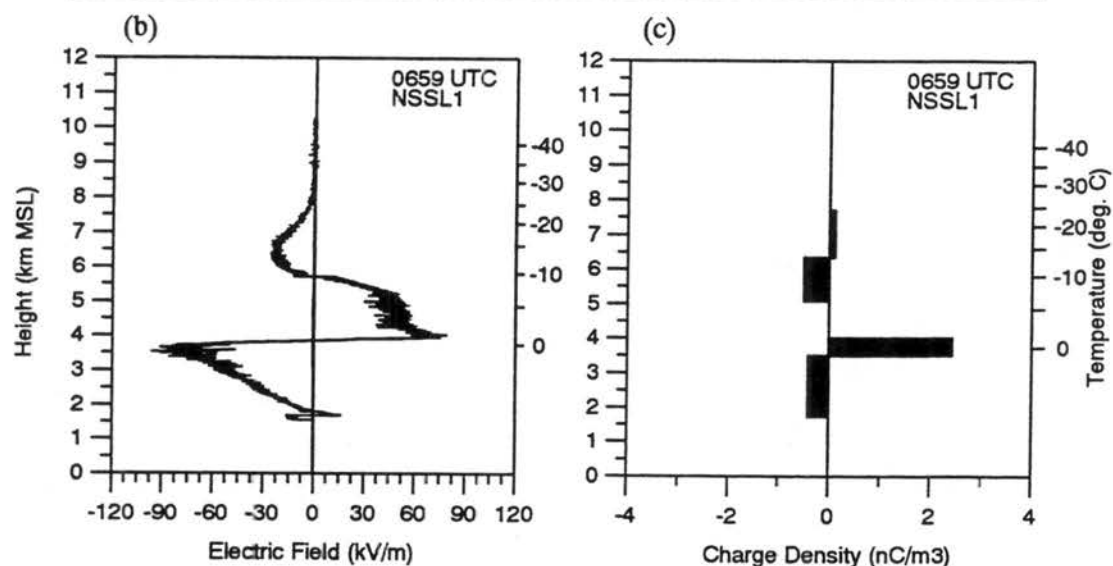
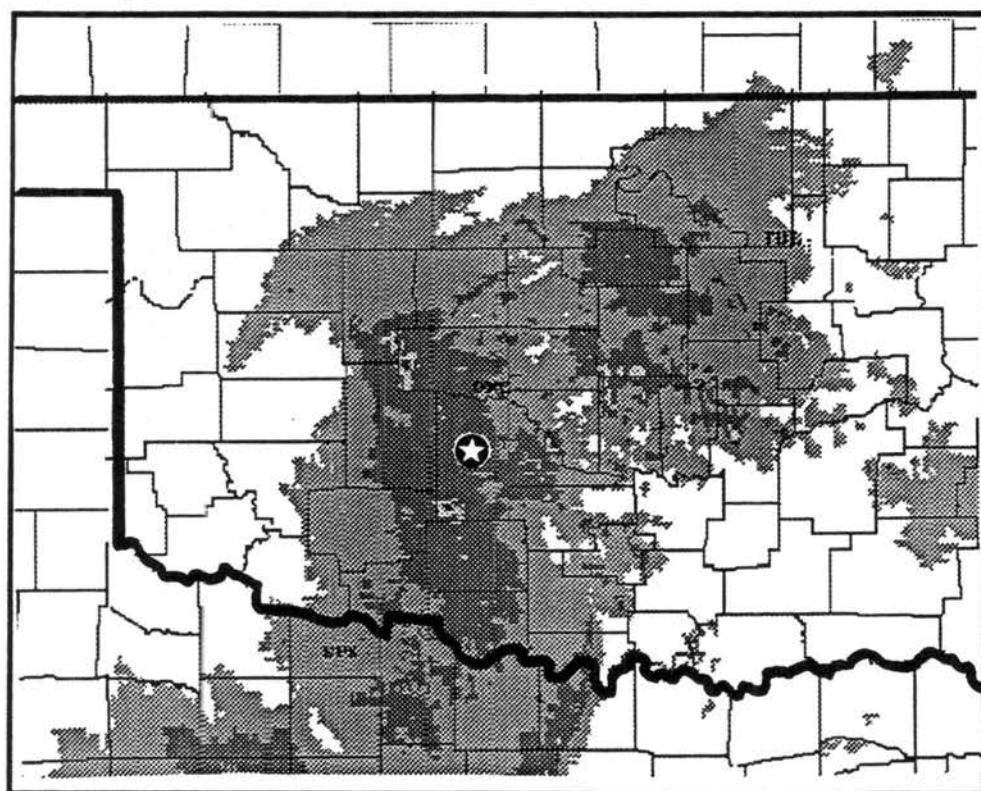


Fig. 3.26. (a) Location of 0659 UTC 8 May 1991 electric field meter launch (indicated by star) overlaid on composite of low-level radar reflectivity patterns of the 8 May 1991 (asymmetric) MCS at 0700 UTC. (b) Electric field profile from the 0659 UTC electric field meter flight. (c) Charge densities (nC m^{-3}) derived from the 0659 UTC electric field meter profile using Gauss' Law (Eqn. 2.3). The four shading intensities in (a) indicate radar reflectivity ranges of 18-30 dBZ, 30-41 dBZ, 41-46 dBZ, and > 46 dBZ, respectively. Radar data from WSI.

(a) 8 May 1991, 0755 UTC

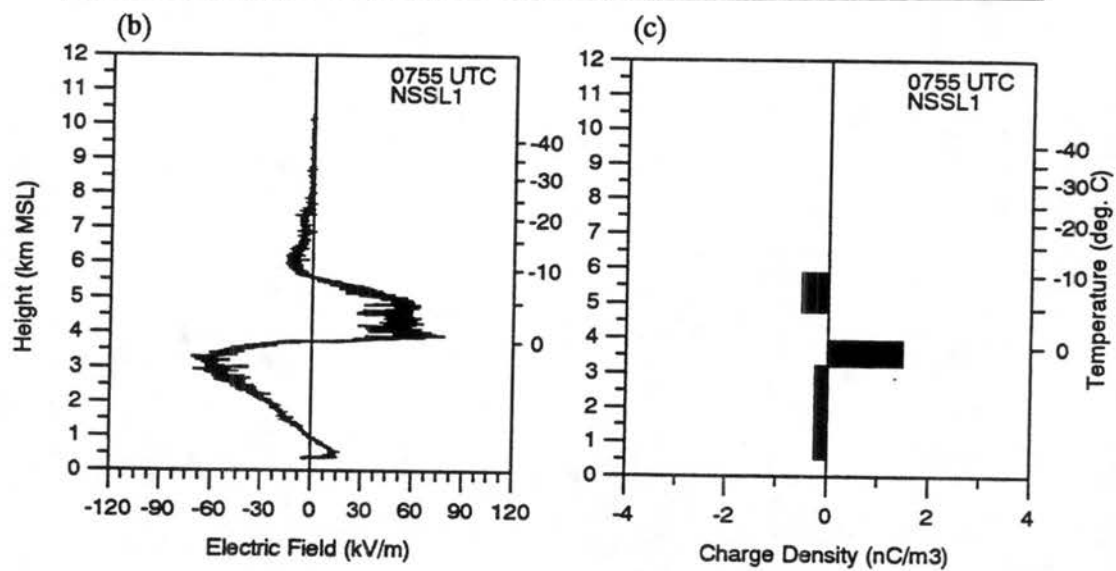
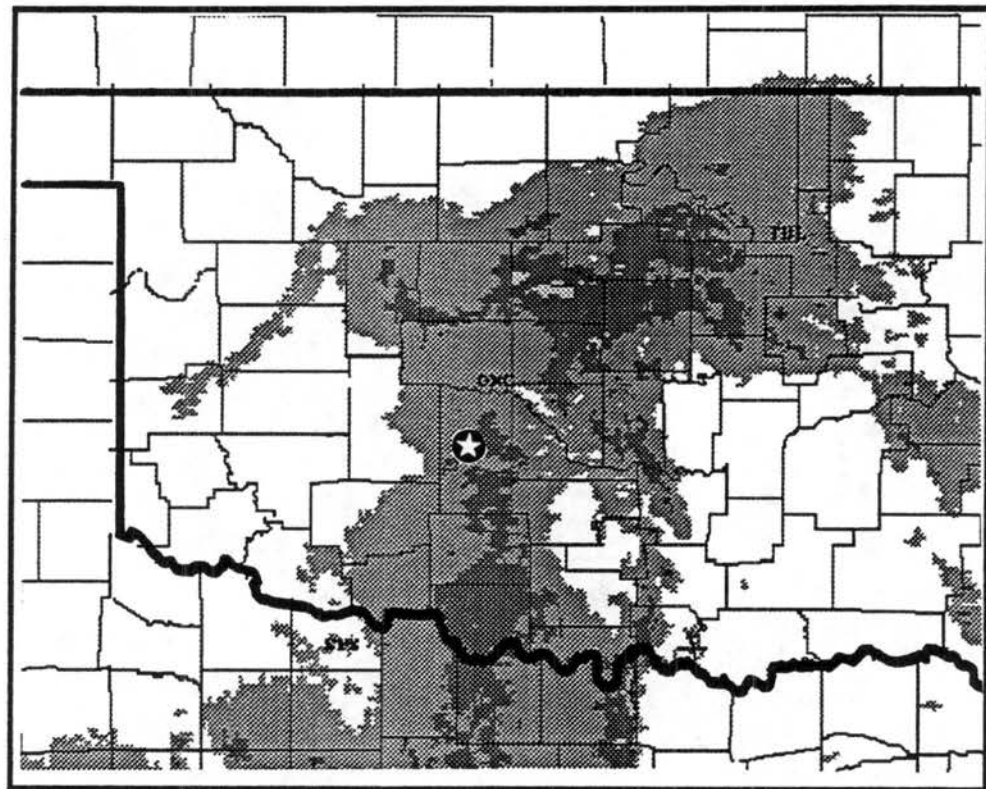


Fig. 3.27. Same as Fig. 3.26 except for 0755 UTC 8 May 1991 electric field meter launch/flight.

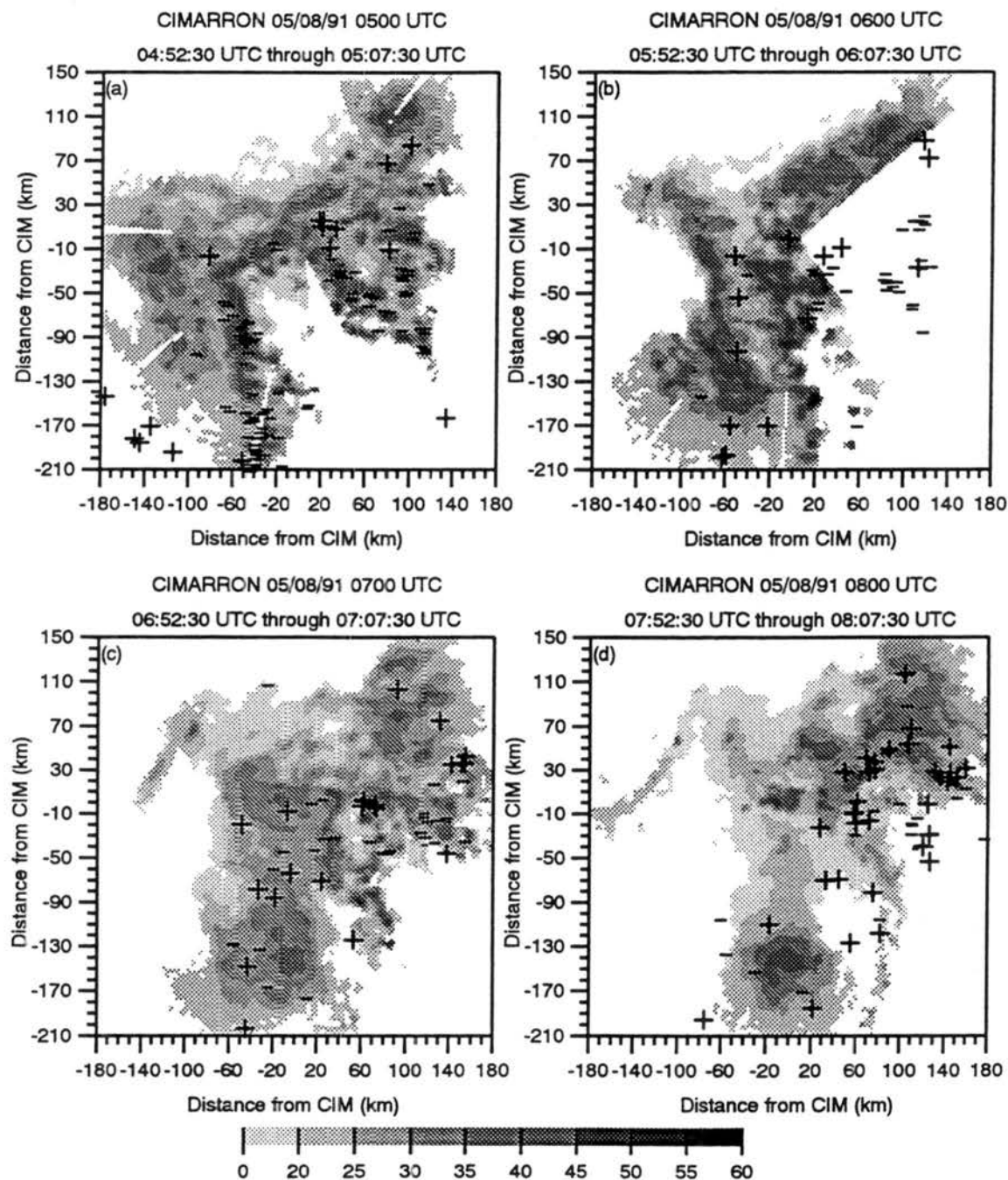


Fig. 3.28. Cloud-to-ground (CG) lightning flashes and low-level radar reflectivity at (a) 0500, (b) 0600, (c) 0700, and (d) 0800 UTC 8 May 1991 (asymmetric MCS). Negative CG flashes are represented by a (-) while positive CG flashes are represented by a (+). Lightning depicted in each figure represents 15 minutes of data (times listed in figure header) centered on each radar analysis time.

CHAPTER 4

COMPARISON OF SYMMETRIC AND ASYMMETRIC MESOSCALE CONVECTIVE SYSTEMS

A detailed comparison of the kinematic, microphysical, and electrical structures of the symmetric and asymmetric MCSs is presented in this chapter. Particular emphasis is placed on understanding relationships between observed charge transition levels, microphysical growth regimes, and MCS kinematic structure and evolution. Finally, we present conceptual models that summarize the results of the observations.

4.1 Electrical structures

In Chapter 3, we noted that the primary differences between the symmetric and asymmetric MCS electrical structures are most evident in the temperature region from -5 to -15°C , where the symmetric MCS exhibits two additional layers of charge not evident in the asymmetric case. We begin our analysis by discussing the two symmetric MCS transition zone electric field profiles (see Figs. 3.12-3.13) relative to the 0 , -12 , and -40°C temperature levels (Fig. 4.1). These three temperature levels were chosen since they are the same temperatures depicted in the microphysical conceptual model shown in Fig. 1.3 (from Houze, 1989). Thus, Fig. 4.1 can be compared to Fig. 1.3 to understand the mean microphysical conditions associated with each layer of charge (addressed in more detail later). In the conceptual model depicted in Fig. 1.3, Houze (1989) noted that the 0 to -12°C layer is significant since ice particles readily aggregate to form

snowflakes over this temperature zone. Additionally, Houze noted that riming may also take place within this zone. Though in situ observations of MCS stratiform LWCs are still quite limited, our observations indicate that mixed-phase conditions were present to -20°C (the maximum altitude attained by the P-3 aircraft).

As can be seen by comparing the profiles in Fig. 4.1, in the 76 minutes that elapsed from the first transition zone EFM launch to the second (0549 to 0705 UTC), electrical stratiform cloud tops, which are defined here as where the electric field becomes nearly zero (due to screening layer formation), dropped from 11.8 km to 8.4 km MSL. Nevertheless, during this same time period, the electrical structure in the 0 to -12°C region remained essentially unchanged. Within this region, both electric field profiles exhibited three intense layers of charge with charge transitions in both profiles occurring at approximately 0 and -12°C (see also Tables 3.1 and 3.2). Above the -12°C level, the overall charge structure in the 0705 UTC profile is also very similar to that observed earlier (i.e., positive charge above negative), although the depth of those two upper layers (combined) is approximately 2.5 km less in the 0705 UTC electric field profile than in the 0549 UTC electric field profile.

The temperatures at which these charge transitions occur are now examined in more detail using the data presented in Tables 3.1 and 3.2. In particular, we focus on the common charge structure in the 0 to -12°C layer. From these data, it is apparent that both symmetric MCS profiles have sharp charge transitions (negative charge above positive) at -11.5°C ; similarly, both symmetric MCS profiles also have common charge transitions at -1.3°C . In the 0549 UTC profile, the charge transition at -1.3°C is positive above negative. In the 0705 UTC profile, on the other hand, the charge transition at -1.3°C is negative above positive with the charge polarity again switching back to negative at a temperature of $+0.5^{\circ}\text{C}$. Thus, while both profiles essentially exhibit a three-layered charge structure in the 0 to -12°C layer, the lower positive

charge layer in the 0549 UTC profile was located entirely above the 0°C isotherm while the lower positive charge layer in the 0705 UTC profile straddled the 0°C isotherm. While we do not understand why the polarity of charge transitions differs between the 0549 and 0705 UTC profiles at the -1.3°C level (though it might be related to the 0549 UTC balloon-borne EFM apparently ascending through the melting level at a location that slightly *preceded* the stratiform bright band), below we will address the possible significance of this lower charge transition level.

In comparing the electric field profiles from the asymmetric MCS (0659 and 0755 UTC, Figs. 3.26 and 3.27 and Tables 3.3 and 3.4, respectively) to those for the symmetric MCS, it is apparent that many differences exist. Most significantly, no charge transition occurs in the mid-levels of the cloud for the asymmetric case. Near the melting level, similarities in the charge structure are evident. When examined in detail using the data in Tables 3.3 and 3.4, we find that both asymmetric MCS profiles have charge transitions of negative above positive that occur at approximately -1.3°C, which is the same temperature level at which lower charge transitions occurred in the symmetric MCS electric field profiles (and also the same *polarity* of charge transition as exhibited by the 0705 UTC symmetric MCS profile). In the 0659 UTC asymmetric MCS profile (Fig. 3.26), the charge polarity switches back to negative at +0.5°C, which is similar to that seen in the 0705 UTC symmetric MCS electric field profile; in the 0755 UTC asymmetric MCS profile (Fig. 3.27), the charge polarity switched back to negative at +2.3°C.

It is interesting to compare these lower charge transitions to the results of Shepherd et al. (1996), who concluded that melting charging (e.g., Simpson, 1909; Dinger and Gunn, 1946; Drake, 1968) may be the primary cause of the charge density found at and below the 0°C isotherm in electrified stratiform clouds. While melting charging might offer an explanation for the charge transitions found *below* the melting level, it does not explain the common charge transition level that is located at -1.3°C (i.e., *above* the melting level) in all of our electric field

soundings. Shepherd et al. (1996) speculated on possible mechanisms that might account for the charge layers in their soundings extending above the 0°C isotherm, suggesting that either turbulent mixing of charge or a lowering of the 0°C isotherm with time might result in the charge transition extending to levels above 0°C. After a review of the microphysical data, we will offer an alternative explanation.

4.2 Kinematic and microphysical structures

As noted in Chapter 3, the composite dual-Doppler analyses suggest that the vertical velocity structures of the symmetric and asymmetric MCSs were quite different, as illustrated by Fig. 4.2, which depicts the mean vertical velocities of the symmetric and asymmetric MCS transition zone (Fig. 4.2a) and stratiform precipitation regions (Fig. 4.2b) with respect to temperature. The most marked difference in the updraft structure of the two MCSs is seen in the transition zone (Fig. 4.2a) where, despite the strong mechanically forced downdraft observed behind the convective line (cf. Figs. 3.5 through 3.7), the symmetric MCS transition zone profile indicates mean ascent at all levels (peaking at approximately 25 cm s^{-1} at -15°C). The vertical velocity profile of the asymmetric MCS transition zone, on the other hand, indicates weak descent at most levels.

The mean vertical velocities of the symmetric and asymmetric MCS stratiform precipitation region are shown in Fig. 4.2b. Though the mean stratiform region vertical velocity profiles from these two systems are very similar, a comparison of those profiles with the line-normal cross-sections depicted in Figs. 3.8 and 3.23 suggests that their overall *structures* are quite different. That is, the symmetric MCS mesoscale updraft appears to be rather deep near the leading edge of the stratiform precipitation region but much shallower towards the rear of the system (with the updraft/downdraft interface sloping upwards slightly with distance from the convective line). The updraft in the asymmetric MCS appears to be more horizontally uniform across the extent of

the MCS stratiform precipitation region. It should be noted that the radar data from which the vertical velocity composites were computed were collected at locations immediately behind the convective line and during the MCSs dissipative stage (at which time the stratiform precipitation region behind the convective line was growing in areal extent, cf. Fig. 3.19). Observational studies have shown that upward vertical motions in mature asymmetric MCSs is typically found in the front-to-rear portion of the vortex whereas descent is more commonly found in the rear-to-front flow (e.g., Jorgensen and Smull, 1993; Scott and Rutledge, 1995).

Thus, when comparing these updraft structures to the balloon-borne EFM flight tracks, we suggest that the two symmetric MCS electric field profiles (Figs. 3.12 and 3.13) were likely obtained in a region of deeper mesoscale ascent than were the two asymmetric MCS electric field profiles (Figs. 3.26 and 3.27). Fig. 4.3 depicts mean charge layers, which are derived from a combination of electric field profiles obtained in each MCS, overlaid on a conceptual model depicting the depositional growth regime (from Hallett, 1984). It is interesting to note that the positive charge layer located immediately below -12°C in our symmetric MCS electric field profiles is also located immediately below the level at which ice growth due to the Bergeron-Findeisen process (e.g., Bergeron, 1935; Findeisen, 1938) would be a maximum. In addition, as can be seen from Fig. 4.3, dendritic growth is favored at temperatures colder than -12°C in water saturated conditions. In Chapter 3, we discussed microphysical observations in the symmetric MCS that indicated dendritic growth and cloud LWCs (averaged over the -8 to -16°C layer) of approximately 0.23 g m^{-3} , both of which are consistent with water saturated conditions through a deep layer. In contrast, microphysical observations in the asymmetric MCS indicated no dendritic growth and cloud LWCs (averaged over the -8 to -16°C layer) of only 0.12 g m^{-3} , which is only slightly above the minimum detectable threshold of the instrument. Though these observations indicate that at least some cloud water was encountered in the asymmetric MCS, it

was clearly not as abundant or widespread as in the symmetric MCS, and therefore widespread water saturated conditions were not generally supported.

The importance of these layers are examined further by analyzing the charge layers shown in Fig. 4.3 for the symmetric MCS in the context of mean and maximum aggregate sizes (Fig. 4.4, from Hobbs et al., 1974). We use data from this study for comparison since our symmetric MCS P-3 optical array probe observations had data gaps through both of these layers. In Fig. 4.4, the solid curve represents the median of observed aggregate sizes whereas the dashed line represents the maximum observed aggregate sizes. As in Fig. 4.3, the solid line drawn at approximately -1.3°C depicts the lower charge transition level that is common to all symmetric and asymmetric electric field profiles (see Section 4.1).

By comparing the peaks in maximum aggregate size to the location of the charge transitions found in our EFM soundings, it is important to note that both the upper and lower charge transitions found in our symmetric MCS electric field profiles are collocated with peaks in maximum aggregate size (where the upper peak is likely related to interlocking dendritic arms and the lower peak to ice particles becoming coated with a pseudo-liquid layer of water). These observations suggest that gravitational separation is a possible candidate to explain the charge transitions at both -11.5 and -1.3°C . That is, if some charging mechanism (e.g., Takahashi, 1978) were responsible for large and small ice particles obtaining opposite polarity charges, subsequent aggregation and gravitational separation resulting from the increased fallspeed may be sufficient to explain a charge transition that coincides with these peak aggregation layers (aggregates separating from smaller crystals). In the asymmetric MCS, conditions did not support dendritic growth at temperatures near -12°C , and a secondary aggregation peak was not observed in the stratiform cloud's mid-levels. Also, a charge transition was not observed in this layer.

4.3 Conceptual models

The results of these kinematic, microphysical, and electrical comparisons are summarized by Figs. 4.5 and 4.6, which depict line-normal conceptual models for the symmetric (aligned along the line AA' in Fig. 3.11) and asymmetric (aligned along the line BB' in Fig. 3.25) MCSs, respectively. At the time of EFM data collection, the symmetric MCS (Fig. 4.5a) exhibited a line-normal, system-relative flow structure that consisted of a low-level, rear-inflow jet and ascending, front-to-rear airflow regime. The mesoscale vertical motion structure consisted of a deep transition zone downdraft, which was apparently forced in the upper levels by convergence of convective outflow (cf. Fig. 3.9c), and a trailing region of deep mesoscale ascent that appeared to be deepest and most intense near the leading edge of the radar bright band.

The 0549 and 0705 UTC electric field profiles (Figs. 3.12 and 3.13) were both obtained near the leading edge of the stratiform precipitation region (though it appears that the 0549 UTC EFM may have ascended through the melting level at a location that slightly preceded the radar bright band). Both EFM's were quickly caught up in the strong mid-level, front-to-rear flow and carried rearward into the deep region of mesoscale ascent. The 0712 UTC electric field profile (Fig. 3.14) was obtained at a location near the trailing edge of the stratiform precipitation region and apparently ascended through a much deeper region of mesoscale descent before eventually reaching the mesoscale updraft. The approximate location of these three EFM launches (with the 0549 and 0705 trajectories overlaid), in a system-relative sense, are depicted in Figs. 4.5a-c.

As can be seen from the corresponding microphysical conceptual model (Fig. 4.5b), as the EFM ascended into the mesoscale updraft, it encountered a deep layer of water saturated conditions that favored dendritic growth at and near the -12°C level. Subsequent aggregation of those dendrites at approximately -12°C (as determined by a comparison with the study of Hobbs

et al., 1974) has been found to be collocated with a sharp charge transition at the same level (cf. Fig. 4.5c). At lower levels, a secondary maximum in aggregational growth has been found to be collocated with another charge transition level located immediately above the 0°C isotherm (cf. Fig. 4.5c). It is also possible that high electric fields in this region served to enhance aggregational growth (e.g., Crowther and Saunders, 1973). The significance of these aggregational layers may lie in their ability to separate charge, in a bulk sense, through gravitational separation (discussed above). The actual mechanism responsible for the charge on the ice particles, however, is unknown. We will later offer speculation on possible charge sources and, furthermore, test several mechanisms in the numerical modeling component of this study.

In contrast, the kinematic structure of the asymmetric MCS was quite different. At the time of the EFM data collection, the asymmetric MCS (Fig. 4.6a) exhibited a weak front-to-rear flow at both lower and upper levels with a weak rear-to-front airflow component centered at approximately 3 km. Whereas the line-parallel airflow was, in general, slightly weaker than the line-normal component in the symmetric MCS, it was much stronger than the line-normal airflow in the asymmetric MCS in response to the mesovortex that was located at the northern end of the MCS. This resulted in a mean horizontal flow structure that was primarily line-parallel in direction. The mesoscale vertical motion structure consisted of deep (but weak) descent in the transition zone region. The vertical velocity structure of the stratiform precipitation region was characterized by mesoscale descent below approximately -15°C and mesoscale ascent above that level. At the time of the EFM data collection, the convective line had dissipated significantly and the transition zone was much wider than that observed in the symmetric MCS.

The 0659 and 0755 UTC electric field profiles (Figs. 3.26 and 3.27) were both obtained in the stratiform precipitation region. The approximate locations of these two EFM launches, in a

system-relative sense, are depicted in Figs. 4.6a-c. As can be seen from the corresponding microphysical conceptual model (Fig. 4.6b), as the EFMs ascended into the stratiform precipitation region, they initially encountered a deep region of mesoscale descent that resulted in water subsaturated conditions through a deep layer (though localized pockets of measurable cloud water were occasionally encountered by the P-3 aircraft). As a result, dendritic growth was likely greatly inhibited, gravitational separation did not occur, and a sharp charge transition near the -12°C level was not found. At lower levels, however, a charge transition was observed immediately above the 0°C isotherm (cf. Fig. 4.6c) at the same temperature of a similar charge transition in the symmetric MCS electric field profiles.

4.4 Comparison to other profiles

In this section we compare these data to the other Type A electric field profiles (see Section 1.2.1.3) investigated by Marshall and Rust (1993). We first investigate the electric field profile presented by Schuur et al. (1991), a balloon-borne EFM that ascended into the transition zone of a symmetric MCS. The EFM ascended through the melting level at a location that preceded the radar bright band by several km and then was carried rearward over the stratiform precipitation region by the MCS's mid-level, front-to-rear flow. In comparing the charge transitions, it is immediately apparent that the Schuur et al. (1991) electric field profile (Fig. 4.7) does not exhibit a charge transition near the melting level; it does, however, have a sharp charge transition (negative above positive) at -13.2°C that is very similar to a transition seen in our electric field profiles at -11.5°C . We do not find the disagreement in the lower levels surprising since the EFM ascended through the melting level in the transition zone where aggregation is typically weak (e.g., Yeh et al., 1991; Biggerstaff and Houze, 1993). Though in situ aircraft data were

not available for that MCS, thermodynamic sounding data indicated that ice saturated conditions (with relative humidities with respect to ice of up to 110%) were present through a deep layer.

We also compare our results to two Type A electric field profiles, which were obtained in two Oklahoma MCSs on 4 and 7 June 1989, as discussed by Hunter et al. (1992). It is more difficult to compare these electric field profiles since their thermodynamic data were of poor quality at many levels (especially in the mid-levels). Nevertheless, both electric field profiles exhibited similar transition levels to those presented here with an important difference: the polarity of charge transitions in Hunter et al. are the opposite to those found at similar levels in the present study. We do not understand why the polarity of the charge transitions in that study are different than those presented here. An comparison of the levels at which the charge transitions occurred, however, reveals many similarities. A detailed examination of the temperature data from one of those flights (4 June 1989, see Fig. 4.8) indicates that charge transitions occurred at -1.3°C and -12.6°C . Since temperature data used in the lower half of the sounding were quite good, the -1.3°C charge transition appears to be accurate; temperature data at locations colder than approximately -5°C were interpolated between good quality data points and therefore may be accurate to only within a few degrees. Given this, we find that the charge transition levels exhibited by the 4 June 1989 electric field profile are nearly identical to those found in the two transition electric field profiles presented here (e.g., Fig. 4.1), but still of opposite sign. In situ aircraft data were also available in that MCS. Though not presented in Hunter et al. (1992), our analysis of that data indicates the presence of significant dendritic growth and cloud LWCs of as high as 0.5 g m^{-3} , which is much higher than observed in our symmetric MCS.

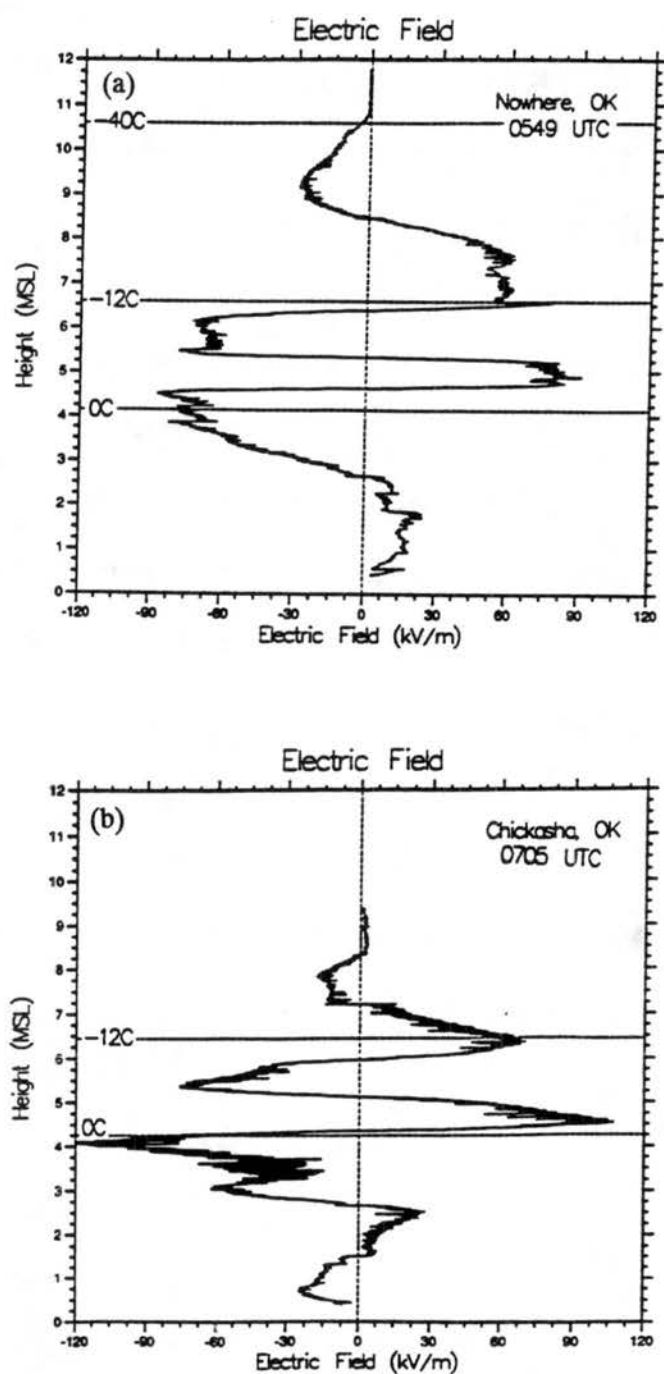


Fig. 4.1. Symmetric MCS electric field profiles at (a) 0549, and (b) 0705 UTC with 0, -12, and -40°C temperature levels overlaid. Fig. 4.1 can be compared with Fig. 1.3, which depicts conceptual model (from Houze, 1989) of MCS microphysical structure at the same temperature levels.

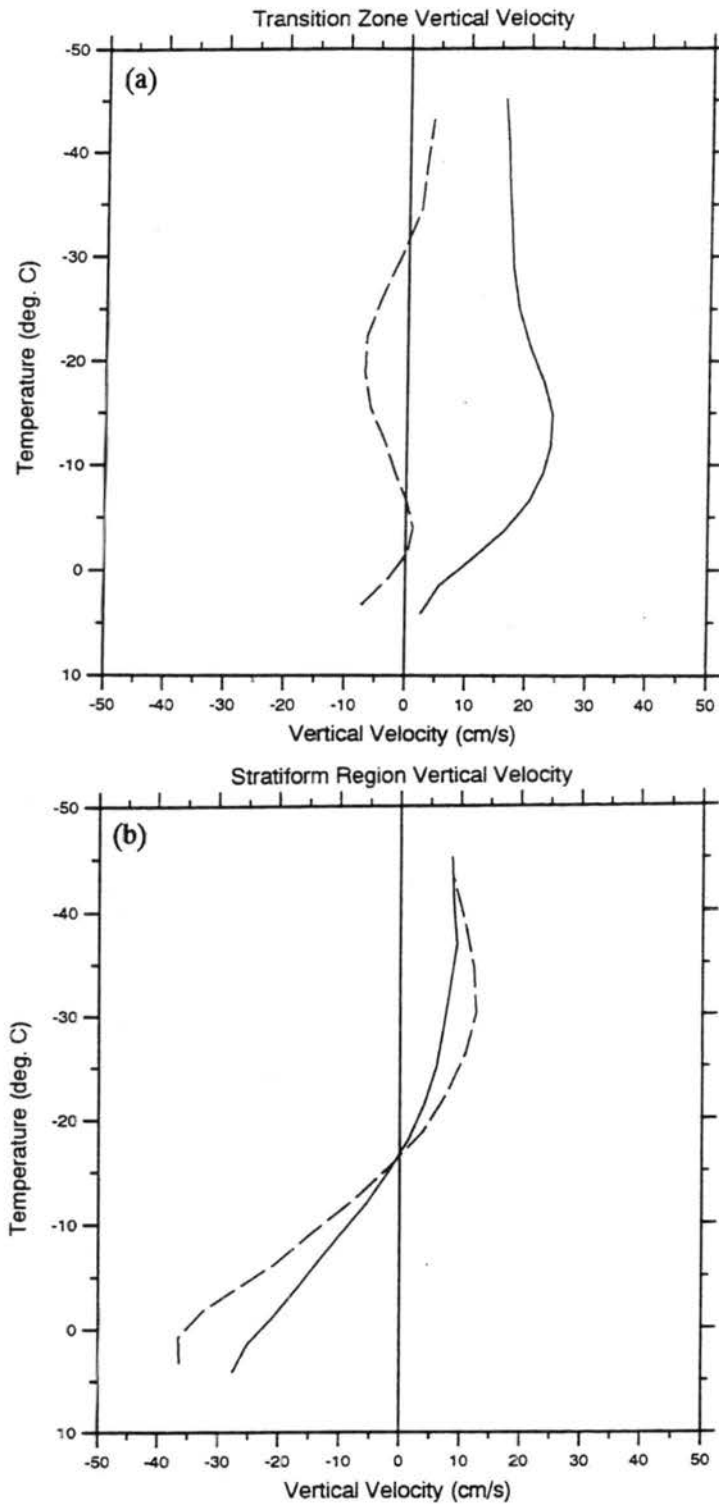


Fig. 4.2. Mean composited vertical velocity of (a) transition zone, and (b) stratiform precipitation region for symmetric (solid line) and asymmetric (dashed line) MCSs, respectively.

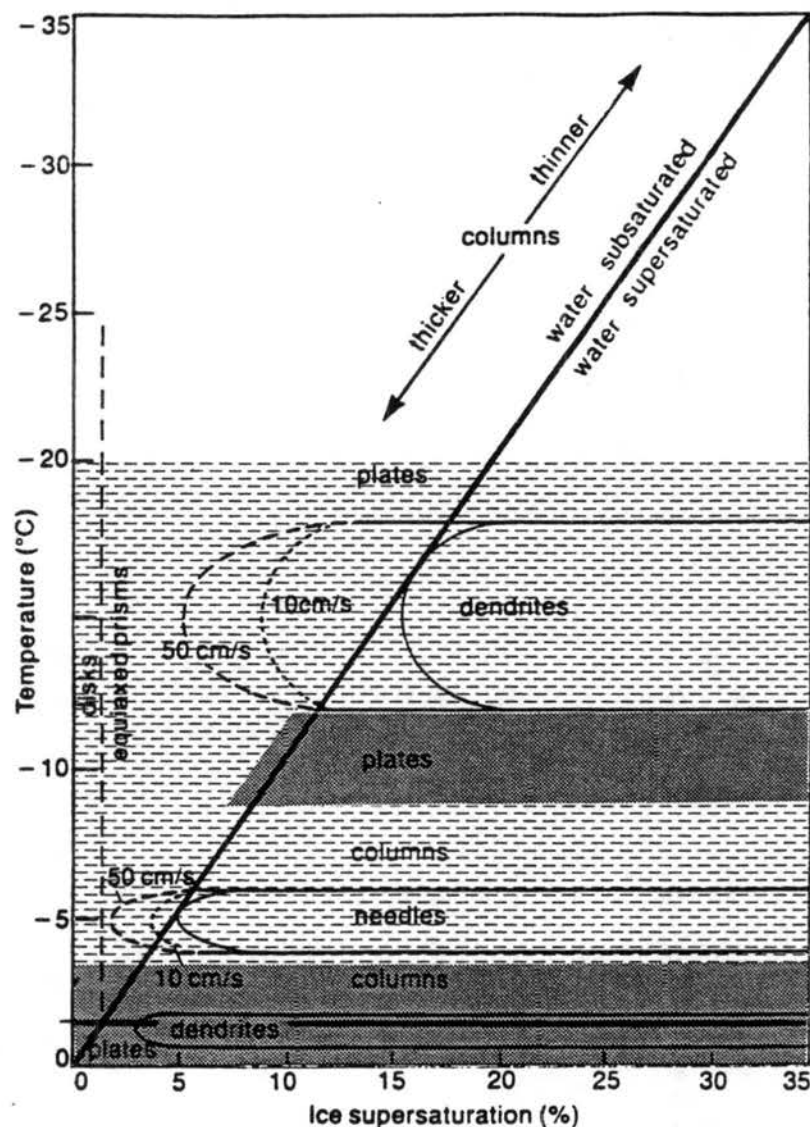


Fig. 4.3. Charge layers from symmetric and asymmetric MCS overlaid on figure depicting depositional growth regime. The solid diagonal line running from the lower left corner to the upper right corner represents boundary between water subsaturated conditions (on the left) and water saturated conditions (on the right). Positive charge layers are indicated by dark shading and negative charge layers by light shading. Charge layers to right of diagonal line represent charge structure of symmetric MCS while charge layers to the left of diagonal line represent charge structure of asymmetric MCS. Symmetric MCS charge structure shown reflects composite of 0549 and 0705 UTC 24 May 1991 electric field profiles while asymmetric MCS charge structure reflects composite of 0659 and 0755 UTC 8 May 1991 electric field profiles. Solid line at approximately -1.3°C delineates the lower charge transition level common to all electric field profiles in this study (see text). (Adapted from Hallett, 1984).

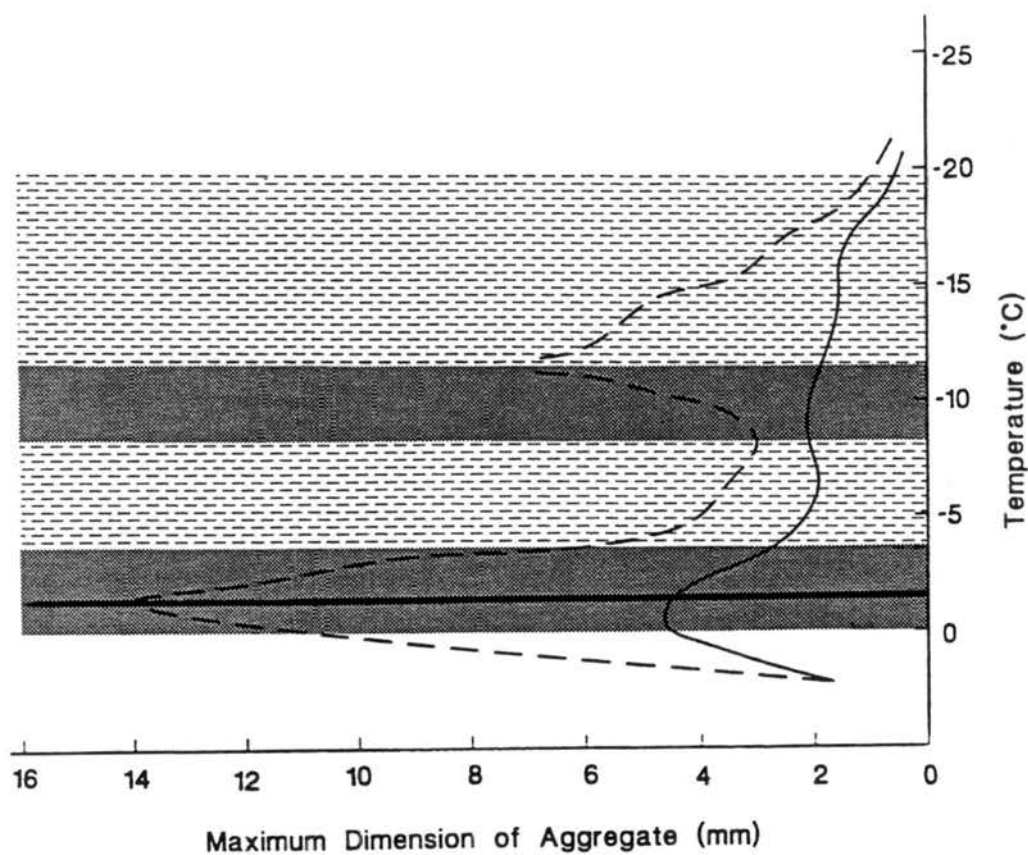


Fig. 4.4. Charge layers from symmetric MCS overlaid on figure depicting mean and maximum aggregate sizes. The solid curve represents the median of observed aggregate sizes and the dashed line the maximum of observed aggregate sizes. Positive charge layers are depicted by dark shading, negative charge layers by light shading. Solid line at approximately -1.3°C depicts the lower charge transition level common to all electric field profiles in this study (see text). (Adapted from Hobbs et al., 1974, and Young, 1993).

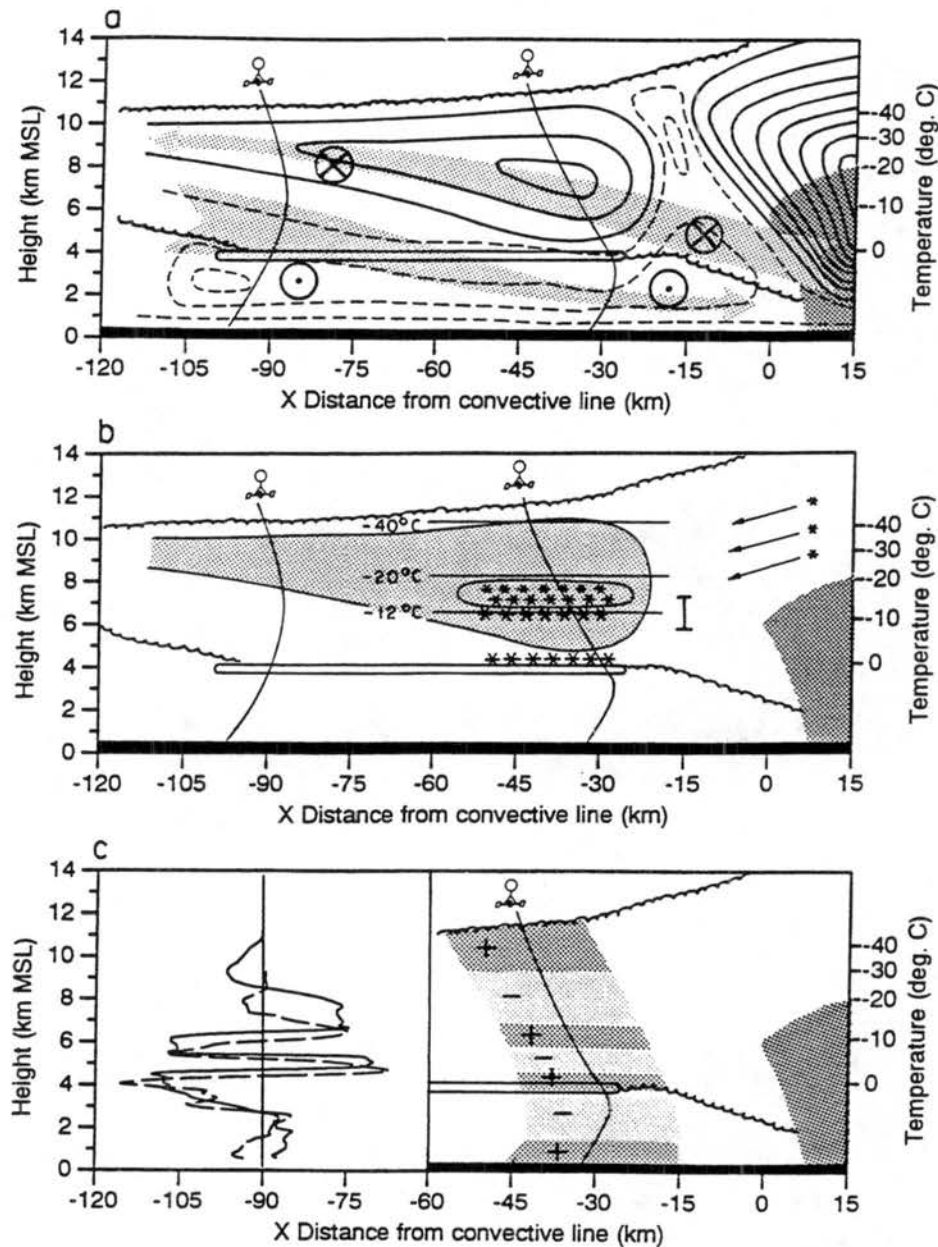


Fig. 4.5. Conceptual model depicting (a) kinematic, (b) microphysical, and (c) electrical structure of symmetric MCS. In (a), light shading indicates line-normal, system-relative airflow while line-parallel, system-relative airflow into and out of the of the cross-section are indicated by ($\otimes$) and ($\odot$), respectively. Solid (dashed) contours represent location and relative strength of updraft (downdraft). In (b), light shading indicates location of mesoscale updraft from part (a). Locations of the aggregation regions (cf. Fig. 4.4) are represented by the (*) while the (I) symbol indicates the -8 to -16°C layer over which P-3 LWCs were averaged. In (c), dark shading represents positive charge layers and light shading negative charge layers (composited from the 0549 and 0705 UTC 24 May 1991 electric field profiles on the left). The system-relative trajectories of the three EFM flights on 24 May 1991 are shown in parts (a), (b), and (c). The 0549 and 0705 UTC trajectories, which were launched in similar system-relative locations, are overlaid and form the right trajectory. The system-relative 0712 UTC trajectory is on the left.

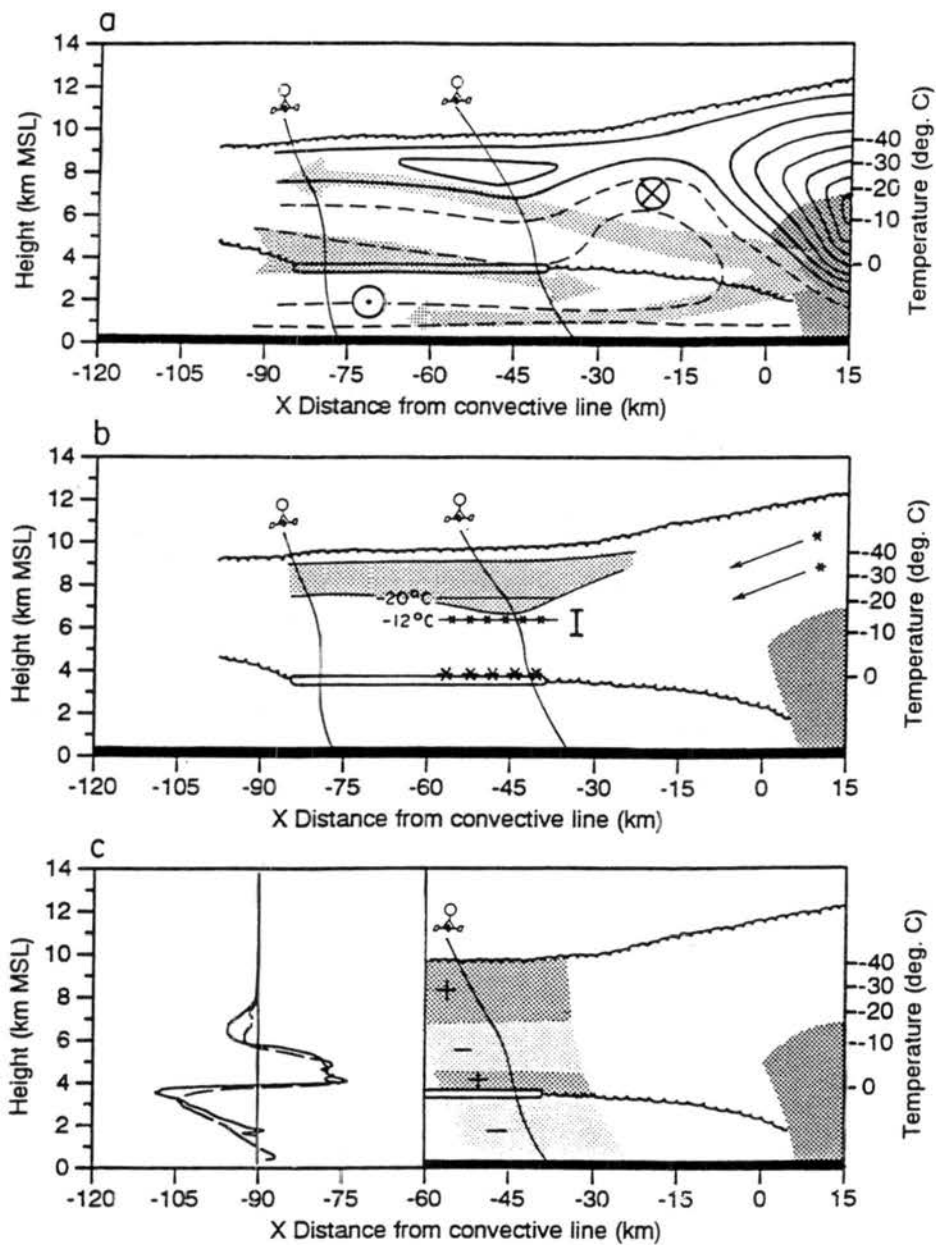


Fig. 4.6. Same as Fig. 4.5, except for asymmetric MCS. The system-relative trajectories of the two EFM flights on 8 May 1991 (0659, and 0755 UTC) are shown in parts (a), (b), and (c).

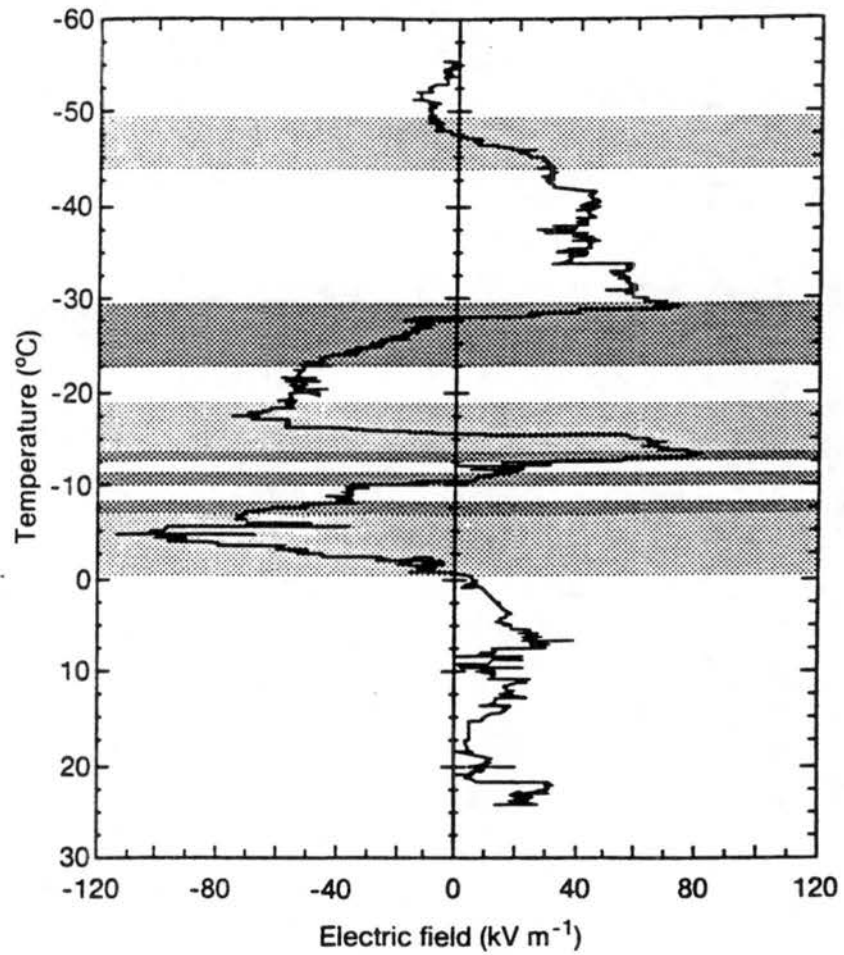


Fig. 4.7. Electric field plotted with respect to temperature for the 18 June 1987 transition zone electric field profile. Dark shading delineates layers of positive charge while light shading delineates layers of negative charge. (From Schuur et al., 1991).

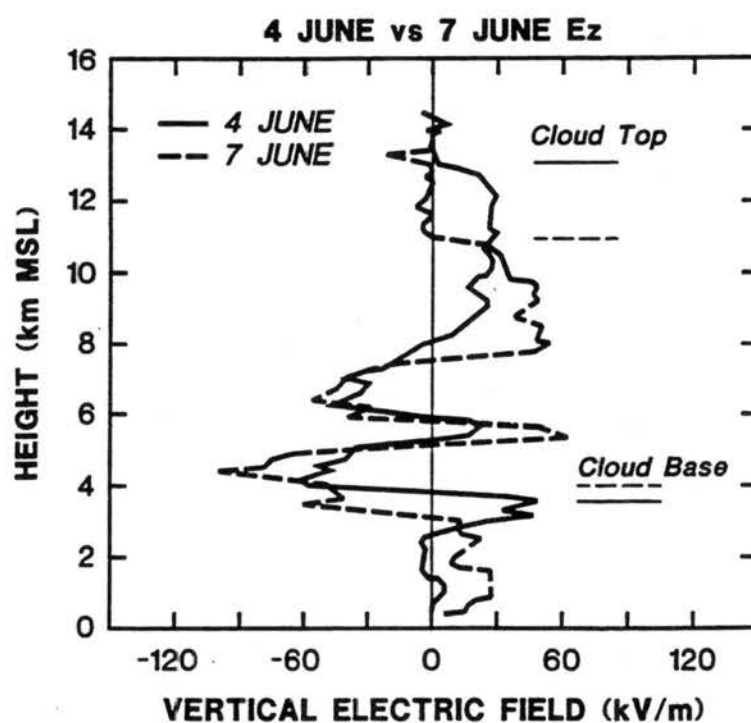


Fig. 4.8. Vertical electric field profiles obtained in the MCS stratiform regions of the 4 June 1989 (solid line) and 7 June 1989 (dashed line). (From Hunter et al., 1992).

CHAPTER 5

THE NUMERICAL MODEL

Over the years, several observational studies have investigated the charge structure and evolution of MCSs. From these studies, two specific hypotheses to explain stratiform charge structure have emerged: 1) advection of charge from the convective line, and 2) local charge generation resulting from microphysical processes within the stratiform region. The primary goal of the modeling component of this study is to examine and compare the relative contributions from each of these charge sources. The model we use is a two-dimensional, bulk microphysical model developed by Gallus (1993), which has been modified to include a variety of electrification processes. To better facilitate the investigation of charge advection, the model is configured to include a "buffer zone" within which the convective microphysical and electrical charge profiles are parameterized for each model run, thereby eliminating complexities and expense that would be required to explicitly simulate cellular convection. In the stratiform precipitation region, we focus on non-inductive charging mechanisms, which numerous laboratory studies (e.g., Takahashi, 1978; Saunders et al., 1991, and many others) have shown to be an effective means of transferring charge during ice/ice collisions in mixed-phase environments (cf. Chapter 1). While several modeling studies (e.g., Helsdon and Farley, 1987; Ziegler et al., 1986, 1991; Norville et al., 1991; Randell et al., 1992; Solomon and Baker, 1994; Ziegler and MacGorman, 1994) have examined the effectiveness of non-inductive charge transfer in convective cells, only one so far has addressed non-inductive charge transfer in the low-liquid water content clouds (Rutledge et al., 1990). In this study, we refer to the non-inductive transfer

of charge during ice-ice collisions, as described by the aforementioned laboratory studies, as the non-inductive charging mechanisms. Other microphysical charging mechanisms (e.g., Dong and Hallett, 1992) are, in the true sense of the word, also non-inductive in that they do not require the presence of electric fields for charging to occur. In order to avoid confusion, however, these other mechanisms will be referred to by alternative names.

5.1 Model description

The dynamic and microphysical framework of the model is described in detail by Gallus (1993, hereafter referred to as G93) and Gallus and Johnson (1995a, hereafter referred to as GJ95). The model is two-dimensional (aligned normal to the squall line in the x-z plane) and designed for specific application to the stratiform region of linear MCSs. The convective line is parameterized at the model's right boundary. This allows the model to be run with a much coarser horizontal grid spacing ($\Delta x=5000$ m, $\Delta z=400$ m) and larger time step ($\Delta t=15$ s) than would be possible if the convective line were explicitly modeled, thereby saving on computational costs and permitting the use of more detailed microphysics. This model framework is depicted conceptually by Fig. 5.1, where the "buffer zone" (consisting of 5 grid points on the right side of the grid) represents the region wherein the convective line's heating, microphysical, and charge profiles are specified. GJ95 found that including convective heating in model simulations was necessary to develop mesoscale motions similar to those typically observed in symmetric MCSs (i.e., cf. Fig. 3.9). In this study, after a comparison with previous work (e.g., Braun and Houze, 1996) and many model sensitivity runs, we have elected to use a convective heating profile that has a peak heating rate that is slightly larger (20 K h^{-1}) and located lower (at ~ 5 km MSL) than that used by GJ95. This heating profile is depicted by Fig. 5.2. The convective (i.e., buffer zone) microphysical and electrical initialization profiles are addressed later.

In general, the equations governing the model dynamics consist of a vorticity equation and a density weighted stream function that are solved at every time step, in the x-z plane, to obtain the two-dimensional (u and w) wind fields. Variations in the line-parallel direction are ignored. In addition, a continuity equation for potential temperature, which includes a source/sink term for diabatic heating resulting from microphysical processes (see Appendix A), is included in the model formulation. Leapfrog finite differencing is used and spatial derivatives are represented with centered differences, except at the model boundaries where one-sided differences are used. The output is filtered (e.g., Asselin, 1972) at each time step to suppress the time mode splitting that is commonly associated with the leapfrog differencing scheme. Subgrid-scale turbulence is parameterized as in Orlanski and Ross (1977) and Szeto et al. (1988). The simulations presented in this study are run on a 65×41 (320 km by 16 km) grid with the lowest grid level assumed to be at a height of 400 m MSL, which is typical of central Oklahoma. While this grid size has proved sufficient to simulate most aspects of MCS kinematic and microphysical evolution (e.g., GJ95), it is too small to contain the entire stratiform cloud region, especially at the latter stages of the simulation. Thus, open radiative (i.e., wave-permeable) boundary conditions, following Orlanski (1976), are used at the model's vertical boundaries. In addition, Newtonian damping (i.e., a spongy layer) is applied to the upper three levels of the model to absorb vertically propagating gravity waves (e.g., Kreitzburg and Perkey, 1977) and, at the lower boundary, rain is assumed to pass through the surface. Additional aspects of the model dynamics are discussed in detail by G93 and GJ95.

In general, the temperature and moisture initialization fields used in G93 (see their Figs. 3.2 and 3.3) have been adjusted slightly to better reflect the cooler environmental conditions on 24 May 1991 (while maintaining the same lapse rate used in the GJ95 simulations). The temperature exhibits a weak surface cold pool within 75 km of the convective line but is

horizontally homogeneous above the melting level. The relative humidity exhibits a gradual moisture gradient immediately behind the convective line which, in the absence of other processes, results in a buoyancy gradient that is capable of producing a weak circulation. The model winds are initialized with a horizontally homogeneous front-to-rear directed flow at all levels, with one wind speed maximum near the surface and a second, stronger one aloft. Once model integration begins, this results in hydrometeors being advected rearward and eventually, through dynamical feedbacks, the evolution of a two-dimensional mesoscale circulation that is quite similar to that exhibited by typical symmetric MCSs (e.g., G93, GJ95).

Given the difficulties of obtaining the high resolution observational data needed to initialize a mesoscale model, it is not a goal of this study to model a specific MCS. Rather, we attempt to model the basic precipitation and kinematic features exhibited by most symmetric MCSs (i.e., see Fig. 1.2) and then, after assuring its concomitant microphysical structure is as realistic as possible, examining its electrical evolution (based on the electrification parameterizations presented later). Since the model is two-dimensional, we only address symmetric MCSs in this study. Based on sensitivity tests presented in Chapter 7, however, we will speculate on how asymmetric MCS kinematic structure may be related to symmetric MCS kinematic structure.

5.2 Model microphysics

Laboratory studies of microphysical charging (reviewed in Chapter 1) indicate that the polarity and magnitude of charge transfer is extremely sensitive to a variety of environmental and microphysical factors. Thus, particular care was taken to assure the model microphysics were as realistic as possible. An initial examination of the model results from GJ95, however, revealed that, while the model simulations were excellent in reproducing higher-moment microphysical characteristics such as rainfall and reflectivity (both of which are dominated by the largest hydrometeors in a distribution), they were relatively poor job in reproducing microphysical

characteristics that are apparently important for charging (i.e., cloud liquid water content and ice concentrations). As a result, significant improvements were made to the model microphysics. Most notable was the addition of a prognostic equation for the number concentration of cloud ice, to better account for the collisions that result in non-inductive transfer of charge. Additionally, several modifications were made to mixing ratio parameterizations to improve the prediction of cloud water. In this chapter, we describe the microphysical framework of the model, as presented in GJ95. Furthermore, we describe the modifications that have been made to the model microphysical package to improve upon the variables necessary to predict microphysical charging processes (e.g., cloud ice concentrations and cloud LWCs). Other modifications are addressed in more detail in Appendix A.

5.2.1 Mixing ratios

As in many previous modeling studies (e.g., Lin et al., 1983; Rutledge and Hobbs, 1983, 1984; Lord et al., 1984, and others), precipitation hydrometeor mixing ratios in this study are represented by an exponential distribution, that is

$$n_x(D_x) = N_{ox} \exp(-\lambda_x D_x) \quad (5.1)$$

where x represents the precipitation hydrometeor type, λ_x the distribution slope, and N_{ox} the slope intercept. The slope of each distribution, in turn, is determined by multiplying Eqn. 5.1 by particle mass, integrating over all diameters, and equating the result to the appropriate water content, yielding

$$\lambda_x = \left(\frac{\alpha \pi \rho_L N_{ox}}{r_x} \right)^{0.25} \quad (5.2)$$

where α is the specific volume of air. In this study, as in GJ95, this derivation includes the modifications suggested by Potter (1991). That is, rather than using hydrometeor densities of the

precipitation classes, the density of liquid water is used to compute hydrometeor slopes. This corrects an error that resulted from a misinterpretation of data from Gunn and Marshall (1958) by previous modeling studies. Further corrections based on Potter (1991) are made to the hydrometeor fallspeed coefficients and are listed in Appendix A.

Earlier versions of this model (e.g., GJ95) used a three-class ice scheme that included graupel, snow, and cloud ice. Since stratiform liquid water contents are generally quite low ($\sim 0.1\text{--}0.3\text{ g m}^{-3}$) and in-situ microphysical observations indicate that graupel are rather uncommon in MCS stratiform clouds (e.g., Yeh et al., 1991), we eliminate the higher density graupel field in this study and run the model with two ice hydrometeor types: snow, and cloud ice. Since N_{∞} is held constant in this study, this additional mass in the snow field effectively increases the concentration of hydrometeors at the large end of the spectrum (as well as the distribution mean diameter, which is inversely related to the profile's slope), thereby creating a hybrid snow/aggregate field. This modification of the model microphysics also greatly simplifies the cloud electrification parameterizations described in Section 5.3.

With this simplification in mind, we now discuss the initialization of the model's buffer zone region. GJ95 used model output from a one-dimensional, time dependent numerical model (e.g., Ferrier and Houze, 1989) to initialize the model's buffer zone region. In this study, we use a modification of that hydrometeor profile. First, in keeping with the simplification discussed above, the mass from the graupel category was added to the snow category. Heights of the snow/aggregate, cloud ice, and cloud water categories were then adjusted with respect to height (temperature) to better reflect the thermodynamic and radar observations on 24 May 1991. Finally, cloud ice concentrations of 300 l^{-1} , as measured by the P-3 aircraft during transition zone flight legs, were used to introduce ice into the mixed-phase region (-5 to -20°C) of the initialization profile. The resultant buffer zone hydrometeor profile for the snow/aggregate,

cloud ice (assuming a low-level concentration of 300 l^{-1} and ice size of $\sim 250 \text{ }\mu\text{m}$), and cloud water categories are depicted in Fig. 5.3. In some of the non-inductive electrification simulations presented later, charge transfer is dependent on the size of ice. Therefore, we later investigate the sensitivity of using several different ice sizes (and therefore different mixed-phase region ice mixing ratios) in our advection profile.

With this two-class ice scheme, five hydrometeor continuity equations are solved using the leapfrog finite differencing scheme. For water vapor, cloud water, and cloud ice, (each assumed to be monodisperse), these equations follow the form

$$\frac{\partial r_x}{\partial t} = -u \frac{\partial r_x}{\partial x} - w \frac{\partial r_x}{\partial z} + \nabla \cdot K_m \nabla r_x + \left(\frac{dr_x}{dt} \right)_s \quad (5.3)$$

while for the precipitating snow/aggregate and rain fields, these equations are represented as

$$\frac{\partial r_x}{\partial t} = -u \frac{\partial r_x}{\partial x} - w \frac{\partial r_x}{\partial z} - \frac{1}{\rho} \frac{\partial}{\partial z} (\rho \bar{V} r_x) + \nabla \cdot K_m \nabla r_x + \left(\frac{dr_x}{dt} \right)_s \quad (5.4)$$

where the additional term present on the right hand side of equation 5.4 represents a hydrometeor sedimentation term. Note that the inclusion of the turbulent diffusion term in both of these equations represents an improvement over earlier versions of this model. In this study, for consistency, the eddy exchange coefficient (K_m) is parameterized in the same manner as the horizontal and vertical eddy diffusivities (see G93). Model test runs investigating the effect of the addition of turbulent diffusion, however, suggest that it has only minor effects on the resultant hydrometeor distributions. As in GJ95, the advection terms in Eqns. 5.3 and 5.4 are computed using the positive definite scheme of Bott (1989) to avoid the problem of negative hydrometeor values that occasionally arise with other advection schemes. Additional details regarding the microphysical parameterizations, including modifications made to improve upon the prediction of cloud water and cloud ice concentrations, are provided in Appendix A.

5.2.2 Cloud ice concentrations

5.2.2.1 Prognostic equation for cloud ice concentration

In order to accurately predict particle collisions which are highly coupled to non-inductive charging rates, a predictive equation for cloud ice concentrations was added to the model. This continuity equation follows the form

$$\frac{dN_i}{dt} = -u \frac{\partial N_i}{\partial x} - w \frac{\partial N_i}{\partial z} + \frac{\partial(N_i V_i)}{\partial z} + \nabla \cdot K_m \nabla N_i + \left(\frac{dN_i}{dt} \right)_s \quad (5.5)$$

where the terms on the right hand side are analogous to those described for Eqn. 5.4. The source/sink term is represented by

$$\begin{aligned} \left(\frac{dN_i}{dt} \right)_s = & \frac{1}{M_o} [PINT] + \left(\frac{N_i}{q_i} \right) \alpha [-PSACI - PRACI - PSMLTI - PCONV] \\ & + \frac{1}{m_{frag}} [PFRAG] \end{aligned} \quad (5.6)$$

Here, PINT, PSACI, PRACI, PSMLTI, and PCONV represent source and sink terms for the ice mixing ratio corresponding to initiation, accretion, melting, and autoconversion processes (see Appendix A), M_o the mass of an initiated ice crystal (1×10^{-12} kg), and $(N_i / q_i) \alpha$ the inverse of the average ice crystal mass. The PFRAG term represents a source of cloud ice due to ice fragmentation processes from the snow category. The assumed mass of each ice fragment (1×10^{-10} kg) is represented by m_{frag} .

As discussed in the next section, laboratory studies have indicated that non-inductive charging is dependent on collision impact velocity and cloud ice size. Therefore additional modifications have been made to the ice microphysics to predict these two variables. Average ice mass is first determined at each grid point from the predicted cloud ice mixing ratio and ice concentration. Using a mass-size relationship for hexagonal plates (from Locatelli and Hobbs,

1974), which is the same habit type used for depositional growth and autoconversion parameterizations in the model, a mean diameter is then determined. This diameter is then used to compute a mean ice fallspeed according to Kajikawa, 1972 (see Fig. 5.4). The maximum ice size in the model is constrained by the ice-to-snow autoconversion threshold, which is typically set to 6×10^{-10} kg (corresponding to a maximum ice size of 400 μm and fall speed of approximately 20 cm s^{-1}). The sensitivity of model electrification to factors that effect ice size are addressed in Chapter 7.

5.2.2.2 Ice initiation

The ice initiation scheme used in previous versions of this model inaccurately accounted for the amount of pre-existing ice at each grid location. Furthermore, it used the somewhat outdated parameterization of Fletcher (1962), which is known to overpredict ice initiation for $T_c < -40^\circ\text{C}$ and underpredict for $T_c > -20^\circ\text{C}$. Therefore, ice initiation in the model was modified to follow the deposition/condensation freezing scheme presented by Ferrier (1994). In this scheme, which follows the work of Cotton et al. (1986) and Murakami (1990), the more recent laboratory work of Huffman and Vali (1973) and Meyers et al. (1992) is utilized to develop a parameterization that takes ice saturation conditions into consideration. That is

$$N_{\text{INT}} = \frac{P_{\text{INT}}}{M_o} \quad (5.7)$$

$$P_{\text{INT}} = \alpha \max(0, w) \frac{\partial n_c}{\partial z} \cdot M_o \quad (5.8)$$

$$n_c = n_{c_1} \text{ if } T_c \geq -5^\circ\text{C} \quad (5.9)$$

$$n_c = n_{c_2} \text{ if } T_c \leq -5^\circ\text{C} \quad (5.10)$$

$$n_{c_1} = n_{\infty_1} \left[\frac{(r_v - r_{is})}{(r_{ws} - r_{is})} \right]^{\alpha_1} \exp(-\beta_1 T_c) \quad (5.11)$$

$$n_{c_2} = n_{\text{co}_2} \exp \left[\alpha_2 \left(\frac{r_v}{r_{is}} - 1 \right) - \beta_2 \right] \quad (5.12)$$

where w is the vertical velocity, r_{is} and r_{ws} the saturation mixing ratios with respect to ice and water, respectively, $n_{\text{co}_1} = 50 \text{ m}^{-3}$ and $n_{\text{co}_2} = 1000 \text{ m}^{-3}$ (so that $n_{c_1} = n_{c_2}$ at $T_c = -5^\circ\text{C}$), $\alpha_1 = 4.5$, $\beta_1 = 0.6 \text{ K}^{-1}$ (from Cotton et al., 1986), and $\alpha_2 = 12.96$, $\beta_2 = 0.639$ (from Meyers et al., 1992). The Meyers et al. data was derived from measurements of crystal concentrations in continuous flow devices at temperatures warmer than -25°C . The lack of data at colder temperatures points to an obvious need for more experimental data for temperatures colder than -25°C . Nevertheless, a comparison of this parameterization to that of Fletcher (1962) for a temperature of -15°C shows an increase of ice initiation from 81 m^{-3} to 3687 m^{-3} with the current scheme in the region where EFM observations indicate high charge densities.

5.2.2.3 Ice multiplication

Even with the improved ice initiation scheme described above, ice concentrations generated by the model were considerably lower than those indicated by the observations of Willis and Heymsfield (1989). In that study, in-situ aircraft observations of cloud ice in a midlatitude MCS revealed concentrations in excess of 100 l^{-1} ($1 \times 10^5 \text{ m}^{-3}$) at locations immediately above the 0°C isotherm. Citing a lack of sufficient cloud liquid water in this region, they noted that the Hallett-Mossop rime splintering mechanism (e.g., Hallett and Mossop, 1974) could not account for these observations and suggested that a fracturing mechanism was responsible for the higher ice concentrations.

Unfortunately, the mechanism by which ice fragmentation occurs is a difficult subject to address. Vardiman (1978) investigated the fracturing of various ice species in a laboratory by examining ice fragment generation that occurred when they were collided with a stationary plate.

Oraltay and Hallett (1989) suggested a mechanism whereby ice fragments are generated during the evaporation process. Other researchers (e.g., Crowther and Saunders, 1973) have noted an increase in ice fragmentation in the presence of high electric fields. Yet, no studies have offered sufficient qualitative data to allow for the development of a physically based model parameterization. Nevertheless, the necessity of accounting for ice concentrations in order to accurately represent collisional charging processes stresses the importance of including such a mechanism in this model. Therefore, a two-step procedure based both on theoretical considerations (primarily the work of Vardiman, 1978) and the observations of Willis and Heymsfield was developed to generate ice fragments in the model.

In the scheme, a mass-adjustment is first applied to the snow/aggregate field to generate ice fragments in regions where the mixing ratio exceeds a predefined, altitude-dependent mass threshold. To define this threshold, a characteristic mean diameter (D_m) for the snow/aggregate category, based on the in-situ observations and experimental model test runs was selected. Since $D_m = 1/\lambda$ for an exponential distribution, it is then possible to use Eqn. 5.2 to derive a fragmentation mass threshold for the snow mixing ratio, r_s , such that

$$r_{frag} = \alpha \pi \rho_L N_{os} D_m^4 \quad (5.13)$$

In this study, $D_m = 500 \mu m$. This results in $r_{frag} = 1.0 \text{ g kg}^{-1}$ at $0^\circ C$ and $r_{frag} = 1.28 \text{ g kg}^{-1}$ at $-15^\circ C$. Therefore, it can be seen that the effect of defining a mass threshold based on a mean diameter is to have to have increased fragment generation (for a given mixing ratio) at locations lower in the cloud. This is agreement with the observations of Willis and Heymsfield (1989), which suggest that fragment generation is maximized immediately above the $0^\circ C$ isotherm. Once the mass threshold is determined, the number of ice fragments generated is solved for as an autoconversion type process. That is, when the mass in the snow/aggregate category exceeds

r_{frag} , the slope of the distribution is adjusted (i.e., mass is removed) and fragments generated at the rate given by

$$\text{PFRAG} = k_{\text{frag}} (r_s - r_{\text{frag}}) \rho \quad (5.14)$$

where $k_{\text{frag}} = 0.5 \times 10^{-5} \text{ s}^{-1}$ is a rate coefficient similar to that used for the cloud water to rain and ice-to-snow autoconversion processes described by Lin et al. (1983) and Rutledge and Hobbs (1983). The number of ice fragments generated by this process is then determined by dividing PFRAG by a typical ice fragment mass, m_{frag} . As mentioned earlier, m_{frag} in this study is set to $1 \times 10^{-10} \text{ kg}$, which is based on typical ice fragment sizes in the Willis and Heymsfield observations.

While this mass adjustment procedure yields realistic ice concentrations, the pulsing of the convective input (discussed later) occasionally results in ice fragment generation patterns that are not stratified in a manner that is expected of the MCS stratiform region. Therefore a secondary procedure is invoked wherein the occasional gaps in the ice generation pattern resulting from the mass adjustment are filled with ice concentrations resembling the observational data. That is, using the assumed ice fragment ice mass, m_{frag} , mass is removed from the snow/aggregate category and distributed into stratified layers with concentrations resembling the profile used in Rutledge et al. (1990). Combined, these procedures give stratiform ice concentrations that depict the stratified nature of the MCS stratiform cloud but also give increased ice concentrations in regions where the snow/aggregate mean diameter exceeds observed values.

5.2.3 Other microphysical modifications

With the addition of a prognostic equation for ice concentration to the model it was necessary to modify the parameterizations for PDEPI (depositional growth of ice) and PCONV

(autoconversion of ice to snow) to better reflect their dependence on the newly predicted cloud ice concentrations. Since these parameterizations only effect the amount of water and ice mass transferred between categories and not the total ice concentration, their modifications are not addressed here. Other modifications to the ice microphysics tested during the course of this modeling study (in an attempt to improve upon the prediction of both the cloud ice concentration and the cloud water mixing ratio) include the development and testing of parameterizations for the Bergeron process (e.g., Lin et al., 1983) and heterogeneous and homogeneous freezing of cloud water (e.g., Wisner et al., 1972; Ferrier, 1994). A more detailed description of microphysical parameterizations is presented in Appendix A.

5.2.4 Cloud water

Initial model simulations indicated that far too much cloud water was being removed by collection, thereby resulting in cloud water contents that were as much as an order of magnitude smaller than indicated by the P-3 observations (e.g., 0.2-0.4 g m⁻³). Since the microphysical charging mechanisms discussed later are highly dependent on cloud liquid water content, improving upon the prediction of cloud water (while simultaneously predicting accurate cloud ice concentrations) became a major objective in the model's microphysical development. In an attempt to better understand the collection of cloud water by the snow/aggregate category, several sensitivity tests were run using microphysical code obtained from the Colorado State University Regional Atmospheric Mesoscale Model (CSU-RAMS, e.g., Flatau et al., 1989), which allowed for a detailed comparison of the stochastic and continuous collection equations.

For the values of N_{∞} commonly used in modeling studies (e.g., 3×10^6 through 2×10^7) it was found that explicitly solving the stochastic collection equation (e.g., Verlinde et al., 1990) rather than the continuous collection equation with an assumed mass weighted fallspeed relationship (e.g., Wisner et al., 1972) resulted in changes to the collection efficiency of -30% to +25% (for

assumed monodisperse droplet distributions with mean diameters ranging from 10 to 50 μm). The reduction in collection efficiency generally occurred at the smaller end of the droplet spectrum (i.e., 10-20 μm drops) with an increase in collection efficiency at the larger droplet sizes (i.e., 40-50 μm). The Willis and Heymsfield FSSP data suggest that a typical cloud water droplet size for an MCS stratiform region is approximately 15 to 20 μm (see their Fig. 10). Therefore, the stochastic collection equation results in a approximate reduction of 25% over that from the continuous collection equation.

More importantly, however, this exercise revealed several other factors that have large effects on collection rates, including the choice of N_∞ (slope intercept), choice of snow/aggregate fallspeed coefficients, and the choice of the distribution shape parameter. In fact, sensitivity tests indicated that even a small increase in the snow/aggregate distribution shape parameter (from an assumed exponential distribution to an assumed gamma distribution) resulted in a reduction in total collection cloud water by an order of magnitude. When these uncertainties in collection were combined with an assumed collection efficiency of 0.3 (Pitter and Pruppacher, 1974, see Fig. 5.5) to reduce the assumed collection efficiency in the model, predicted cloud water contents came into agreement with the in situ observations made by the P-3.

5.3 Model electrification

5.3.1 Continuity equations

The first step in adding electrification to the stratiform cloud model was to develop and add continuity equations and source/sink parameterizations for charge density. For the non-precipitating hydrometeor categories (e.g., cloud ice and cloud water) these equations follow the form,

$$\frac{\partial Q_x}{\partial t} = -u \frac{\partial Q_x}{\partial x} - w \frac{\partial Q_x}{\partial z} + \nabla \cdot K_m \nabla Q_x + \left(\frac{dQ_x}{dt} \right)_s \quad (5.15)$$

where Q is the charge density (presented here in units of nC m^{-3}) while for the precipitating categories (e.g., snow/aggregate and rain) the corresponding continuity equations are,

$$\frac{\partial Q_x}{\partial t} = -u \frac{\partial Q_x}{\partial x} - w \frac{\partial Q_x}{\partial z} - \frac{1}{\rho} \frac{\partial}{\partial z} (\rho \bar{V} Q_x) + \nabla \cdot \mathbf{K}_m \nabla Q_x + \left(\frac{dQ_x}{dt} \right)_s \quad (5.16)$$

In both of these equations the terms on the right hand side are analogous to the ones described for hydrometeor mixing ratios by eqns. 5.3 and 5.4. As in those equations, turbulent diffusion has been added to the charge continuity equations presented here in order to test the diffusion of charge (e.g. Rutledge et al., 1993; Stolzenburg et al., 1994). The last term in eqns. 5.15 and 5.16 represents the charge density source/sink term, i.e.,

$$\left(\frac{dQ_x}{dt} \right)_s = \left(\frac{dQ_x}{dt} \right)_{\text{nic}} + \left(\frac{dQ_x}{dt} \right)_{\text{qec}} + \left(\frac{dQ_x}{dt} \right)_{\text{qmelt}} + \left(\frac{dQ_x}{dt} \right)_{\text{int}} \quad (5.17)$$

where the subscript nic refers to non-inductive charging (e.g., Takahashi, 1978; Saunders et al., 1991), qec evaporation-condensation charging (Dong and Hallett, 1992), and qmelt melting charging (e.g., Drake, 1968). The last term on the right side of Eqn. 5.17 represents a source/sink term for charge exchange resulting from microphysical interactions (e.g., collection, autoconversion, fragmentation, sublimation, evaporation, and melting). In this model, these exchanges are addressed as in Ziegler et al. (1991), where the amount of charge transferred is linearly related to the amount of mass transferred in the microphysical interaction. For processes such as evaporation and sublimation, the charge is lost to the environment (since this model does not contain continuity equations for ions due to the computation constraints that would result from ion fluxes). In addition to the parameterizations listed above, parameterizations for inductive charging (e.g., Ziv and Levin, 1974; Kuettner et al., 1981) and screening layer formation (e.g., Ziegler and MacGorman, 1994) were developed and tested in the model. Since these two parameterizations are not discussed in detail, their parameterizations are not presented here.

Given the model grid spacing ($\Delta x=5000$ m, $\Delta z=400$ m) and time step ($\Delta t=15$ s), order of magnitude calculations using clear air ion mobilities given by Shreve (1970) indicate that ion flux speeds would surpass Courant-Friedrich-Levy (CFL) criteria and lead to mathematical instabilities with electric fields as small as 22.75 kV m^{-1} . Using the same grid spacing, an electric field of 150 kV m^{-1} (as indicated by the EFM soundings and generated by preliminary model runs) would require a time step of less than 2.5 s. Therefore, unlike the SDSMT model, charge continuity equations for clear air ion concentrations are not included. In their place, a rather simple screening layer parameterization similar to that described by Ziegler et al. (1991) is used (to be described later).

The charge continuity equations are solved in a similar manner to that described for the hydrometeor mixing ratio continuity equations. That is, a leap-frog forward in time integration scheme is used wherein the turbulent diffusion coefficient, K_m , is solved for in the same manner as that used in the mixing ratio continuity equations.

5.3.2 Electric fields

After the total charge is determined, Poisson's equation is solved to obtain the scalar electric potential. That is,

$$-\nabla^2 \Phi = \frac{Q_T}{\epsilon_0} \quad (5.18)$$

where Φ is the electric potential, Q_T the charge density, and ϵ_0 the permittivity of air. A general purpose Helmholtz solver in cartesian coordinates was obtained from the NETLIB software library and installed in the model to accomplish this task. After the electric potential is determined, the two-dimensional electric field vector, E , is solved for by taking the gradient of the potential

$$E = -\nabla \Phi \quad (5.19)$$

The magnitudes of the horizontal and vertical components of the electric field vector are respectively,

$$E_x = -\frac{\partial\Phi}{\partial x} \quad \text{and} \quad E_z = -\frac{\partial\Phi}{\partial z} \quad (5.20)$$

To solve Poisson's equation using the Helmholtz equation solver, assumptions must be made about the magnitude of either the electric potential or the electric field vector at the model boundaries. Generally, the assumption made is that the electric field at the model boundaries is zero. Since the entire model grid is filled with charge when the MCS reaches maturity, however, such an assumption is usually not justified. Therefore, before the equation solver is invoked at each time step, the model cloud grid (with horizontal and vertical dimensions of 320 and 16 km) is nested within a much larger grid (1000 km $\times$ 24 km). Charge densities at all grid points outside of the cloud grid are then set to zero. Therefore, electric fields at the boundaries of the grid over which Poisson's equation is solved (which are now separated from the cloud grid by over 300 km in the horizontal) may reasonably be set to zero. At the model lower boundary, the scalar potential is assumed to be equivalent to that from the grid level immediately above it. Sensitivity tests in which the cloud charge was nested within much larger charge free grids indicated that this nested grid size reasonably approximated the electric field structure. A schematic depicting this nested grid approach is shown by Fig. 5.6.

5.3.3 Charge advection

As noted earlier, the primary goal of the modeling component of this study is to examine and compare the relative contributions of 1) charge advection from the convective line, and 2) in situ charge generation within the stratiform precipitation region. Since relatively few observations of particle charge have been made in MCS convective cells, developing a charge initiation profile for the buffer zone region (cf. Fig. 5.1) is a difficult process. Therefore, we base our initiation

profile on conceptual models of thunderstorm charge structure (e.g., Simpson and Scrase, 1937; Williams, 1989), which suggest a basic thunderstorm charge structure of positive charge over negative charge (with the presence of an additional region of positive charge located near cloud base). Numerous laboratory studies (e.g., Takahashi, 1978; Jayaratne et al., 1983; Saunders and Jayaratne, 1986, and others) suggest that ice/ice collisions in high LWC environments result in a net positive charge transfer to the ice crystal, consistent with the charge structure model. Thus, in our charge initialization profile, positive charge is assigned to cloud ice while negative charge is assigned to the snow/aggregate category (with charge density magnitudes on the order of that observed in convective cells by Marshall and Rust, 1991). This initialization profile is depicted in Fig. 5.7, which shows the charge density assigned to each hydrometeor category at all levels in the convective cloud (i.e., buffer zone). When combined, the composite charge profile (Fig. 5.7b) exhibits a simple dipole structure, with the exception of a weak region of positive charge near cloud base. Peak charge density magnitudes in the initialization profile are approximately 3 nC m^{-3} and the primary charge transition occurs at approximately -20°C . With the exception of a negative charge near cloud top, this profile exhibits marked similarities to the composite MCS convective line electrical structure in Stolzenburg (1996).

5.3.4 Charging parameterizations

5.3.4.1 Non-inductive charging

As reviewed in Chapter 1, numerous laboratory studies have investigated the non-inductive charging process. In particular, studies of Takahashi (1978) and Saunders et al. (1991) have both shown promise in explaining at least some of the charge features observed in convective clouds. Yet, they also exhibit marked disagreement for some temperature/LWC regimes (e.g., Williams, 1989). Therefore, we here investigate data from both laboratory studies.

The parameterization for the non-inductive charging is developed here as in Rutledge et al. (1990). That is, using the continuous collection equation, the charge generation rate per unit volume can be written as,

$$\left(\frac{dQ}{dt}\right)_{\text{nic}} = \int_0^{\infty} \frac{\pi}{4} D_s^2 |\bar{V}_s - \bar{V}_i| E_{is} N_i N_s \delta q dD_s \quad (5.21)$$

where E_{is} is the separation efficiency, δq the charge separation per collision (discussed below), and N_i the cloud ice concentration (predicted by eqn. 5.5). Substituting for Eqn. 5.1 for N_s and integrating over the particle size distribution, we obtain the charge generation rate per unit volume, i.e.,

$$\left(\frac{dQ}{dt}\right)_{\text{nic}} = \frac{\pi}{4} |\bar{V}_s - \bar{V}_i| E_{is} N_i N_{os} \frac{\Gamma(3)}{\lambda_s^3} \delta q \quad (5.22)$$

where λ_s and N_{os} are the slope and intercept of the exponential snow/aggregate distribution, respectively, and Γ represents the Gamma function. For the Takahashi (1978) data set, δq at each grid point is then determined from a charge lookup table (cf. Fig. 1.6) developed by Randell et al. (1994). With the Saunders et al. (1991) data set, however, δq is dependent on both cloud ice size (D_i) and collision impact velocity (V) and is determined according to

$$\delta q = B D_i^a V^b q' \quad (5.23)$$

where q' is computed from Eqns. 2-8 and the coefficient B , a , and b obtained from Table 2 in Saunders et al. (1991), respectively. The prediction of ice size and fallspeed are discussed in Section 5.2.2.1. The sensitivity to the Saunders et al. (1991) non-inductive charging parameterization to factors that affect cloud ice size and fallspeed in the numerical model will be addressed in Chapter 7. Computer code to compute charge transfer using the Saunders et al. (1991) data was obtained from Dr. John Helsdon of the South Dakota School of Mines and Technology (SDSMT). Since the Saunders et al. data were only determined for $T < -7.5^\circ\text{C}$, the

SDSMT parameterization extrapolates those results to warmer temperatures. In agreement with Ziegler and MacGorman (1994), we use an effective collection efficiency of 0.5 in our Saunders et al. non-inductive charging simulations.

5.3.4.2 Evaporation-condensation charging

Though, by definition, the “depositional charging” mechanism proposed by Dong and Hallett (1992) is a non-inductive process (e.g., Stolzenburg et al., 1994), the mechanism of proposed charge transfer (see review in Chapter 1) is quite different than that of the collisional process described above. Therefore, we address this as a distinct category of in situ charging in our parameterization development. Stolzenburg et al. (1994) and Rutledge and Petersen (1994) both suggested that the evaporation-condensation charging mechanism might be responsible for at least some of the charge observed in MCS stratiform precipitation regions. Simple order of magnitude calculations conducted by Rutledge and Petersen (1994), however, suggested that charge density magnitudes generated by this mechanism are two orders of magnitude smaller than those observed in MCS stratiform regions. In this section, we develop a parameterization to more formally test this mechanism.

In their laboratory study, Dong and Hallett (1992) found a complicated charging behavior during evaporation and vapor growth of ice and cloud water. In particular, they found that the ice particle acquired positive charge during growth and negative charge during sublimation, except at temperatures between 0 and -4°C where the ice behaved like a cloud water droplet and acquired negative charge during growth. They quantified their laboratory results by producing a table that listed charging current as a function of ice/water surface area (A cm^{-2}) with respect to temperature ($^{\circ}\text{C}$). Therefore, our model parameterization approach was to determine the total

surface area associated with the model ice distributions and multiplying that area by the charging rate from Dong and Hallett (1992), that is

$$\sigma_x = \int_0^{\infty} \pi D_x^2 N(D_x) dD_x \quad (5.24)$$

$$\left(\frac{dQ_x}{dt} \right) = \sigma_x F_q(T_c) \quad (5.25)$$

where σ_x is the total surface area of the distribution (cloud ice or snow/aggregate categories) and $F_q(T_c)$ derived from fitting a third order polynomial to the data presented in Fig. 2 of Dong and Hallett (1992). The data from their Fig. 2, along with the best fit, third order polynomial, is depicted in Fig. 5.8, which shows that $F_q(T_c)$ can be approximated as

$$F_q(T_c) = a_{q3}T_c^3 + a_{q2}T_c^2 + a_{q1}T_c + a_{q0} \quad (0^\circ\text{C} \geq T_c \geq -20^\circ\text{C}) \quad (5.26)$$

where T_c is the temperature in $^\circ\text{C}$ and $a_{q3}=0.0008$, $a_{q2}=-0.0025$, $a_{q1}=-0.4477$, and $a_{q0}=-1.286$. The factor 10^{-16} converts the charging rate into units of A cm^{-2} . For the snow/aggregate distribution (assumed to be exponential), the total charging rate is then

$$\left(\frac{dQ}{dt} \right)_{q\text{dep}} = \frac{2\pi N_{os}}{\lambda_s^3} F_q(T_c) \quad (5.27)$$

and, for cloud ice (assumed to be monodisperse), the total charging rate is

$$\left(\frac{dQ}{dt} \right)_{q\text{dep}} = (\pi N_i)^{1/3} \left[\frac{6q_i \rho}{r_i} \right]^{2/3} F_q(T_c) \quad (5.28)$$

Experimental model runs using this parameterization indicate that this mechanism, on mesoscale time scales (i.e., approximately 1 hour), gives charge density magnitudes that are approximately two orders of magnitude less than those indicated by the EFM soundings. However, Dong and Hallett (1992) noted that the charging rates determined in their study (Fig. 5.8) were also highly dependent on numerous other environmental and microphysical factors (discussed later). Given

this, the applicability of this charging mechanism to the MCS stratiform region will be discussed in detail in Chapter 6 after a presentation of the model runs.

5.3.4.3 Melting charging

Numerous recent observational studies (e.g., Marshall and Rust, 1993; Stolzenburg et al., 1994; Shepherd et al., 1996) have suggested the “melting charging” mechanism as a possible source of charge generation near the melting level of MCS stratiform regions. With one such mechanism by Drake (1968), it is hypothesized that strong convection currents develop in the melt water shell surrounding a melting ice particle. The convection currents are then believed to carry bubbles from the embedded ice to the water surface where they burst, thereby releasing ions and resulting in a separation of charge. Drake quantified this process by developing a bulk charging rate per ice mass melted, which can be parameterized as

$$\left(\frac{dQ}{dt} \right)_{\text{qmelt}} = \frac{\text{PMELTS}}{\rho_L} Q_m \quad (5.29)$$

where PMELTS is the snow/aggregate category melting rate (see Appendix A) and Q_m the charging rate, where $Q_m = 6 \times 10^6$ esu per cubic meter of melt water (1×10^{-4} esu = 33 fC) was obtained from Drake (1968). It should be noted that the laboratory study of Drake (1968) primarily examined melting ice spheres. Therefore, in MCS stratiform regions, the applicability of this mechanism is unknown since the predominant ice hydrometeor type near the melting level of MCS stratiform regions are aggregates (which have a much different melting pattern than do quasi-spherical graupel). Nevertheless, given the abundance of recent observational studies that indicate intense charge transitions near the melting level of MCS stratiform regions (which lend support to the potential importance of this mechanism), we examine its relative contribution in more detail in Chapter 6.

5.4 Pulsing and data averaging

GJ95 demonstrated that pulsing the convective input (heat and hydrometeors) leads to more vigorous and realistic mesoscale circulations. Additionally, it simulates the growth and decay of convective cells along the convective line, as is commonly seen in both observations and modeling studies that explicitly simulate the convective line (e.g., Dudhia et al., 1987; Fovell and Ogura, 1988; LaFore and Moncrieff, 1989). Given this, we also pulse the convective heating, hydrometeor, and charge profiles with a few modifications to that used by GJ95. In that study, the amplitude of the pulse dropped to zero at the end of each 30 minute period. In the present study, we do not allow the pulsing amplitude to fall below 25% of the maximum amplitude used by GJ95. Thus, at no time during our simulations are hydrometeor and heat advection from the convective line completely turned off. In order to assure that the total water mass associated with each convective pulse remains unchanged from that used in GJ95, however, we have also slightly reduced the maximum amplitude of the pulse so that the total integrated water mass (over a pulse) agrees with the symmetric MCS water budget study presented by Gallus and Johnson (1991). As in GJ95, the simulations presented here are integrated to 7 hours, with the convective input gradually decreasing after five hours in order to simulate the convective line's dissipative stage. This also is slightly different from GJ95 who decreased the convective heating and hydrometeor advection at different rates.

These pulse amplitudes, over the 7 hour integration period, are plotted with respect to time in Fig. 5.9. Since the balloon-borne EFM measures an environment resulting from contributions from many convective cells (along the line) at various stages of their life cycles, we also average much of the data presented in this study over one hour time periods (i.e., two convective pulses) to assure that the data presented here are representative of mean MCS stratiform conditions. Finally, these averages are used to produce a charge budget analysis that compares the relative

contributions of charge advection from the convective line to in situ charge generation within the MCS stratiform precipitation region over one hour periods.

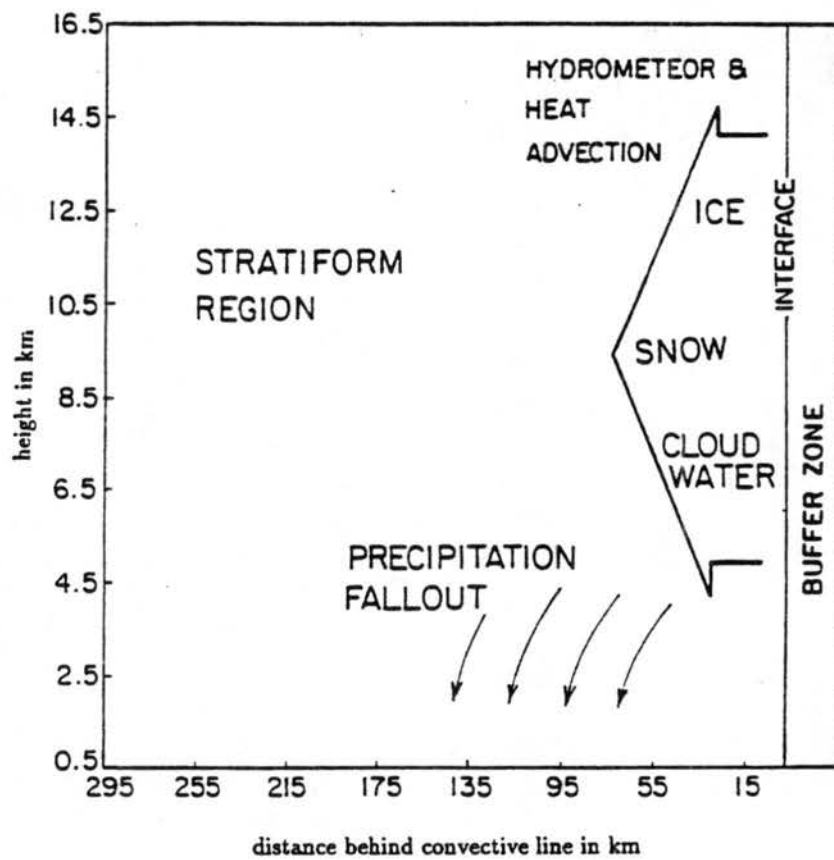


Fig. 5.1. Schematic of the model domain. Convective line hydrometeor, heating, and electrical profiles are specified within the buffer zone region. (Adapted from Gallus, 1993).

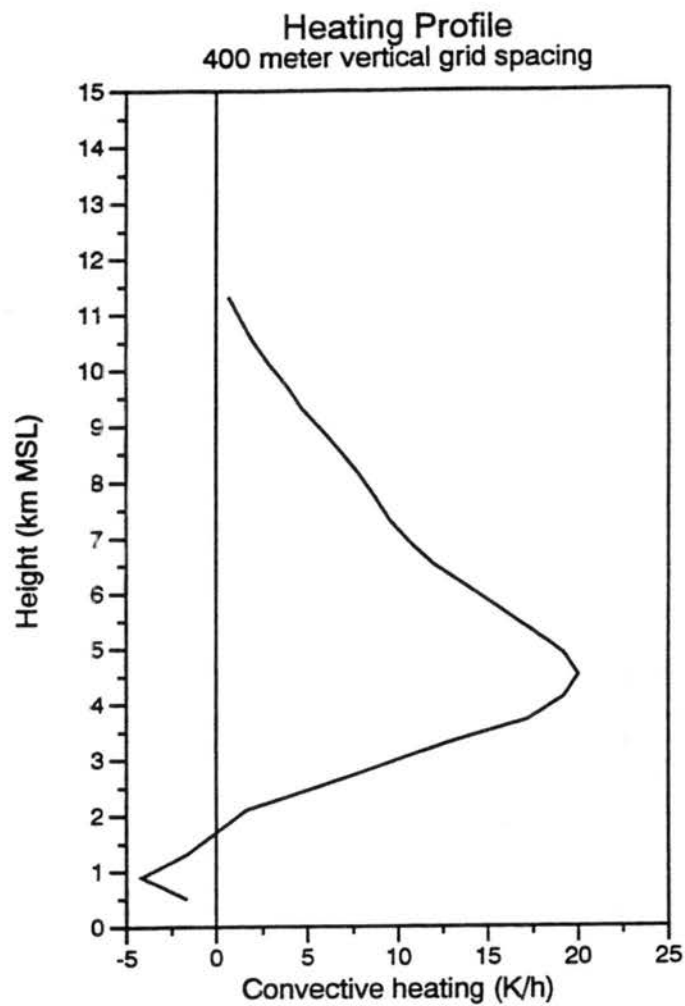


Fig. 5.2. Peak heating rate used to initialize the model buffer zone (cf. Fig. 5.1).

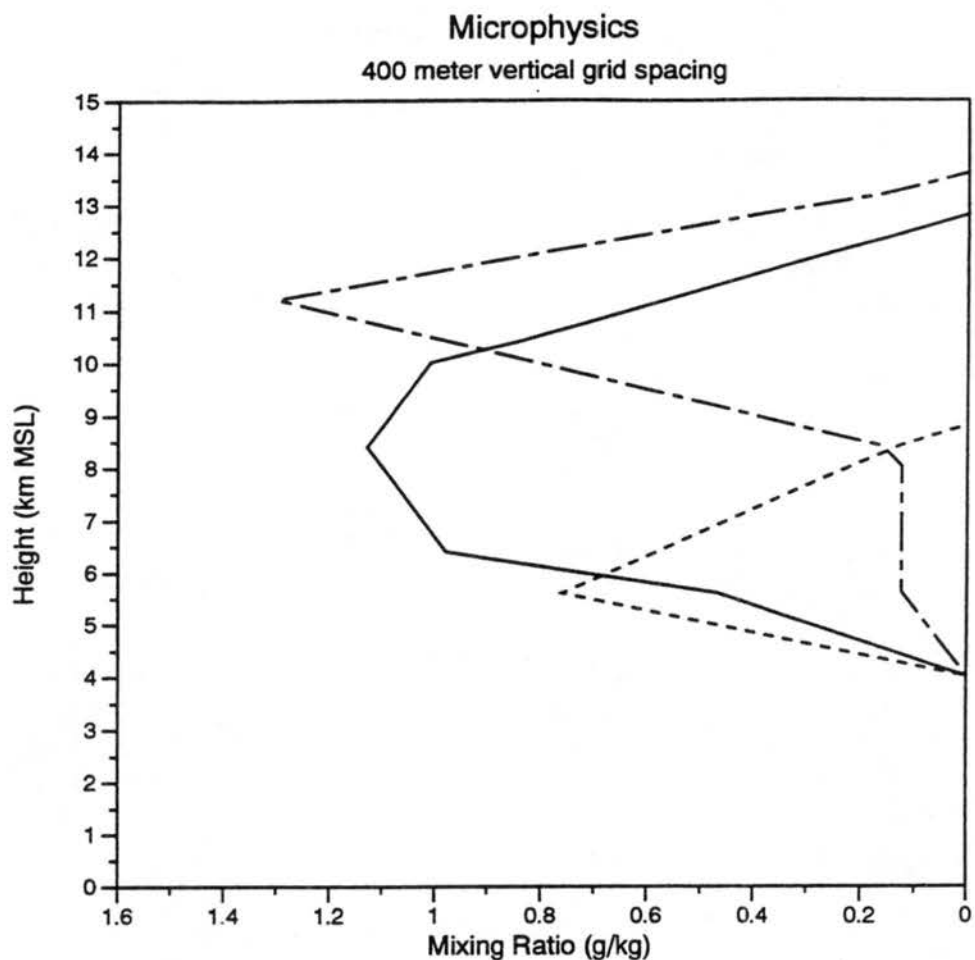


Fig. 5.3. Peak hydrometeor mixing ratios used to initialize model buffer zone (cf. Fig. 5.1). Hydrometeor profiles depicted are snow/aggregate (solid line), cloud ice (long and short dashed line), and cloud water (short dashed line). Ice mixing ratio depicted in mixed-phase region (below approximately 8.5 km MSL) represents mass associated with $250\text{ }\mu\text{m}$ crystals with a concentration of 300 l^{-1} (see text).

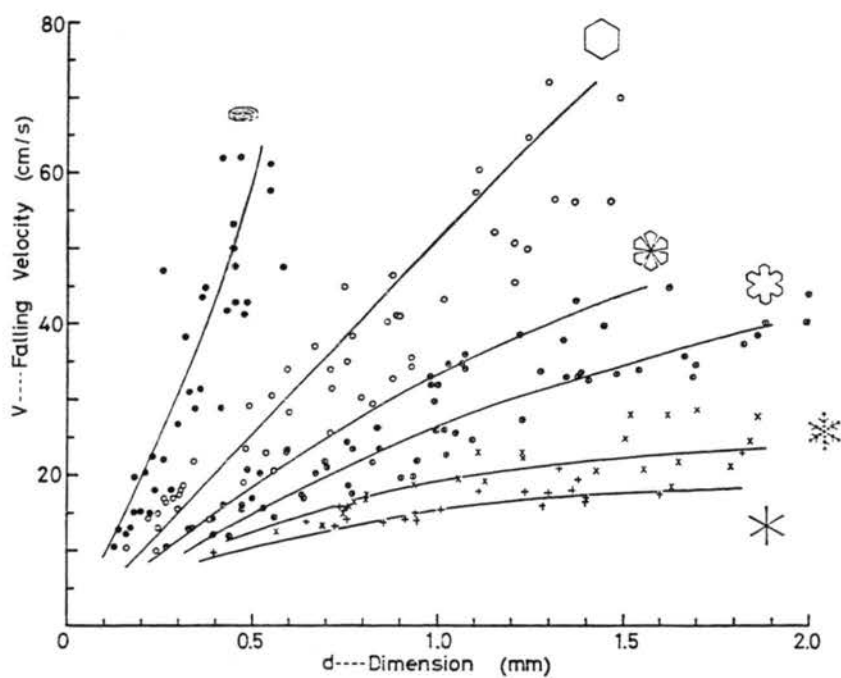


Fig. 5.4. Ice crystal fallspeeds from Kajikawa (1972). Fallspeeds profile used for cloud ice in this model is that for hexagonal plates (the second line from the left).

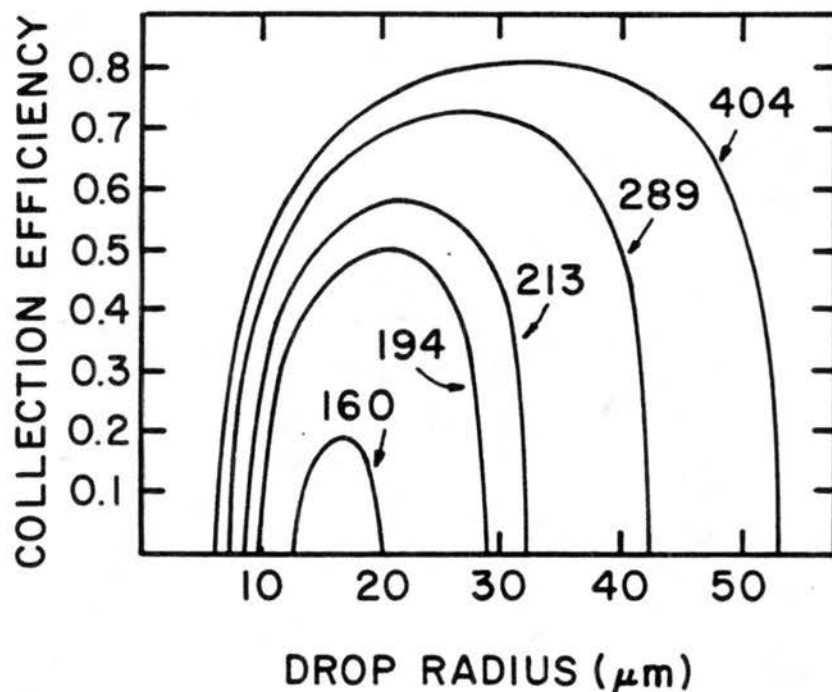


Fig. 5.5. Collection efficiencies for thin plates collecting water droplets based on numerical simulations by Pitter and Pruppacher (1974). Plate radii are 160 μm , 194 μm , 213 μm , 289 μm , and 404 μm , respectively. Comparing cloud droplet sizes (15 to 20 μm) from Willis and Heymsfield (1989) suggests a mean collection efficiency of approximately 0.3.

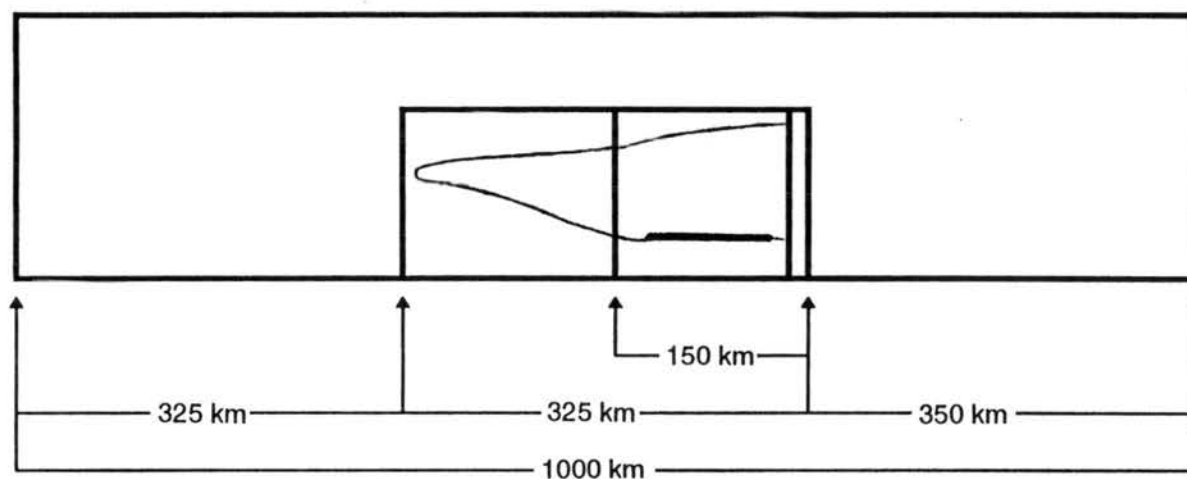


Fig. 5.6. Schematic showing grid sizes used in this modeling study. Model grid is nested in 1000×24 km grid to solve for electric field (see text for details).

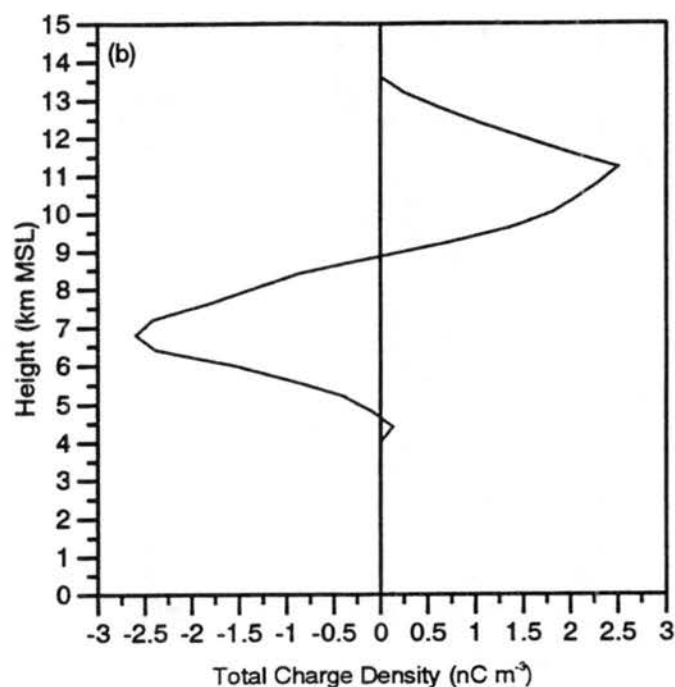
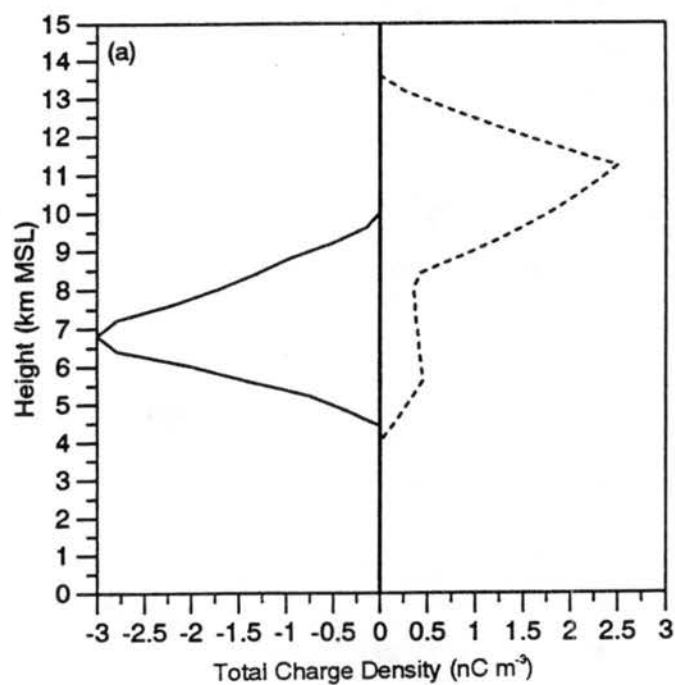


Fig. 5.7. Charge profiles used to initialize the buffer zone region (cf. Fig. 5.1). (a) Charge densities (nC m^{-3}) of snow (solid line) and cloud ice (dashed line). (b) Total charge density (nC m^{-3}).

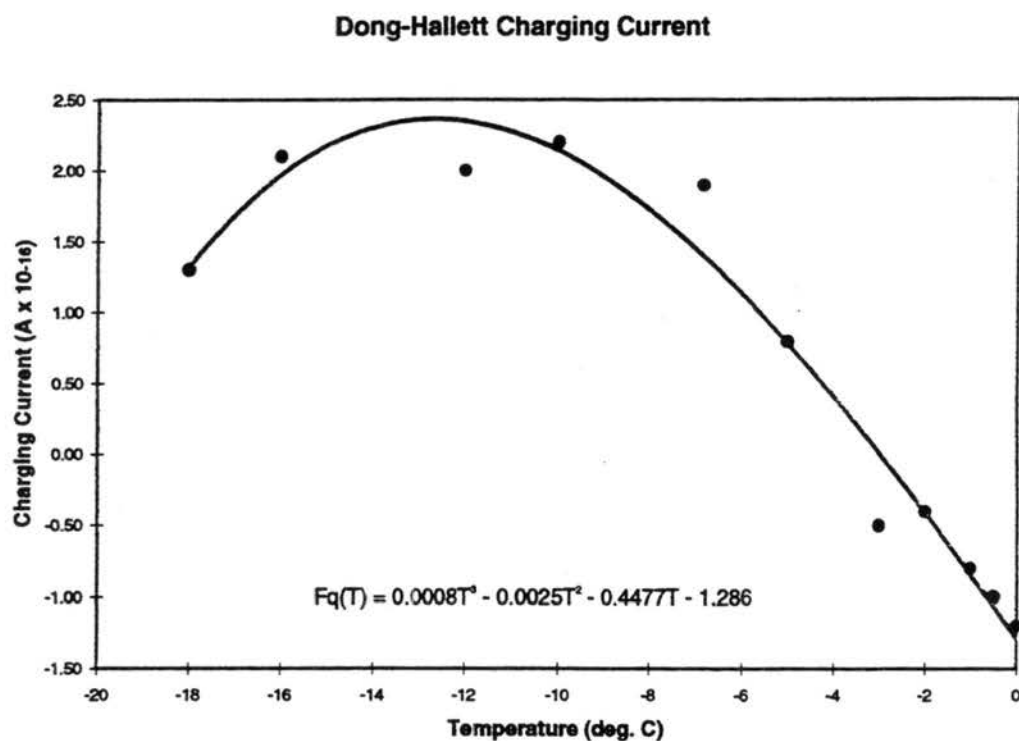


Fig. 5.8. Polynomial fit of charging current ($A \text{ cm}^{-2} \times 10^{-16}$) versus temperature ($^{\circ}\text{C}$) from the data of Dong and Hallett (1992).

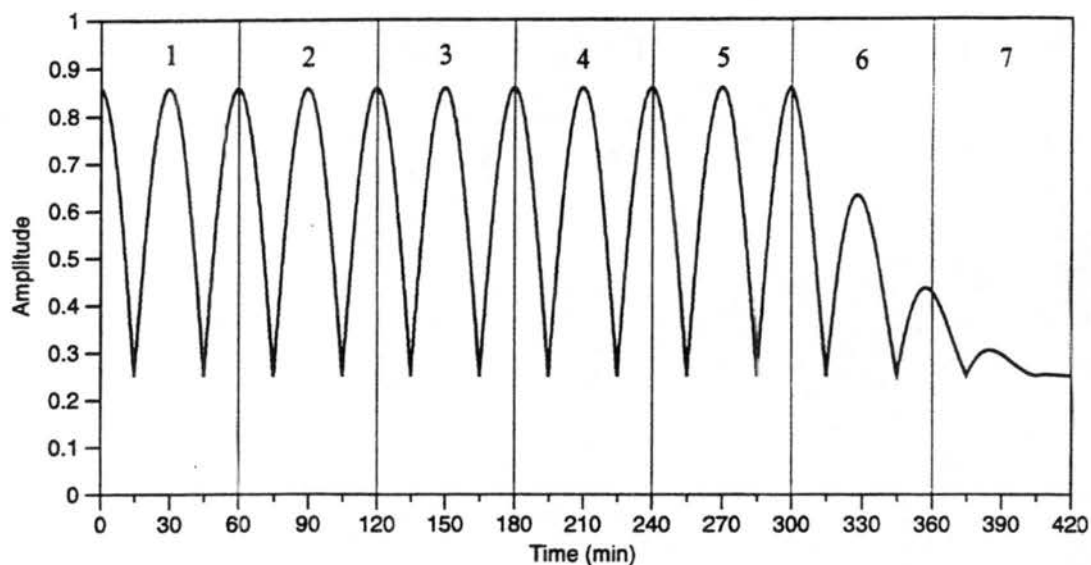


Fig. 5.9. Pulsing amplitude used for convective line hydrometeor, heating, and electrical profiles (cf. Figs. 5.2, 5.3 and 5.7). Maximum pulse amplitude has been reduced to agree with water budget study of Gallus and Johnson (1991). After five hours of integration, pulsing amplitude gradually decreases to simulate dissipation of convective line. Vertical lines indicate one hour periods over which output data are averaged for budget study.

CHAPTER 6

NUMERICAL SIMULATIONS

In this chapter, we present a series of model simulations that investigate MCS charging mechanisms addressed in Chapter 5. Particular emphasis is placed on understanding the relative contributions of 1) the advection of charge from the convective region, and 2) non-inductive charge transfer resulting from ice/ice collisions. Evaporation-condensation and melting charging are also discussed. Unless otherwise stated, all simulations use the same convective input (cf. Figs. 5.2, 5.3) and microphysical parameters specified in Appendix B. Thus, the output from the model runs may be easily intercompared to examine the relative importance of each charging mechanism. The sensitivity of the non-inductive charging process to the size of ice crystals advected from the convective line is tested by creating a budget analysis that quantifies its contributions for different ice sizes. Model sensitivity to other input and initialization parameters are presented in Chapter 7.

In this presentation, we show both instantaneous and averaged model output. The averaged output represents data averaged over two convective pulses (i.e., a one hour period) and is meant to depict the mean MCS stratiform conditions, which are likely the result of dynamic, microphysical, and electrical contributions from many convective cells at various stages of development, as sampled by the balloon-borne EFMs. Furthermore, the averaged model output offer a more effective means to compare the model output with the composited observational data sets presented in Chapters 3 and 4.

6.1 Kinematics and microphysics of the control run

The set of microphysical and convective region inputs used in all simulations was arrived at after numerous model sensitivity runs that were exhaustively compared to the observations presented in Chapters 3 and 4. The control run presented here represents the most realistic combined, kinematic, microphysical, and precipitation structure attainable with the model. Though we used the same model grid size employed by Gallus and Johnson (1995a), which has a horizontal dimension of 320 km, we only presented model output within 125 km of the convective region (i.e., buffer zone), since that is where the radar bright band develops and, furthermore, where all of our electric field profiles from EFM flights were obtained.

The model radar reflectivity, averaged over one hour periods to eliminate asymmetries resulting from the convective pulsing (cf. Fig. 5.9), is depicted by Fig. 6.1, which shows the evolution of the radar bright band over a 4 hour period. Since reflectivity is primarily dependent on the largest hydrometeors in a distribution, only the precipitation-sized hydrometeor categories (i.e., snow/aggregate and rain) have been used to compute the radar reflectivity; cloud ice and cloud water have been neglected by virtue of their small assumed sizes. By 240 minutes, the bright band has grown to 80 km in width and achieved maximum reflectivities of approximately 45 dBZ, both of which agree well with the MCS observations presented in Chapter 3. Fig. 6.1 also shows the evolution of a transition zone region immediately behind the convective line, though the width of the modeled transition zone is significantly less than that seen in the observations. Smull and Houze (1985) and Rutledge and Houze (1987) demonstrated that much of the transition zone reflectivity deficit can be explained by fallspeed sorting, whereby heavier hydrometeors fell out close to the convective line and lighter hydrometeors (i.e., snow and cloud ice) are carried rearward by the MCS's upper-level front-to-rear flow. Since we have eliminated the graupel category from the convective initialization profile and added its mass to the snow

category (cf. Fig. 5.3), it is possible that much of this underestimation is related to the use of a two-class ice scheme. In agreement with observations, the transition zone broadens considerably when the convective pulsing is turned off later in the model simulation (not shown).

The evolution of instantaneous and hourly averaged horizontal winds are depicted by Figs. 6.2 and 6.3, respectively. Within 60 minutes, the model develops a rear-to-front mesoscale flow at locations immediately below the melting level. At the same time, an acceleration of the front-to-rear flow occurs aloft, eventually resulting in a highly sheared region situated above the melting level. By 240 minutes, peak front-to-rear velocities are $>20 \text{ m s}^{-1}$ while peak rear-to-front velocities are approximately 10 m s^{-1} . Since the initialization wind field has front-to-rear directed flow at all levels, an examination of horizontal wind *perturbations* (not shown) indicate the development of a system-relative rear inflow jet that slopes upwards slightly towards the rear edge of the MCS, in agreement with observations of many MCSs (e.g., Smull and Houze, 1987b). Maximum horizontal velocities associated with this rear inflow jet also agree well with observations (cf. Fig. 3.9).

Corresponding instantaneous and hourly averaged vertical motions are depicted by Figs. 6.4 and 6.5, respectively. Peak upward motions generated by the model are approximately 50 cm s^{-1} (Fig. 6.4) though the mean vertical motion associated with the broad region of mesoscale ascent (Fig. 6.4b-d) is typically much weaker ($10\text{-}20 \text{ cm s}^{-1}$). Below the mesoscale updraft, an extensive region of mesoscale subsidence develops in response to melting and evaporative cooling, eventually broadening and achieving peak magnitudes of approximately 30 cm s^{-1} . These vertical motions are comparable to the flight leg averaged magnitudes presented in Figs. 3.5 and 3.7. This mean vertical motion structure is depicted better by the averaged data in Fig. 6.5, which effectively smoothes out any gravity waves that might exist in the instantaneous model output (e.g., LaFore and Moncrieff, 1989; Gallus, 1993).

Qualitatively, the hourly averaged vertical motions exhibit many similarities to the composited observations presented in Chapter 3 (e.g., cf. Fig. 3.8). As an example, the interface between the updraft and downdraft near the leading edge of the radar bright band is situated at temperatures between approximately -5 and -10 °C, in accord with observations. It is also interesting to note the region of deep descent that is located approximately 10 to 15 km behind the convective line from 120 through 240 minutes in the hourly averaged output (Fig. 6.5b-d). Though a deep transition zone downdraft in this region was indicated by our observations (Figs. 3.5-3.9), consistent with previous studies (Smull and Houze, 1987a, Biggerstaff and Houze, 1993), the transition zone downdraft generated by the model must be driven entirely by microphysical processes and not mechanically forced by diverging convective drafts.

We now discuss the evolution of the model microphysical structure. In particular we focus on the evolution of the snow/aggregate, cloud water, and cloud ice categories since they are the fields that are important towards understanding the in situ microphysical charging processes to be discussed later. The evolution of the instantaneous and hourly averaged snow/aggregate (hereafter referred to as snow) mixing ratios are depicted by Figs. 6.6 and 6.7, respectively. Snow advects into the model domain and moves rearward, gradually experiencing increased upward motion (cf. Figs. 6.4 and 6.5) and enhanced depositional growth. The effect of the convective pulsing on the snow category can be clearly seen in Fig. 6.6, as three distinct pulses are evident at each analysis time (from 120 through 240 minutes). When examining the snow category during this time period in more detail, it is apparent that each subsequent pulse experiences slightly greater mass growth, which is most likely the result of enhanced depositional growth as the mesoscale updraft intensifies with time (cf. Fig. 6.5). This is even more apparent in the hourly averaged data presented in Fig. 6.7, which shows a distinct increase in maximum snow mixing ratio from 120 through 240 minutes.

As snow advects rearward, it also increases in mass through the collection of cloud water. In addition, a small amount of ice is converted to snow immediately behind the convective line due to both collection and autoconversion processes. Given the smaller collection efficiencies used in this study (see discussion in Chapter 5) and also because autoconversion has been modified to be dependent on predicted ice size rather than a set mixing ratio, these processes all occur at a much slower rate than in Gallus and Johnson (1995). As a result, peak snow mixing ratios during the first four hours of model simulation are typically 1.2 g kg^{-1} , which is approximately 0.2 g kg^{-1} greater than the maximum value advected rearward from the convective line. At later times (not shown), snow mixing ratios in the model increase slightly as the mean updraft velocity increases, thereby resulting in increased deposition rates.

The instantaneous and hourly averaged cloud water fields are depicted by Figs. 6.8, and 6.9, respectively. Since cloud liquid water content (LWC; rather than cloud water mixing ratio) is the variable commonly addressed by laboratory electrification studies (e.g., Takahashi, 1978; Saunders et al., 1991), it is also what we present here. As noted earlier, cloud water is collected by snow at a much slower rate in this study than in Gallus and Johnson (1995) which, based on our aircraft measurements, greatly underpredicted cloud water in the stratiform precipitation region. Even with the lower collection efficiencies in our simulations, however, cloud water advected from the convective line (with peak LWCs of $> 0.5 \text{ g m}^{-3}$) is quickly depleted in the stratiform region. The hourly averaged data (Fig. 6.9) show cloud LWCs at locations 15 to 25 km behind the convective line (and at a temperature level of approximately -10°C) that are typically between 0.15 and 0.25 g m^{-3} , which is comparable to the average 0.23 g m^{-3} measured at similar temperature levels in the symmetric MCS by transition zone P-3 aircraft penetrations. Since we do not have any aircraft measurements further rearward in the stratiform cloud,

however, we do not know whether the small LWCs predicted by the model for those regions are realistic.

Finally, we present instantaneous model output of predicted ice concentrations and mass, which are depicted by Figs. 6.10 and 6.11, respectively (hourly averaged profiles vary little from the instantaneous model output for these two fields and are therefore not presented here). In the upper levels, cloud ice quickly advects rearward across the model grid with concentrations between 500 and 5000 l^{-1} . However, given their small sizes ($< 150 \mu\text{m}$) and fallspeeds ($< 7.5 \text{ cm s}^{-1}$), these upper-level ice crystals typically do not fall into the mixed-phase region ($T > -40^\circ\text{C}$) and are therefore not important to the non-inductive charging process discussed later. At lower levels, a plume of ice crystals with an initial concentration of 300 l^{-1} (based on P-3 aircraft observations in the transition zone region) advects rearward and is slowly depleted by snow collection. This eventually results in mixed-phase region ($0^\circ\text{C} > T > -40^\circ\text{C}$) ice concentrations of between 100 and 200 l^{-1} at locations within 40 km of the convective line. Furthermore, when comparing these concentrations to predicted ice mass (Fig. 6.11), it is evident that the average ice mass increases dramatically as the ice is advected rearward into the mesoscale updraft, eventually reaching a maximum ice mass of approximately $6.0 \times 10^{-10} \text{ kg}$ (corresponding to the ice to snow autoconversion threshold size of $400 \mu\text{m}$) at a location 45 km behind the convective line.

We also compare the predicted ice concentrations and mass at locations immediately above the melting level. Though the fragmentation parameterization (discussed in Chapter 5) works to maintain ice concentrations of between 100 and 200 l^{-1} just above the melting level (in agreement with the measurement of Willis and Heymsfield, 1989), the corresponding predicted ice mass in this region is typically quite small. The reason for this is apparent when Figs. 6.10 and 6.11 are compared to the model predicted vertical velocity structure (Figs. 6.4 and 6.5), which indicates

that the region immediately above the melting level is situated in the mesoscale downdraft. Even though model ice concentrations are continually replenished by the fragmentation function in this region, the ice particles are also in a constant state of sublimation that reduces their size. Since we do not have high quality microphysical observations at these temperature levels, we do not know whether these predicted ice sizes agree well with observations in the 24 May 1991 MCS (though deep subsidence in this region was indicated by the our observations). Nevertheless, this model-generated microphysical feature is important towards understanding the non-inductive charging model simulations, one of which is dependent on ice crystal size (Saunders et al., 1991) and one which is not (Takahashi, 1978).

Since high quality, multi-level microphysical measurements in MCS stratiform clouds are quite rare, we do not know whether the model predicted microphysical structure presented here is characteristic of all symmetric MCSs. Many uncertainties regarding MCS microphysical structure remain. Based on data from our observational analysis (Chapter 3) and insight gained from previous observational and theoretical studies (e.g., Willis and Heymsfield, 1989; Yeh et al., 1991; Biggerstaff and Houze, 1993; Braun and Houze, 1994), however, we believe the model microphysical structure presented here is consistent with all available in situ measurements, particularly at locations near the leading edge of the radar bright band where most of our electric field measurements were obtained. Given this, we now investigate the electrification processes in detail.

6.2 Charge advection

The first electrification process that we investigate is charge advection. For this simulation, we use the convective line charge initialization profile presented in Fig. 5.7, which agrees well with the composite MCS convective line profile produced by Stolzenburg (1996). In this simulation, which uses the microphysical control run presented above, the charge profile from

Fig. 5.7 is allowed to advect rearward without any contributions by in situ charge generation processes. Thus, the only modification that occurs to the advected charge structure is due to sedimentation, turbulent diffusion, and charge exchange between hydrometeor categories by collection, autoconversion, fragmentation, sublimation, evaporation, and melting (discussed in Chapter 5). The results from this simulation are depicted in Fig. 6.12.

In general, the advection profile maintains its dipole structure, with positive charge situated above negative, to locations approximately 60 km behind the convective line. The weak lower positive charge center, which is primarily associated with positively charged ice in the initialization profile (see Fig. 5.7), is quickly overwhelmed by the faster falling, negatively charged snow. Since snow in the model falls at the bulk (i.e., density weighted) fallspeed, a significant lowering of the negative charge center occurs immediately behind the convective line. Fig. 6.13 depicts snow/aggregate fallspeeds for typical mixing ratios (computed using fallspeed coefficients from Appendix B and an approximate air density corresponding to a temperature of -15°C). Calculations using these values suggest that using a fallspeed of 1.2 m s^{-1} and horizontal wind of 15 m s^{-1} (cf. Fig. 3.9), charge on the snow category is lowered by approximately 2.5 km as it crosses a 30 km transition zone region.

The effect of turbulent diffusion on advected charge structure was also investigated (see Eqns 5.15 and 5.16). Turbulent diffusion of charge has previously been investigated by Rutledge et al. (1993) and Stolzenburg et al. (1994). Though the turbulence parameterization used in this model differs slightly from that used by Stolzenburg et al. (1994), our results are quite similar. Using typical eddy exchange coefficients, Stolzenburg et al. (1994) concluded that turbulent diffusion resulted in a 10 nC m^{-3} charge core being reduced to 6.7 to 9.0 nC m^{-3} at distances 60 km from the convective line. Since our model is more complicated than theirs in that it includes interactions between multiple hydrometeor categories, we do not similarly quantify the

effect of turbulent diffusion. From sensitivities model runs wherein turbulent diffusion was not included, however, we find that the addition of turbulent diffusion results in only minor changes to the advected charge distribution.

6.3 Non-inductive charging

We now investigate contributions from non-inductive charging processes, both with and without charge advection. In particular, studies of Takahashi (1978) and Saunders et al. (1991) have shown promise in explaining at least some of the charge features observed in convective clouds. Yet, they also exhibit marked disagreement for some temperature/LWC regimes (e.g., Williams, 1989). Therefore, we investigate both laboratory studies in the context of the model simulations.

6.3.1 Takahashi (1978)

6.3.1.1 Without charge advection

The evolution of instantaneous and hourly averaged charge densities from a simulation that uses the laboratory data of Takahashi (1978) are depicted by Figs. 6.14 and 6.15, respectively. Within the first hour, the model produces an inverted dipole charge structure consisting of negative charge overlying positive charge. This charge structure is similar to the results of Rutledge et al. (1990), whose one-dimensional model produced an inverted dipole and charge generation rates of up to $3.6 \text{ nC m}^{-3} \text{ h}^{-1}$ (using the laboratory data of Saunders and Jayaratne, 1986). From this, they concluded that in situ charge generation might be responsible for the production of positive CG lightning in MCS stratiform precipitation regions. In our simulations, maximum charge density magnitudes associated with the inverted dipole, during the first hour of integration, are typically $< 2 \text{ nC m}^{-3}$ (with hourly averaged magnitudes $< 1 \text{ nC m}^{-3}$). By 180 minutes, however, maximum charge densities in both the instantaneous and hourly averaged

model output exceed 5 nC m^{-3} , which compares well with charge densities derived from the balloon-borne electric field meter observations (see Chapter 3).

We also examine the electric field structure of this charge distribution since that is the field most comparable to the EFM observations. Electric fields (as determined by solving Poisson's equation) are as high as 300 kV m^{-1} , which is significantly larger than observed in MCSs and also larger than is required for dielectric breakdown (e.g., Marshall et al., 1995c). Nevertheless, even with the absence of a discharge parameterization (e.g., Helsdon et al., 1992; Solomon and Baker, 1996), one of the more interesting results of the non-inductive charging simulations is that, acting alone, the non-inductive mechanism tends to produce a quasi-steady charge structure after approximately two hours of integration. That is, once observed charge densities are achieved, it appears that the sink of charge resulting from fallspeed divergence becomes approximately equal to the rate of charge generation due to non-inductive processes. This might lead to a quasi-steady layered structure similar to that commonly indicated by the observations.

When detailed comparisons are made between model output and observations, the most realistic feature of the simulated charge distribution is a sharp charge transition (negative above positive) near the melting level (cf. Figs. 3.12, 3.13, 3.26, and 3.27). Several hypothesis have been advanced to explain this charge transition (e.g., Shepherd et al., 1996). In our model simulations, insight into its origin can be gained by individually examining charge in the snow and ice categories (Figs. 6.16 and 6.17). An analysis of these output reveal that although the total charge density above the melting level is negative, the charge polarity of the snow category (resulting from collisions with the slower falling ice crystals) is positive. Thus, the negative total charge density is primarily the result of the large amount of negative charge remaining on the ice crystals. Near the 0°C isotherm, the cloud ice is quickly removed by sublimation (or melting and subsequent evaporation), thereby exposing the positive charge on the larger hydrometeors and

creating a sharp charge transition near the melting level. Unfortunately, due to computational constraints addressed in Chapter 5, our model did not include continuity equations for ions. Therefore, it is not possible for us to address the disposition of the ions that are released to the atmosphere during this sublimation/evaporation process. It is interesting to note, however, that the in situ particle charge measurements by Bateman et al. (1995) indicated that the smaller particles were primarily responsible for the measured electric field, which is also the case in this simulation. The sensitivity of this process to collection efficiency will be tested in Chapter 7.

The depth of this low-level positive charge layer is also similar to observations. Though positive charge is found at all locations below the melting level at locations within 60 km of the convective line, the depth of the high charge density region (i.e., $> 1 \text{ nC m}^{-3}$) is rather small (Figs. 6.14, 6.15), in agreement with observations, which indicate that the depth of this lower positive charge region is typically about 500 m. In the model results, the reduction in positive charge density below this level is primarily due to fallspeed divergence and evaporation, both of which occur as the positively charged aggregates melt. While the model output exhibit marked agreement with the observations near the melting level, they show much less agreement aloft. It might be recalled that the charge structure derived from the balloon-borne electric field profiles exhibited a distinct (negative above positive) charge transition at approximately -12°C . No such charge transition is present in this simulation, for the Takahashi non-inductive parameterization with no charge advection. The possible origin of this charge region will be discussed in more detail after simulations involving charge advection and simulations using the Saunders et al. (1991) non-inductive charging parameterization are discussed.

6.3.1.2 With charge advection

The simulation using the non-inductive charging data of Takahashi (1978) with charge advection activated is now presented. The evolution of instantaneous and hourly averaged

charge densities from this simulation are depicted by Figs. 6.18 and 6.19, respectively. The primary effect of adding charge advection from the convective line is a slight reduction in the intensity of the positive charge near the melting level, as well as the establishment of positive charge aloft associated with the advection of positive charge from the convective region. Hence a three layer model is predicted. As noted earlier, collisions between the snow and cloud ice categories in the stratiform precipitation region results in a positive charge transfer to the snow and negative to the ice. Since the advected snow has an initial negative charge, however, it does not attain a net positive charge in this simulation until it reaches a location approximately 15 km behind the convective line. Nevertheless, even with the advection of negatively charged snow, a broad region of positive charge with densities $> 1 \text{ nC m}^{-3}$ still develops near the melting level by 180 minutes into the simulation. At upper levels, advection merely results in a deep layer of positive charge (with charge densities from $1\text{-}2 \text{ nC m}^{-3}$) that gradually mixes down into the negative charge below, eventually lowering the upper-level charge transition to approximately -20°C .

When comparing the model results with observed charge structures (see Chapter 3), the most striking similarity, as noted earlier, is near the melting level. However, at upper levels, this simulation compares rather poorly with observations. Most notably, since aircraft observations indicated water saturated conditions to temperatures of only -25°C , it appears likely that most charge found above the -25°C level is the result of the advection of charge on small, slowly falling ice particles since significant non-inductive charging would not be possible in the absence of supercooled water. The EFM observations indicate that charge in this region is of the negative polarity, which disagrees not only with observations, but with the composite MCS convective charge profile (e.g., Stolzenburg, 1996) that is used to initialize the convective line charge structure in the model.

The total charge density from this model run is summarized by Fig. 6.20, which depicts grid averaged charge densities at 180 minutes into the simulation, and Fig. 6.21, which depicts charge densities derived from the second balloon-borne EFM flight into the symmetric MCS (Fig. 6.21a) and a vertical cross-section of charge density constructed through model grid at a location 35 km behind the convective line (Fig. 6.21b, a location analogous to where the balloon-borne EFM ascended). As noted earlier, the advection of low-level negative charge on snow resulted in a reduction in the intensity of the lower positive region in this simulation. Nevertheless, the grid averaged data show a distinct charge transition (negative above positive) located at the 0°C isotherm, with grid averaged charge density magnitudes above this level of approximately 4 nC m⁻³. It is interesting to note that an analogous charge transition in the observational data commonly occurs at a temperature of approximately -1°C. In agreement with the observations of Willis and Heymsfield (1989), ice concentrations in our model (resulting from the fragmentation parameterization) typically peak at -2°C and decrease below that level. Given this rapid decrease in cloud ice concentrations near the melting level, it is not unreasonable to expect that this sharp transition may, in some cases, occur at temperatures slightly colder than 0°C.

6.3.2 *Saunders et al. (1991)*

6.3.2.1 Without charge advection

Using the same microphysical control run as above, we now investigate non-inductive charging using the laboratory data from Saunders et al. (1991). The evolution of instantaneous and hourly averaged charge densities from this simulation are shown in Figs. 6.22 and 6.23, respectively. When compared in detail to the charge structures generated with the Takahashi data (cf. Figs. 6.14-6.20), several differences are found. Foremost, this simulation does not develop a sharp charge transition near the melting level. As noted earlier, this is most likely due

to the Saunders et al. parameterization's dependence on ice size (see Eqn. 5.23). Since the predicted ice size at locations above the melting level is typically $<150\ \mu\text{m}$ (Fig. 6.11), the amount of charge transfer per collision is also quite small. Thus, in contrast to the simulation that uses the Takahashi data, a large amount of charge does not build up on the cloud ice category at locations immediately above the melting level and, subsequently, a sharp charge transition at the melting level does not occur.

While this simulation does not reproduce observed charge structures near the melting level very well, it does appear to exhibit much better agreement with observations at higher levels. Most notably, a positive layer of charge develops at locations immediately behind the convective line and at a temperature level quite similar to that indicated by the observations. Charge densities associated with this positive charge layer are as high as $2\ \text{nC m}^{-3}$. When charge on the snow and ice categories are analyzed individually (Figs. 6.24 and 6.25), we find that the bulk of this positive charge resides on the ice and the negative charge below it is predominantly on the snow. Initially, the top of the positive layer is at a temperature of approximately -20°C , but it gradually descends to about -15°C at a location 30 km behind the convective line. This is quite similar to a charge transition (negative above positive) at -12°C in the symmetric MCS transition zone electric field profiles (cf. Figs. 6.24 and 6.25). Overall, maximum charge densities in this simulation are somewhat smaller ($< 3\ \text{nC m}^{-3}$) than in the Takahashi simulation and electric fields remain below $150\ \text{kV m}^{-1}$. Both of these values, however, agree well with observations.

6.3.2.2 With charge advection

The simulation using the data of Saunders et al. (1991) is now presented with the addition of charge advection. The evolution of instantaneous and hourly averaged charge densities from this simulation are depicted by Figs. 6.26 and 6.27, respectively. The primary effect of the adding

charge advection is to delay the onset of the mid-level positive charge layer (discussed above) to a location approximately 30 km behind the convective line, thereby resulting in a transition zone charge structure that is quite different from that found in the stratiform precipitation region. Furthermore, the advection of negative charge on the snow category combines with the non-inductive charging to give a much sharper charge transition (positive above negative) at approximately -10°C , compared to the no-advection simulation. It should be noted that the symmetric MCS transition zone electric field meter profiles, which were obtained at a location approximately 30 km behind the convective line, showed a quite similar charge structure in the -5 to -20°C layer. Given the much weaker charge densities associated with these layers, however, a vertical profile of the electric field through this region (not shown) is quite different than that indicated by the observations.

As with the Takahashi simulation, the largest disagreement with the observations is in the cloud's upper levels, where observations indicate the presence of a deep layer of negative charge. The total charge density from this model run is summarized by Fig. 6.28, which depicts grid averaged charge densities at 180 minutes into the simulation, and Fig. 6.29, which depicts charge densities derived from the second balloon-borne EFM flight into the symmetric MCS (Fig. 6.29a) and a vertical cross-section of charge density constructed through model grid at a location 35 km behind the convective line (Fig. 6.29b, a location analogous to where the balloon-borne EFM ascended). When Fig. 6.28 is compared with the balloon-borne electric field meter observations, a similar overall structure is apparent, though the intensity and location of some of the charge layers are quite different than indicated by the observations. In the average, a charge transition of positive above negative occurs at a temperature of approximately -1°C . In comparing this average to the model output (Figs. 6.26 and 6.27), however, it can be seen that this charge transition does not typically occur until a location approximately 40 km behind the convective

line. In advance of this location, the predominant charge polarity above the melting level is negative. At higher levels, a thin low charge density (0.5 nC m^{-3}) region of negative charge overlies the lower positive charge region, creating a quasi-layered structure that is similar to that indicated by the observations. Given the positive charge aloft, this mean charge structure is probably more similar to the profile presented by Schuur et al. (1991) for another symmetric MCS (cf. Fig. 4.7) than to the two electric field profiles shown in Chapter 3. That is, a sharp transition does not occur at 0°C but at a location of approximately -5°C . Furthermore, the predominant charge polarity in the upper levels is positive, with a negative screening layer.

6.3.3 Charge budget analysis

As stated earlier, one of the primary goals of the modeling component of this study is to determine the relative contributions of 1) charge advection, and 2) local charge generation resulting from non-inductive charging in the stratiform region. In this section, we present a charge budget analysis that investigates the contributions from each of these mechanisms. In this analysis, charge advection rates are computed by determining fluxes through the stratiform region's right boundary (cf. Fig. 5.1). These fluxes are then compared with volume-integrated in situ charge generation rates determined using the Takahashi (1978) or Saunders et al. (1991) laboratory results. The charging rates are then summed over the two convective pulses that constitute the third hour of the model integration (cf. Fig. 5.9) to determine a mean charging rate over a one hour period.

Since the Saunders et al. parameterization is dependent on ice crystal size, we have determined charging rates for four different advected ice sizes. That is, the concentration of ice in the mixed-phase region is held constant at 300 l^{-1} , in agreement with observations, and the average size (mass) of the advected ice crystal is modified. Therefore, with each model run, a different amount of total ice mass (i.e., ice mixing ratio) is advected rearward in the lower

portion of the profile (cf. Fig. 5.3, which depicts the ice mixing ratio associated with the control run ice size of 258 μm). Due to dynamic feedbacks in the model associated with the advection of these different low-level ice mixing ratios, the Takahashi parameterization (which is not dependent on the size of the ice crystal) is also tested with these different ice sizes.

Non-inductive charging statistics for the Takahashi (cf. Fig. 1.6) and Saunders et al. (cf. Fig. 1.8) data sets are presented in Tables 6.1 and 6.2, respectively. For the four ice sizes tested in these simulations, the maximum charging rate produced with Takahashi input data (1.84 C h^{-1}) is approximately six times larger than that produced with the Saunders et al. data (0.32 C h^{-1}). It is also interesting to note that the Takahashi data produces its largest charging rates when small ($52 \mu\text{m}$) ice crystals are advected from the convective line whereas the Saunders et al. data produces its largest charging rates when large ($326 \mu\text{m}$) ice crystals are advected rearward. For the Takahashi simulation, there is no dependence on ice crystal size, and as a result, there is less water vapor removed by depositional growth, resulting in slightly higher LWCs and a larger charge transfer per collision (at small ice sizes). Another difference can be seen by comparing the maximum charge densities. Though the amount of charge transfer associated with the Takahashi parameterization is not dependent on the ice size, dynamical feedback in the model resulted in maximum charge densities produced in the Takahashi simulations that varied much more (with ice size) than did maximum charge densities in the Saunders et al. simulations. For all ice sizes, maximum charge densities produced with the Saunders et al. data ($\sim 3 \text{ nC m}^{-3}$) were all quite similar to that indicated by observations. It should be noted, however, that even though the high charge density magnitudes were produced with the Takahashi data ($> 20 \text{ nC m}^{-3}$), these high charge densities were typically short-lived.

A comparison of the non-inductive and advective charging rates for the Takahashi and Saunders et al. data sets are shown in Tables 6.3 and 6.4, respectively. Advective charging rates

changed very little over the range of ice sizes tested, with a charge advection rate of 0.58 C h^{-1} associated with the advection small ice crystals and a charge advection rate of 0.49 C h^{-1} associated with the advection of large ice crystals. It should be noted that these changes are entirely due to dynamical feedbacks in the model associated with the advection of different low-level ice mixing ratios as the amount of charge assigned to the ice category in the initialization profile for each of these simulations was not modified. When comparing the charge advection rates to the rate of charge generation due to the non-inductive process, it can be seen that the change in ice size had very little overall affect on the charge budget in the Takahashi simulations. That is, though the overall non-inductive charging rate decreased with increasing ice size, dynamical feedbacks also resulted in a simultaneous decrease in charge advection. As a result, the in situ charge generation with the Takahashi data consistently constituted between 70 and 76% of the total charge introduced into the stratiform precipitation region. In contrast, the Saunders et al. simulations exhibited much more sensitivity to the advected ice size, and as a result, non-inductive charging accounted for between 15-40% of the total charge introduced into the stratiform precipitation region. This percentage appears to be fairly constant between 30-40% for the larger ice sizes. It is interesting to note that the contribution of in situ charge generation to the total charge in the stratiform region using the Takahashi charging results (70-76%) is consistent with the contribution that mesoscale ascent (in situ condensate production) makes to the total stratiform precipitation (80%, Rutledge and Houze, 1987). This consistency suggests that stratiform electrification is strongly tied to stratiform precipitation processes.

6.4 Evaporation-condensation charging

Stolzenburg et al. (1994) and Rutledge and Petersen (1994) both suggested that the evaporation-condensation mechanism (proposed by Dong and Hallett, 1992) might be responsible for at least some of the charge observed in MCS stratiform precipitation regions. In

particular, Rutledge and Petersen (1994) provided order-of-magnitude calculations that suggested charge densities produced by this mechanism were two orders of magnitude smaller than those observed in MCS stratiform regions. Here, we formally test this mechanism using the parameterization presented in Chapter 5. In this simulation, in accord with the Dong and Hallett (1992) results, the snow and cloud ice categories acquire a positive (negative) charge when in a state of deposition (sublimation) at temperatures between -4 and -20°C . At temperatures warmer than -4°C , negative (positive) charge is acquired during deposition (sublimation). The results from this simulation are presented in Fig. 6.30.

The most notable result of this simulation, in agreement with the calculations of Rutledge and Petersen (1994), is the very small charging rates it produces. In general, the maximum charge density produced with this mechanism was $< 0.05 \text{ nC m}^{-3}$, which is approximately two orders of magnitude smaller than indicated by the observations. It should be noted that Dong and Hallett (1992) found that charge transfer magnitudes were increased significantly when cloud ion concentrations were large and also when previous riming had taken place on the growing ice particle. When these uncertainties are taken into consideration (potentially increasing charging rates by a factor of 50), it is possible to generate charge densities that are similar to those indicated by the observations. However, cloud ion concentrations are not well known and riming in MCS stratiform regions is typically quite weak, so charging via this mechanism remains uncertain.

Another notable feature in this simulation is the relative lack of negative charge. Though negative charging does take place at locations immediately above the melting level, this negative charge is easily overwhelmed by the fallout of positive charge generated in the deep ascent layer above. As a result, the only appreciable region of negative charge in the model output is towards the rear edge of the stratiform precipitation region where dry air intrusion associated with the

developing rear inflow jet (cf. Figs. 6.2 and 6.3) results in an extensive region of sublimation, thus causing ice particles to charge negatively.

6.5 Melting charging

Finally, we investigate charging from the melting mechanism of Drake (1968). As noted in previous chapters, the melting mechanism has recently been addressed by several studies (e.g., Marshall and Rust, 1993; Stolzenburg et al., 1994; Shepherd et al., 1996) as a potential source of charge near the melting level of MCS stratiform precipitation regions. We do not investigate the inductive melting mechanism of Simpson (1909) since that requires an examination of fallspeed divergence with the rain category (at a given level), which is not possible with this bulk microphysical model.

The results from the melting charging simulation are presented in Fig. 6.31. As discussed in Chapter 5, given that Drake (1968) studied the melting of ice spheres (i.e., graupel) rather than aggregates, the applicability of this mechanism to the MCS stratiform precipitation region is questionable. Nevertheless, applying Drake's results to melting aggregates produces a deep region of positive charge with charge densities that are typically $< 0.4 \text{ nC m}^{-3}$. Using the charging rate from Drake (1968) and converting the model maximum snow mixing ratio of 1.2 g kg^{-1} (see Fig. 6.6) to melt water, one would expect that this mechanism could yield charge densities as high as 1.85 nC m^{-3} . However, the model simulation indicates that fallspeed divergence just below the melting level resulted in the maximum density being limited to $\leq 0.35 \text{ nC m}^{-3}$. Our electric field meter profiles indicate the presence of predominantly negative charge below cloud base though other studies have found positive charge at those levels (i.e., Hunter et al., 1992). It is possible that processes not included in our model, such as point discharge from the ground, are important towards modifying the charge structure in this region.

Table 6.1: Takahashi (1978) non-inductive charging statistics.

ice mass (10^{-10} kg)	ice size (μm)	max + charge (nC m $^{-3}$)	max - charge (nC m $^{-3}$)	charging rate (C h $^{-1}$)
0.1	52	6.21	-20.20	1.840
1.0	163	4.82	-10.41	1.441
2.5	258	3.92	-10.15	1.323
4.0	326	2.71	-8.11	1.183

Table 6.2: Saunders et al. (1991) non-inductive charging statistics.

ice mass (10^{-10} kg)	ice size (μm)	max + charge (nC m $^{-3}$)	max - charge (nC m $^{-3}$)	charging rate (C h $^{-1}$)
0.1	52	1.82	-1.91	0.094
1.0	163	2.47	-1.86	0.242
2.5	258	3.19	-2.17	0.325
4.0	326	2.74	-3.21	0.318

Table 6.3: Comparison of Takahashi (1978) non-inductive charging with charge advection from the convective line.

ice mass (10^{-10} kg)	ice size (μm)	NIC rate (C h $^{-1}$)	Advection rate (C h $^{-1}$)	Percent NIC (%)
0.1	52	1.840	0.575	76.2
1.0	163	1.441	0.549	72.4
2.5	258	1.323	0.524	71.6
4.0	326	1.183	0.491	70.7

Table 6.4: Comparison of Saunders et al. (1991) non-inductive charging with charge advection from the convective line.

ice mass (10^{-10} kg)	ice size (μm)	NIC rate (C h $^{-1}$)	Advection rate (C h $^{-1}$)	Percent NIC (%)
0.1	52	0.094	0.575	14.1
1.0	163	0.242	0.549	30.6
2.5	258	0.325	0.524	38.3
4.0	326	0.318	0.491	39.3

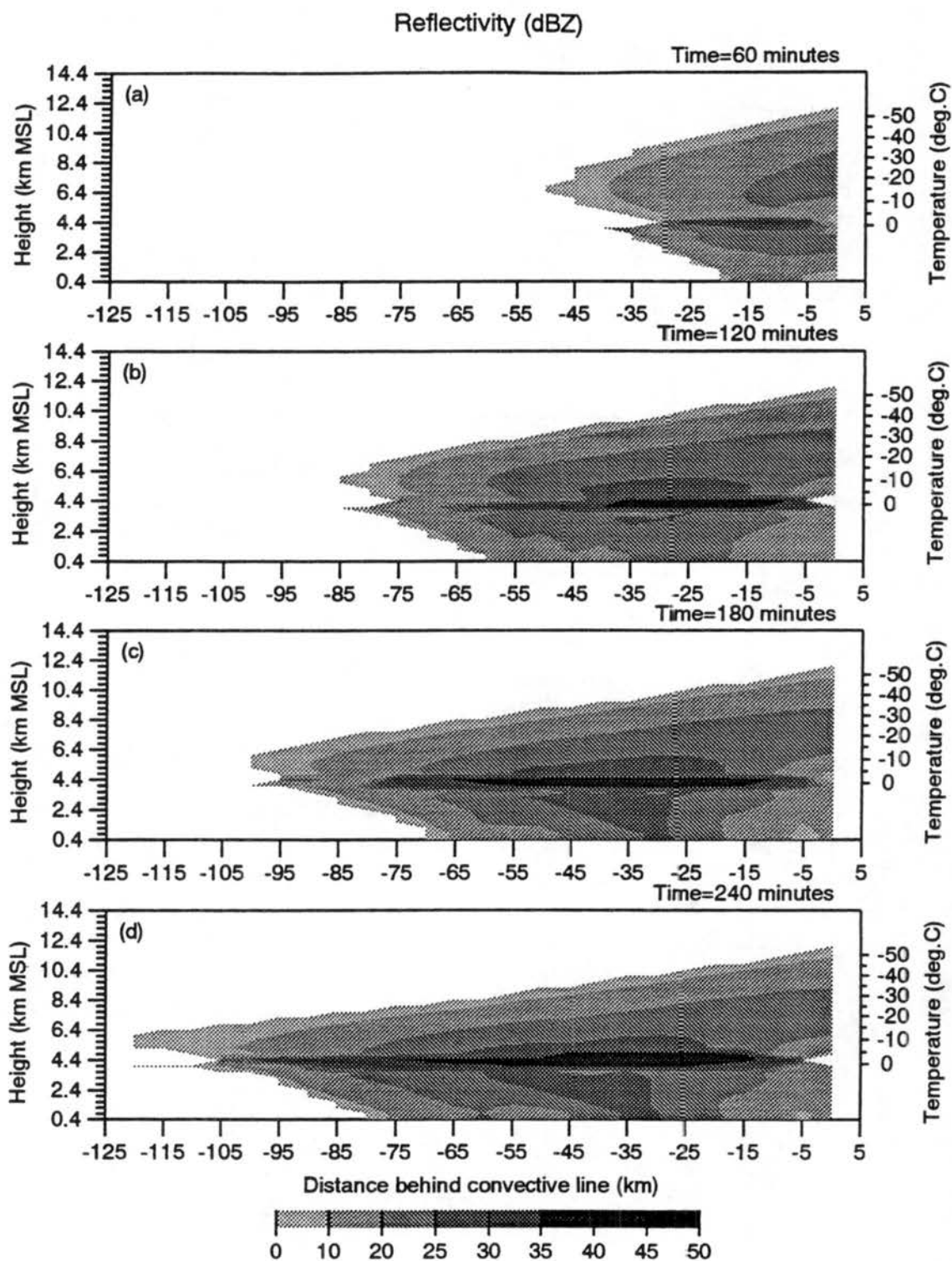


Fig. 6.1. Control run radar reflectivities (dBZ) computed from precipitation-sized hydrometeor fields at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Shading denotes reflectivity intensity (key at bottom). Model output is averaged over two convective pulses (cf. Fig. 5.9).

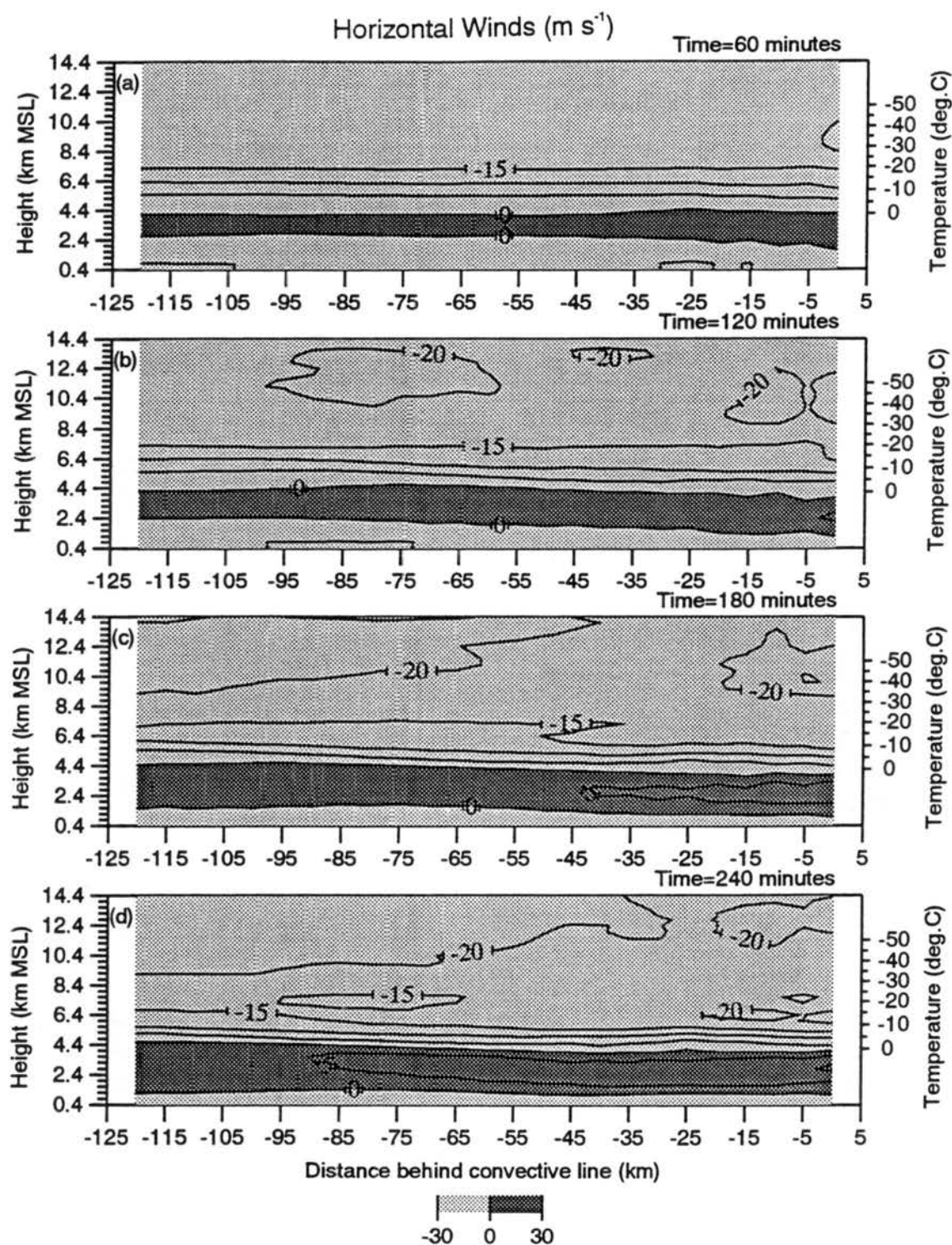


Fig. 6.2. Control run horizontal winds (m s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes front-to-rear flow while dark shading denotes rear-to-front flow. Contours are every 5 m s^{-1} .

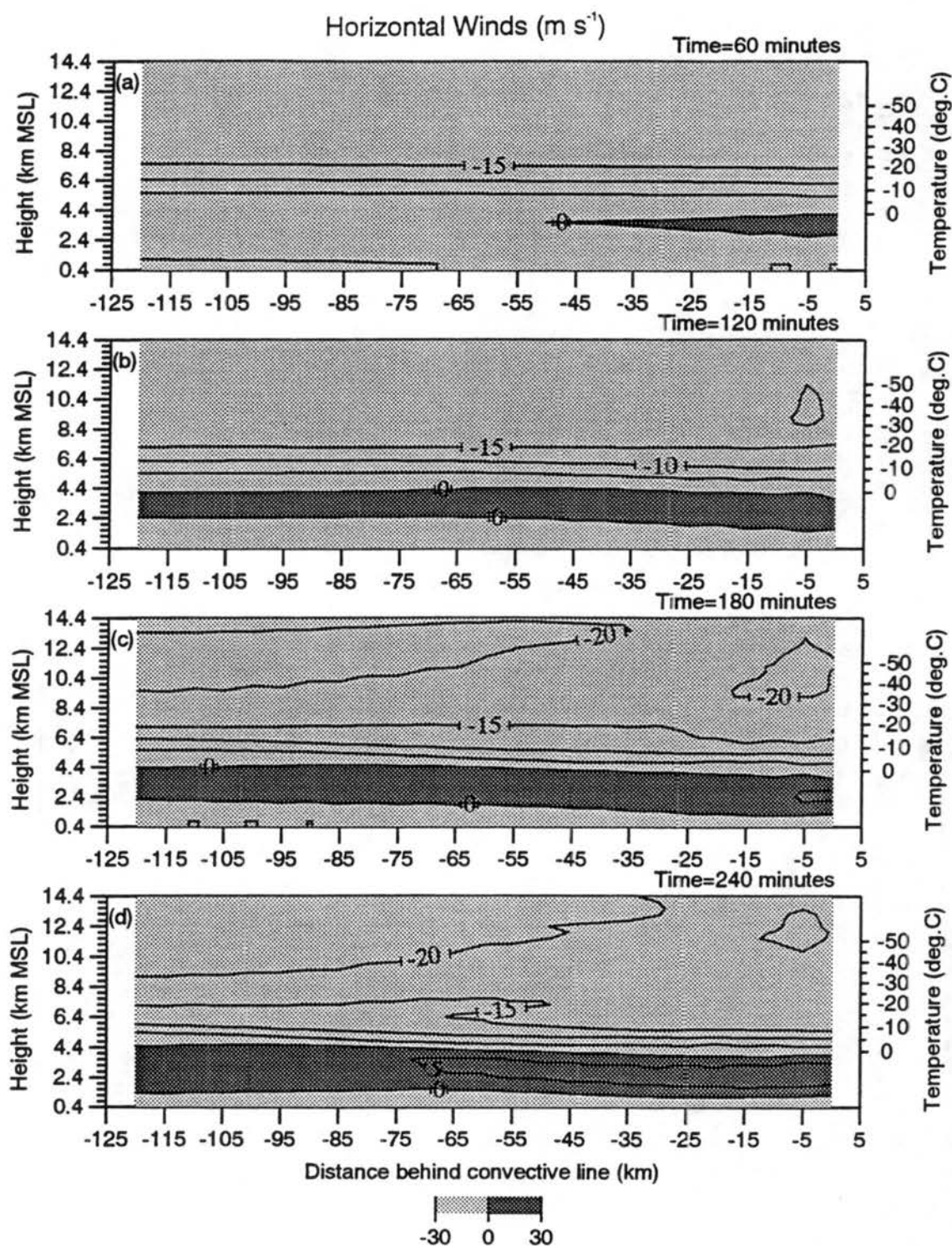


Fig. 6.3. Same as Fig. 6.2 except model output is averaged over two convective pulses.

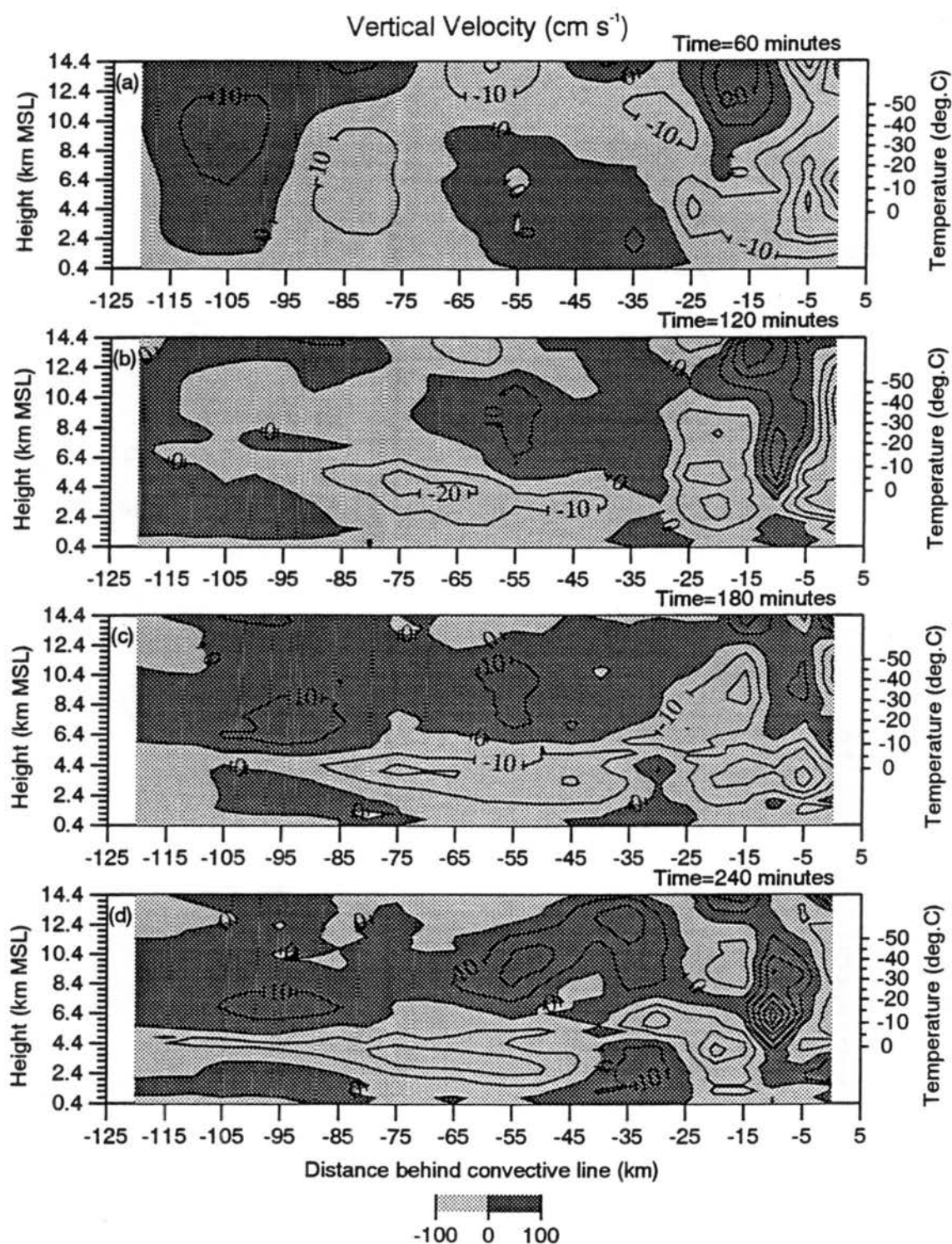


Fig. 6.4. Control run vertical velocities (cm s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes downward motion while dark shading denotes upward motion. Contours are every 10 cm s^{-1} .

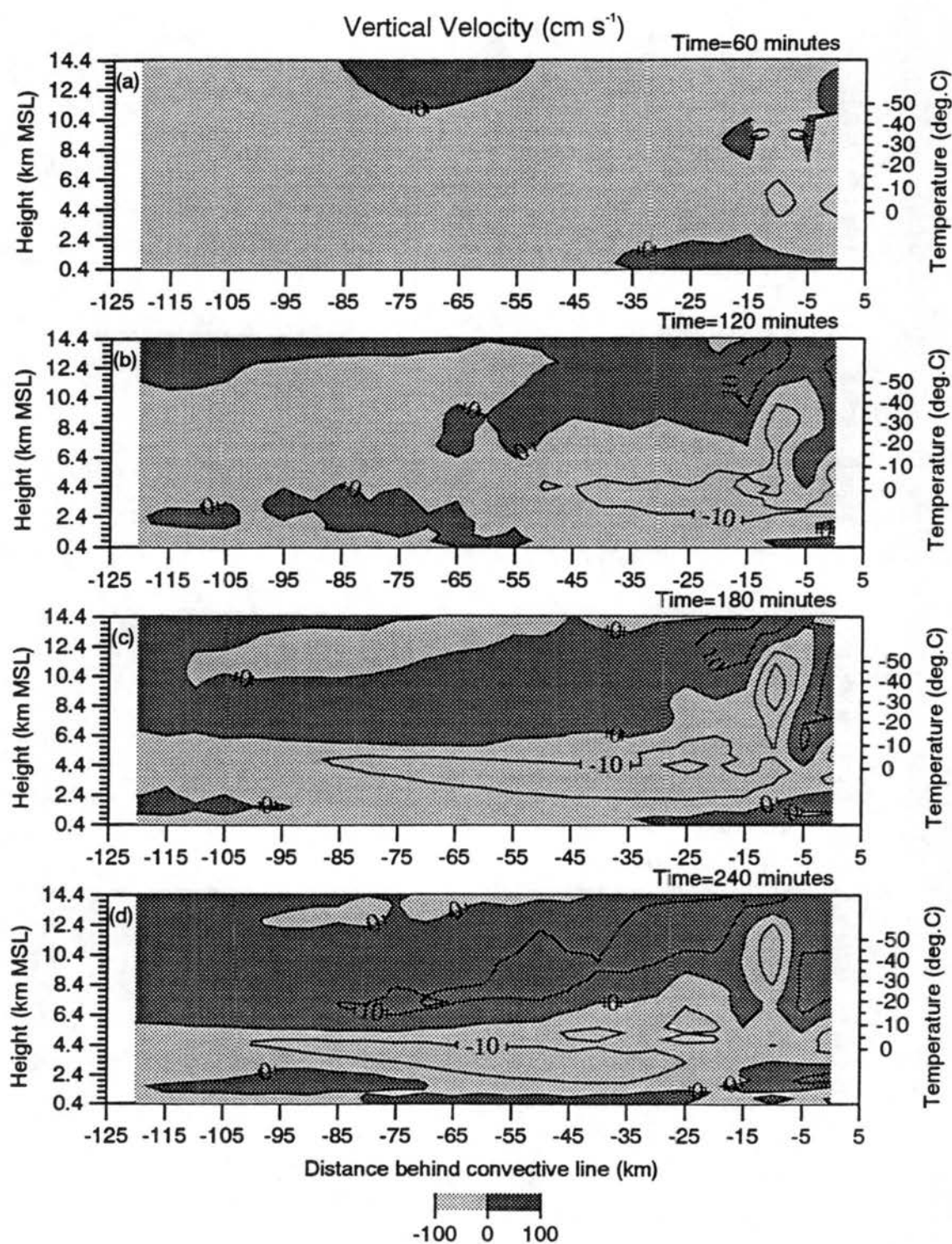


Fig. 6.5. Same as Fig. 6.4 except model output is averaged over two convective pulses.

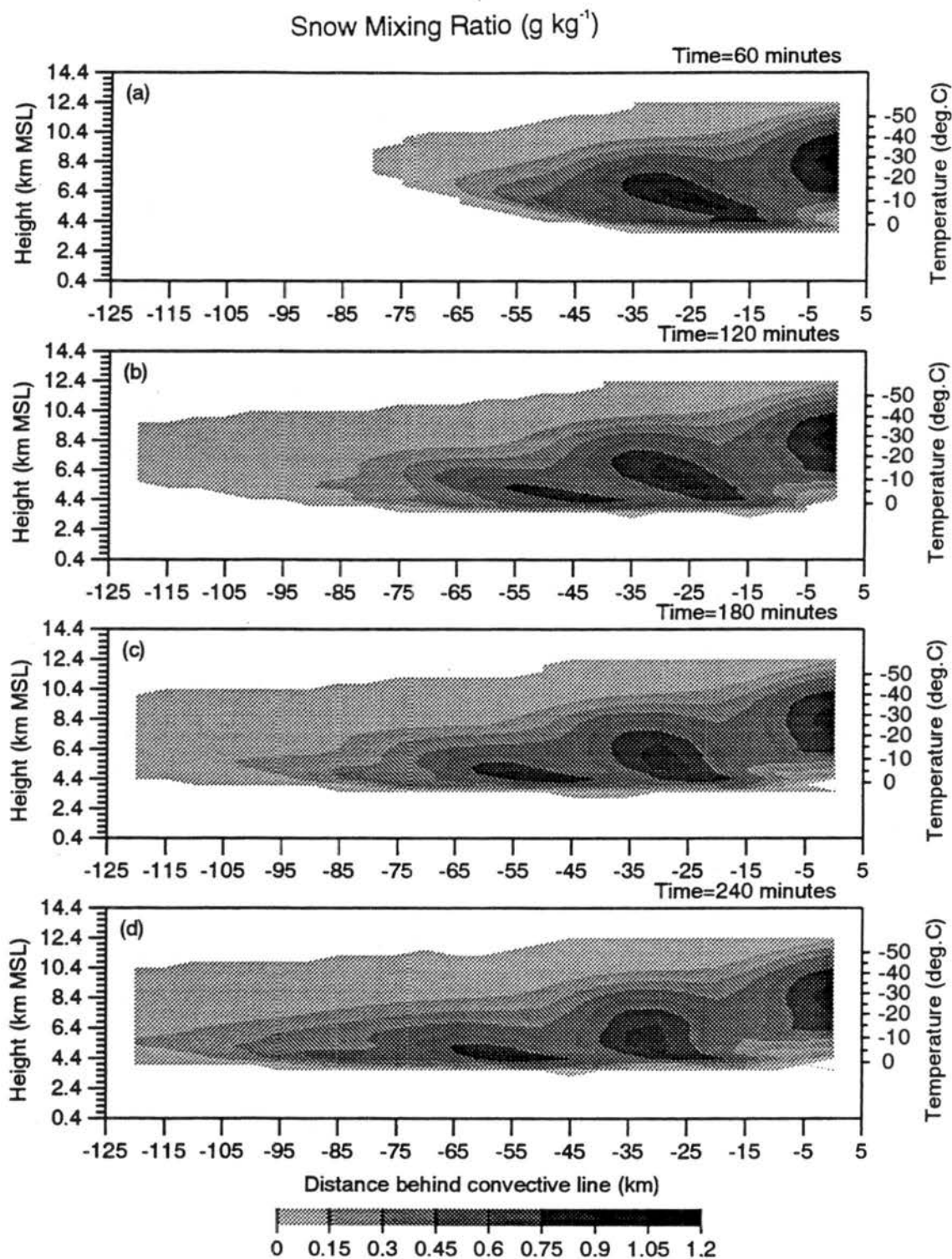


Fig. 6.6. Control run snow mixing ratios (g kg^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Shading denotes mixing ratio at 0.15 g kg^{-1} intervals (key at bottom).

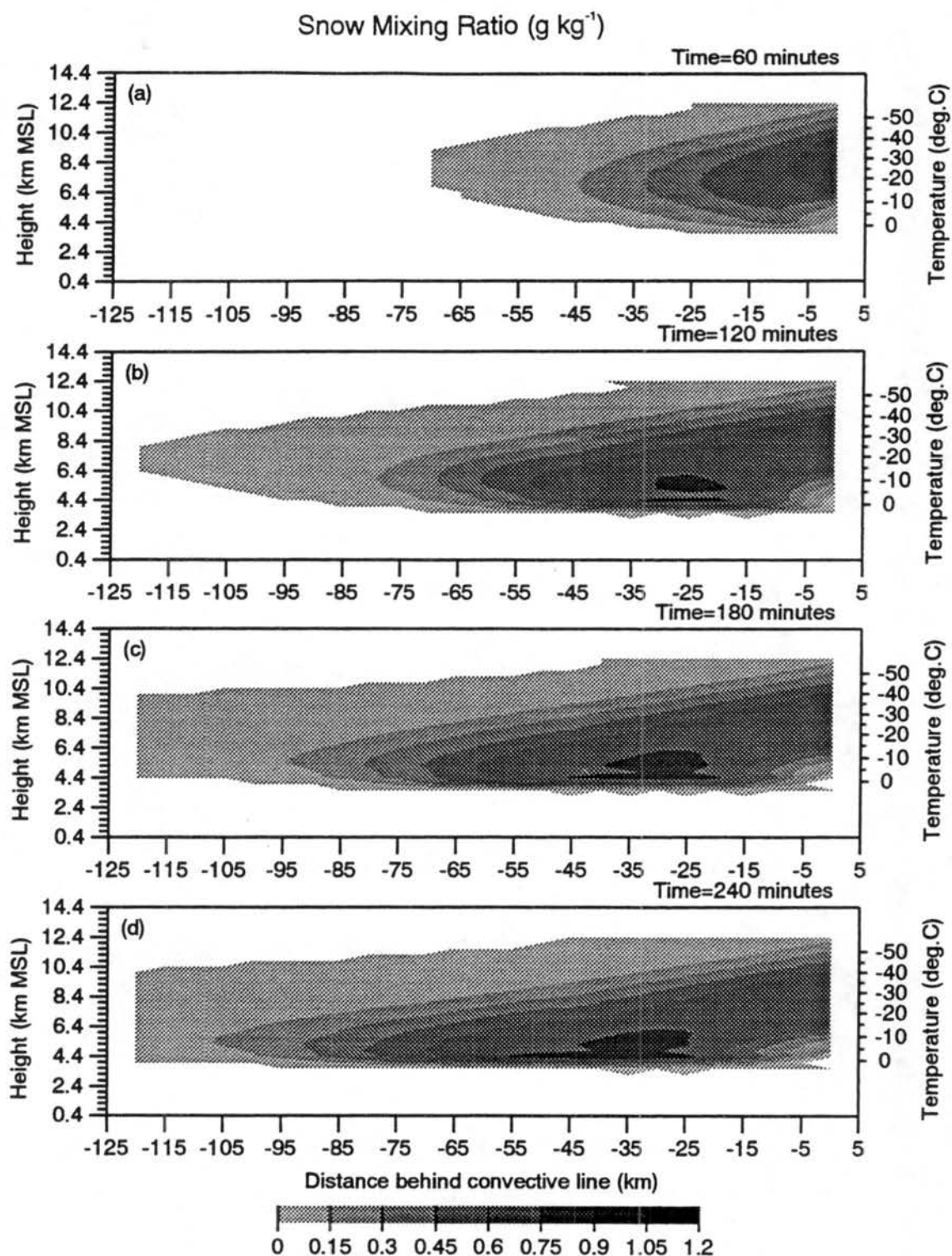


Fig. 6.7. Same as Fig. 6.6 except model output is averaged over two convective pulses.

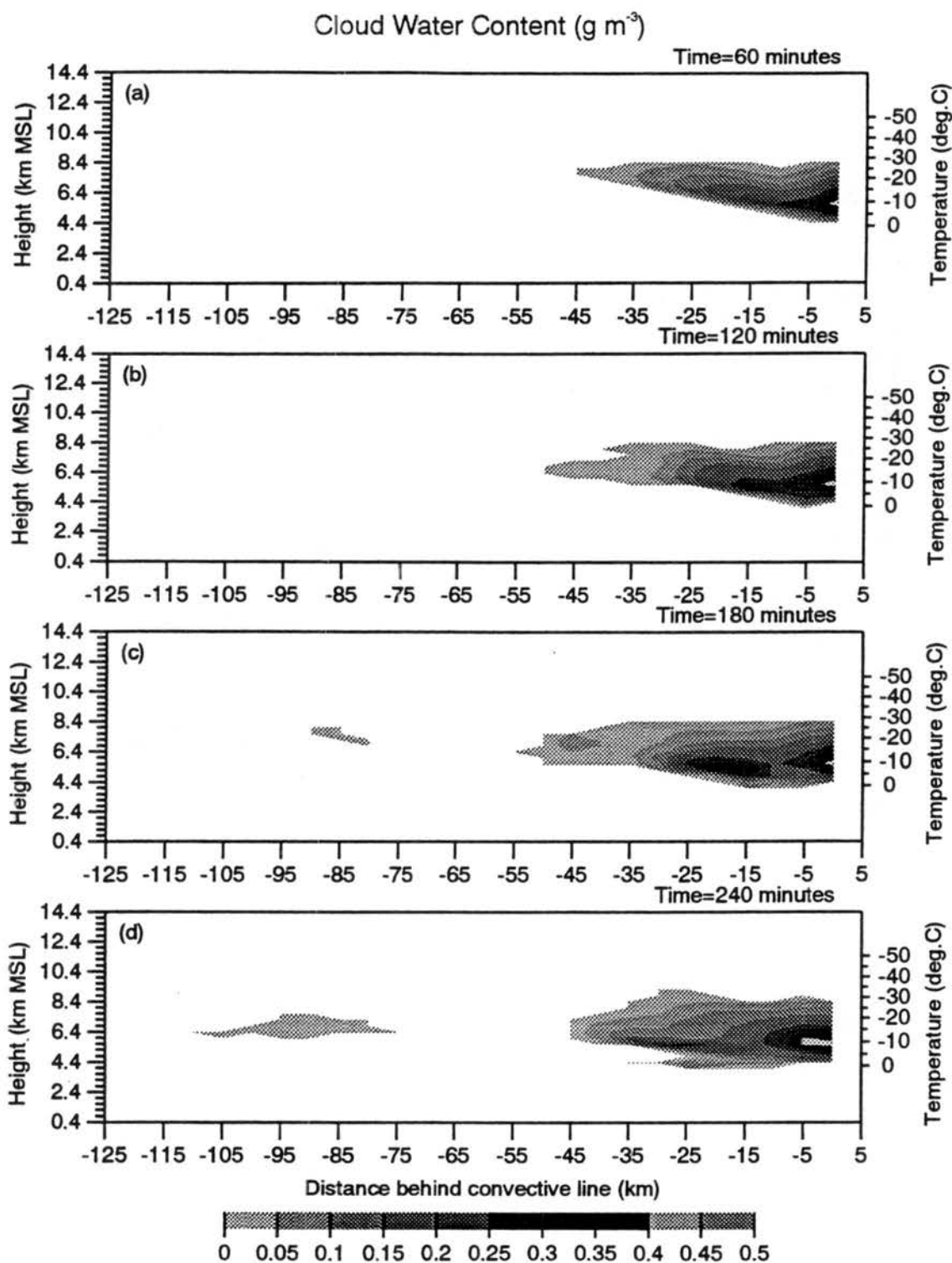


Fig. 6.8. Control run cloud water liquid water content (g m^{-3}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Shading denotes mixing ratio at 0.05 g kg^{-1} intervals (key at bottom).

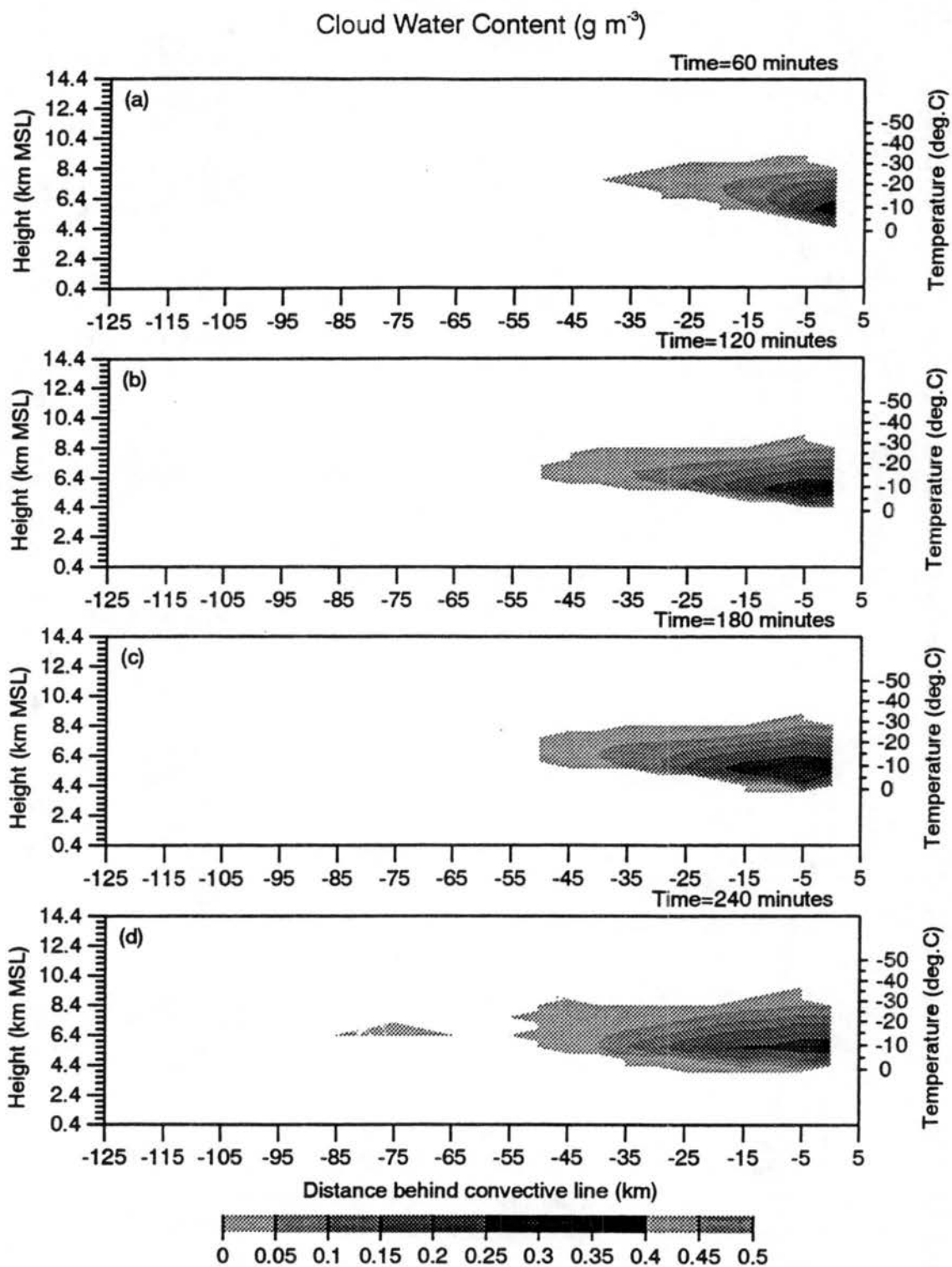


Fig. 6.9 Same as Fig. 6.8 except model output is averaged over two convective pulses.

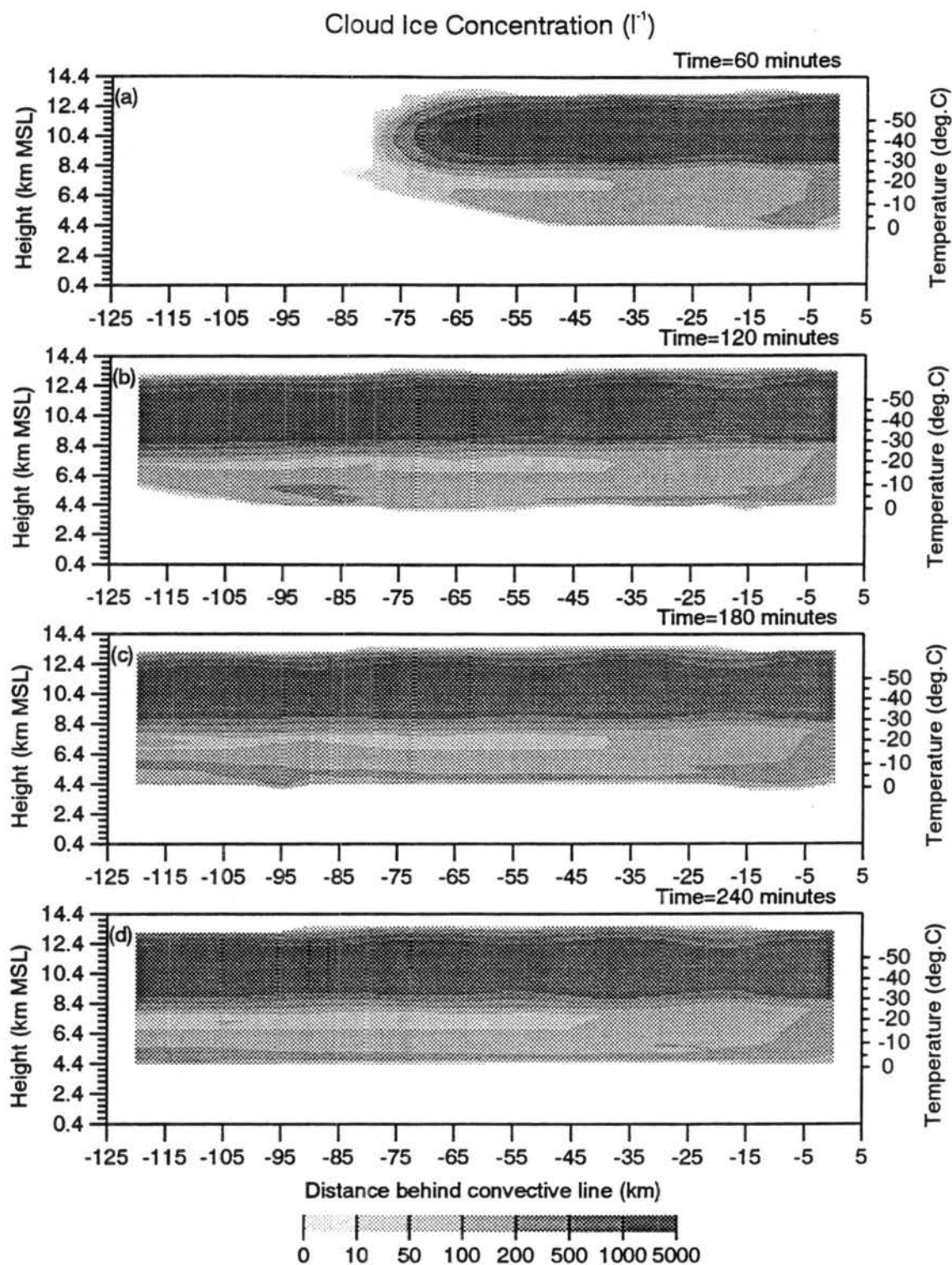


Fig. 6.10. Control run cloud ice concentration (l^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Shading denotes ice concentrations at variable intervals (key at bottom).

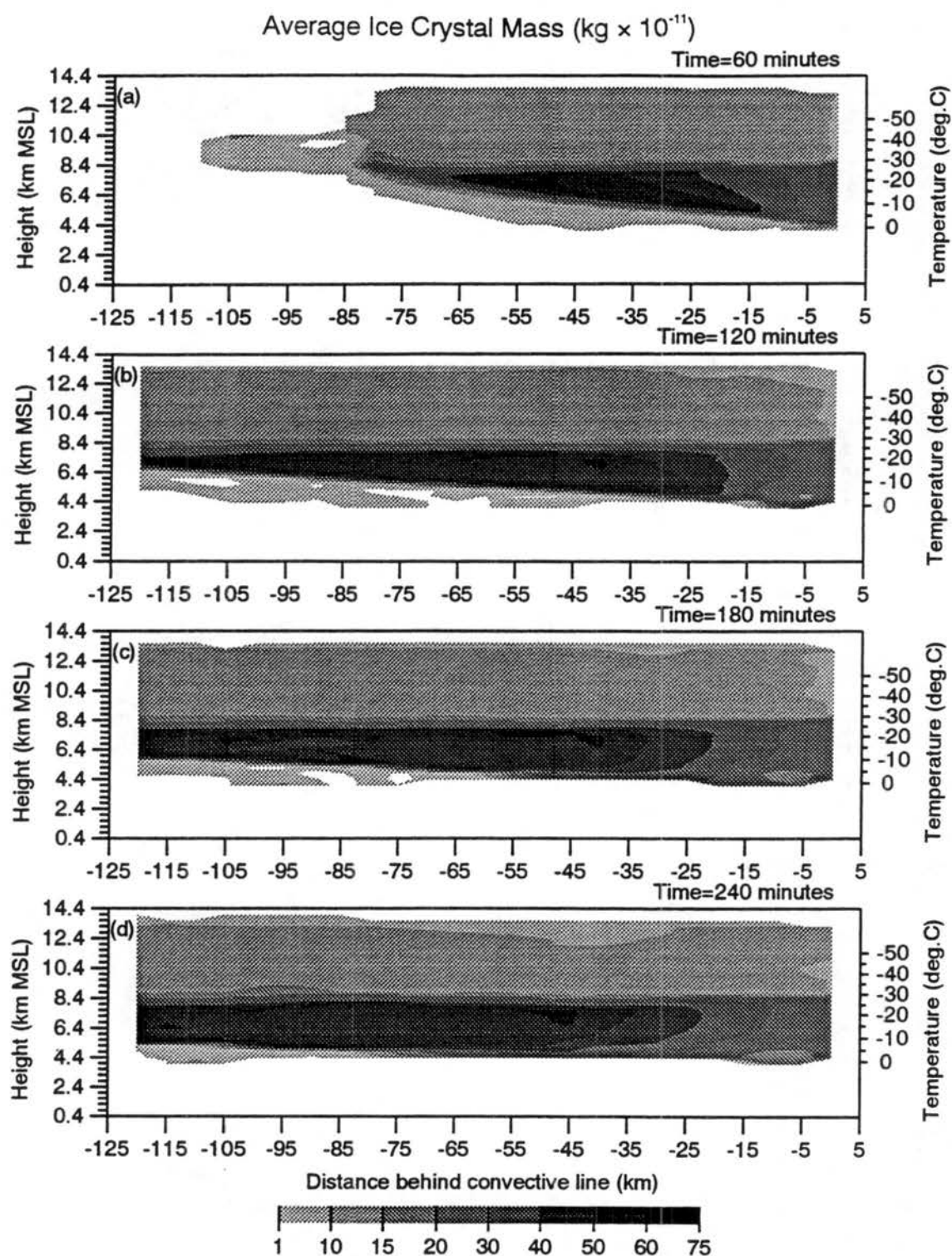


Fig. 6.11. Control run average ice mass ($\text{kg} \times 10^{-11}$) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Shading denotes ice mass at variable intervals (key at bottom).

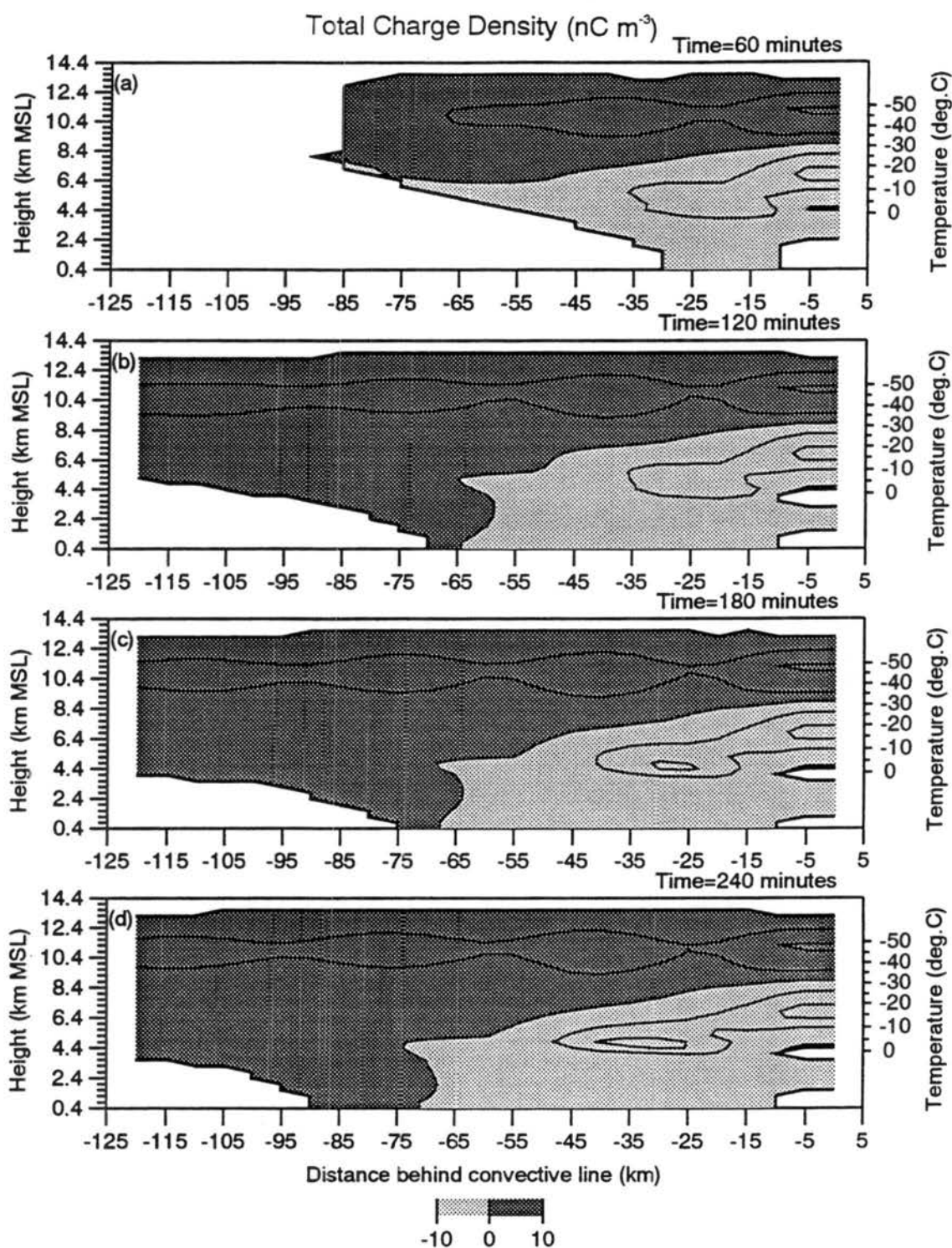


Fig. 6.12. Total charge density magnitudes (nC m^{-3}) for charge advection model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

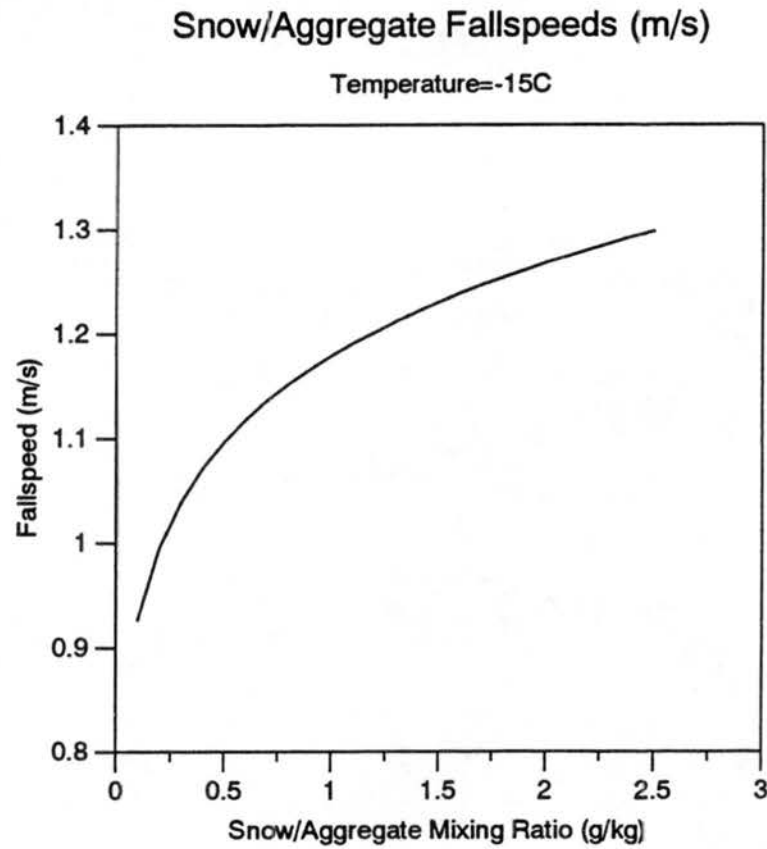


Fig. 6.13. Model snow/aggregate fallspeeds computed using fallspeed coefficient from Appendix B. Fallspeeds are determined using an air density corresponding to the -15°C level.

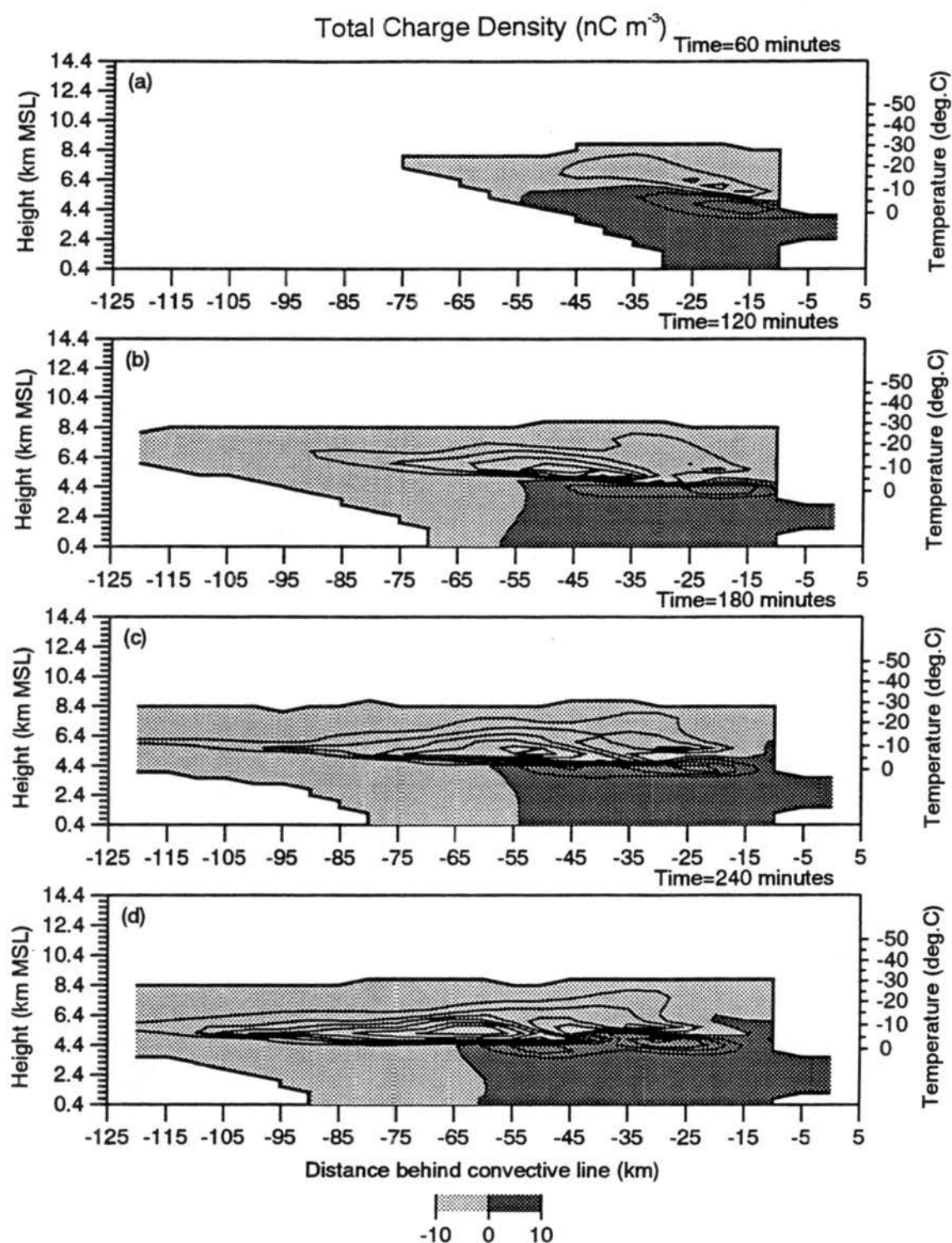


Fig. 6.14. Total charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

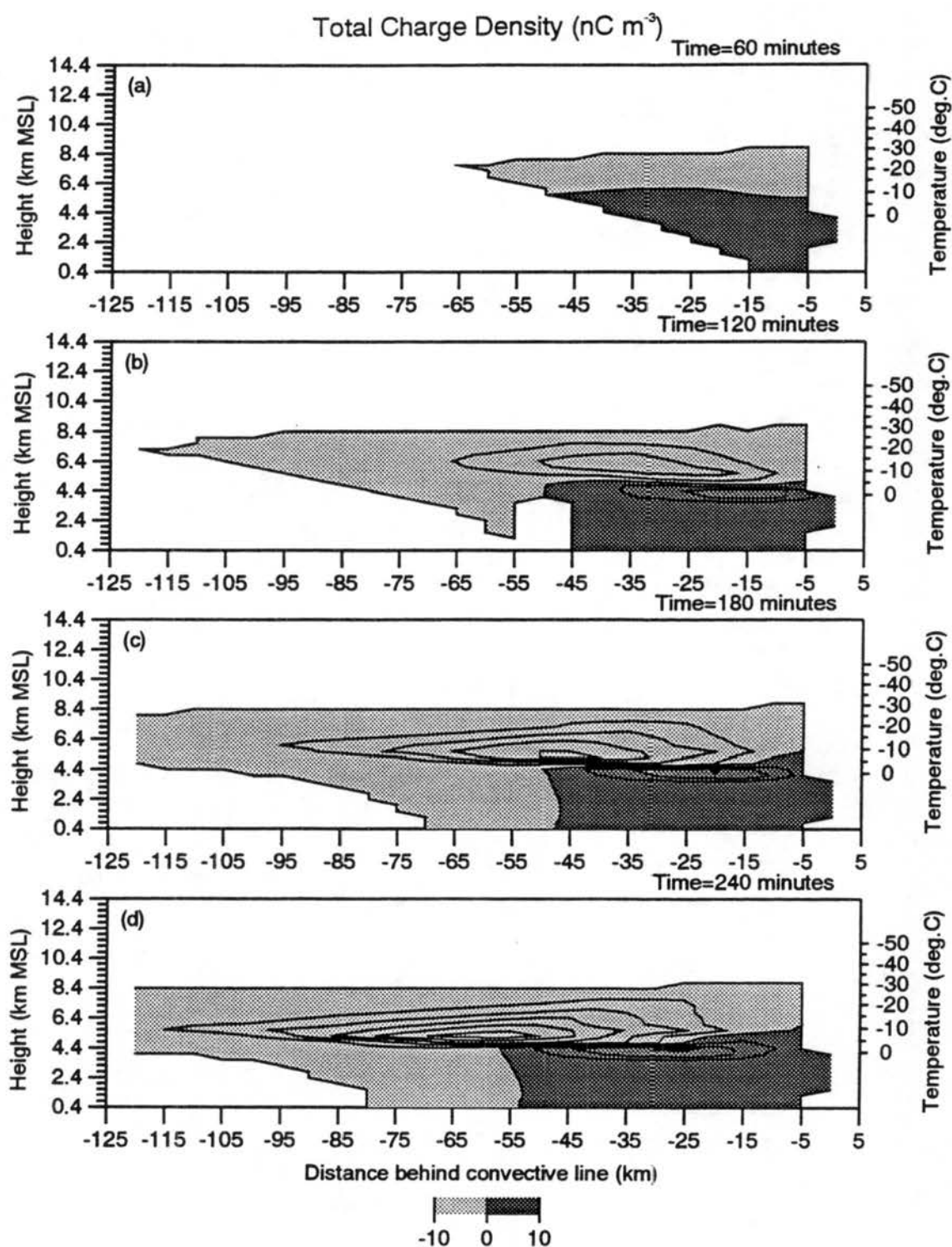


Fig. 6.15. Same as Fig. 6.14 except model output is averaged over two convective pulses.

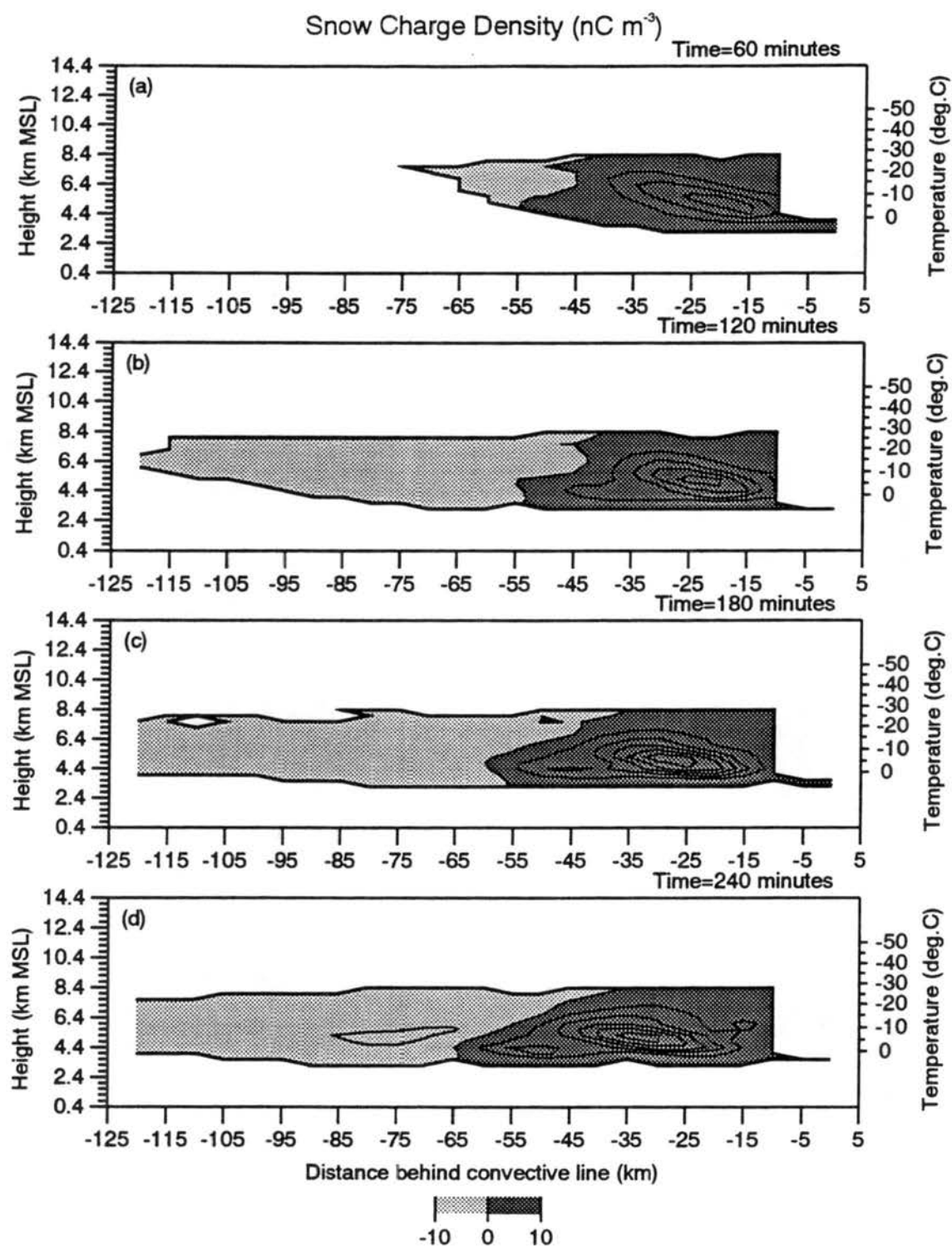


Fig. 6.16. Snow charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

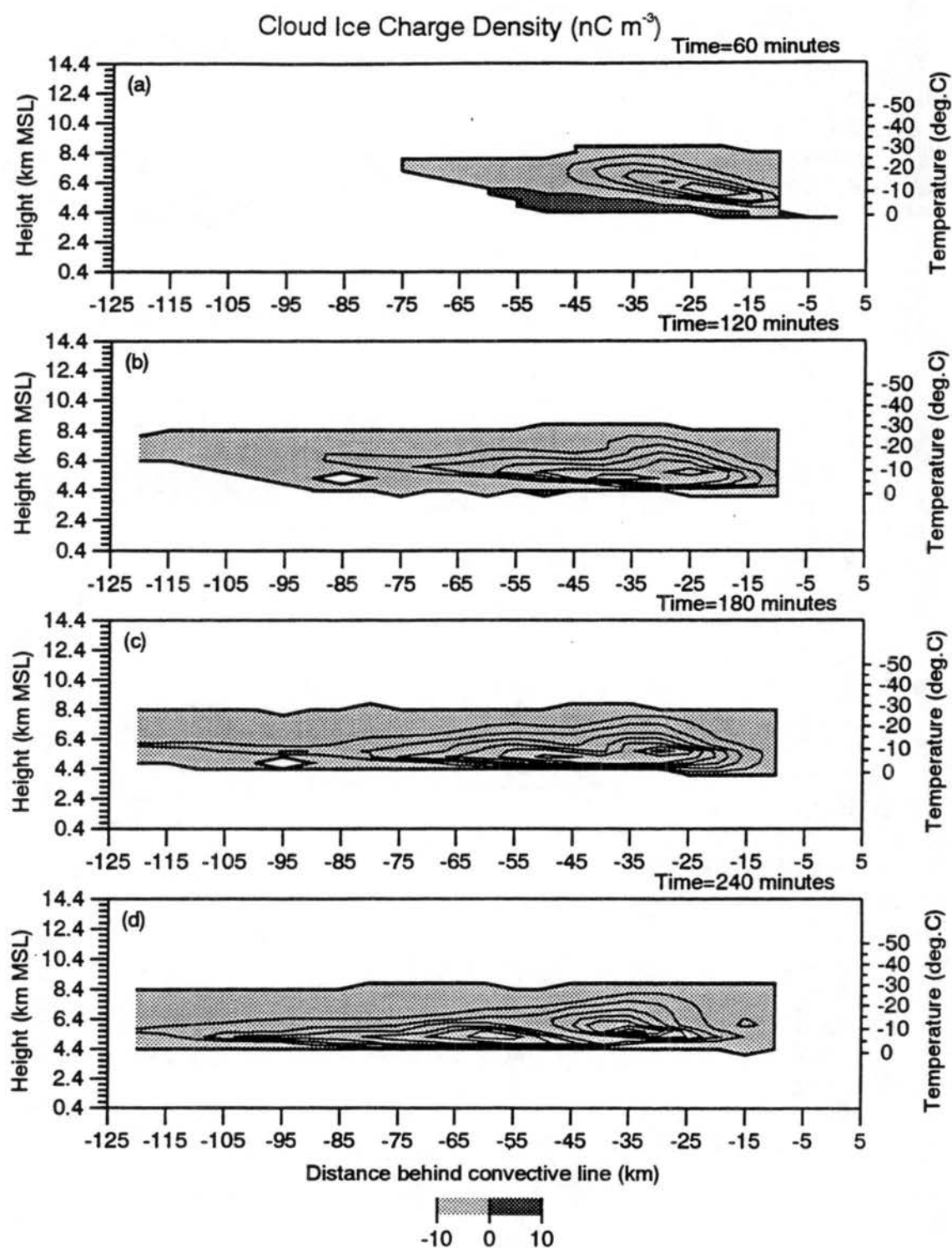


Fig. 6.17. Cloud ice charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

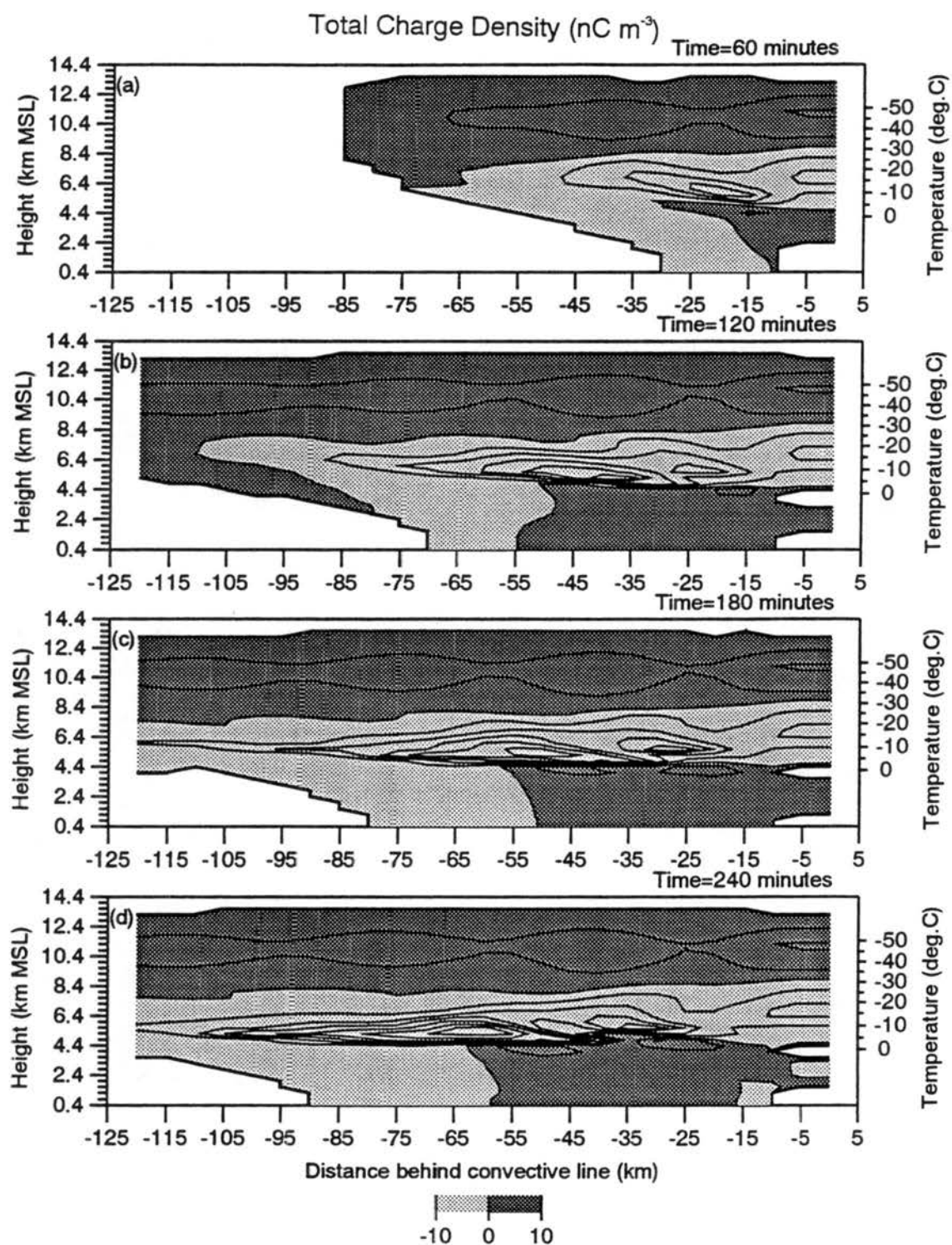


Fig. 6.18. Same as Fig. 6.14 except with the addition of charge advection.

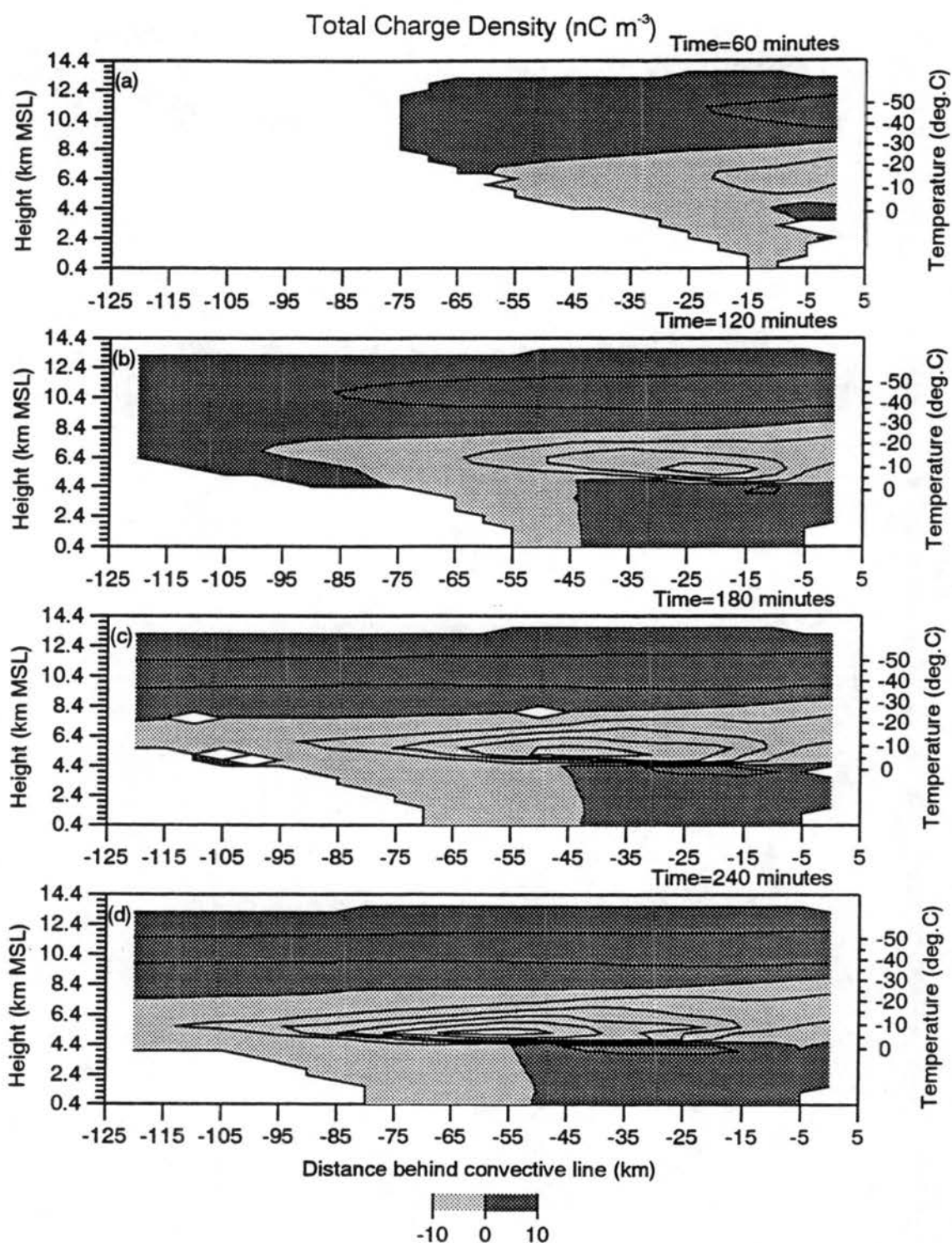


Fig. 6.19. Same as Fig. 6.18 except model output is averaged over two convective pulses.

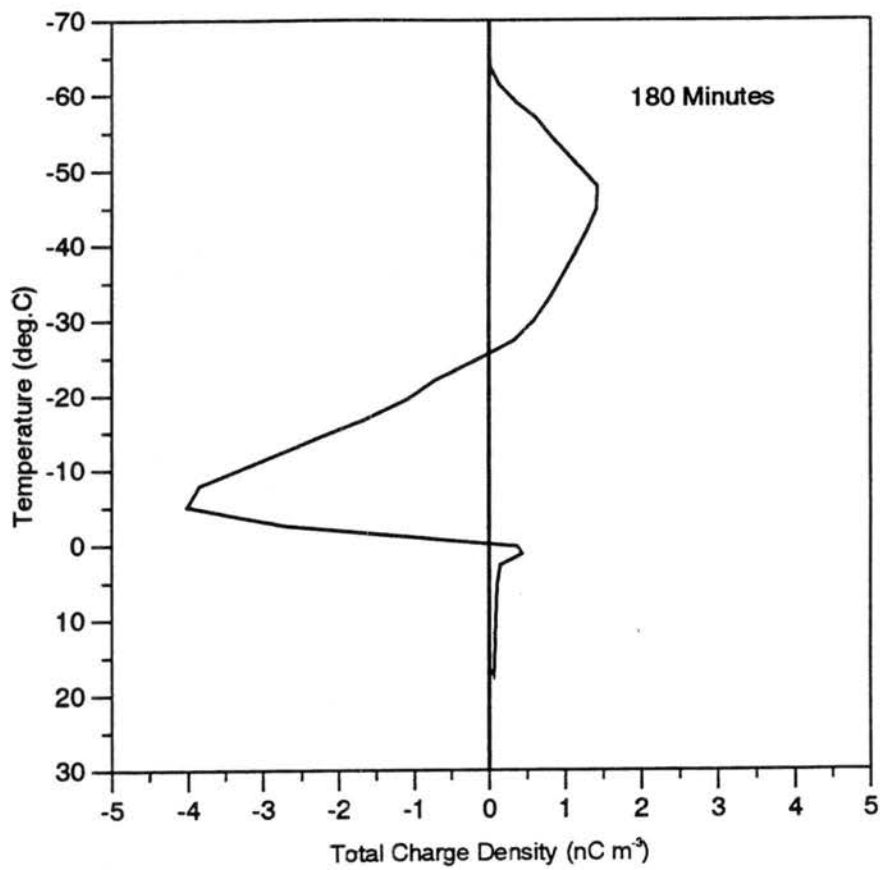


Fig. 6.20. Grid averaged total charge density (nC m⁻³) for the simulation depicted in Fig. 6.18 at 180 minutes into the simulation.

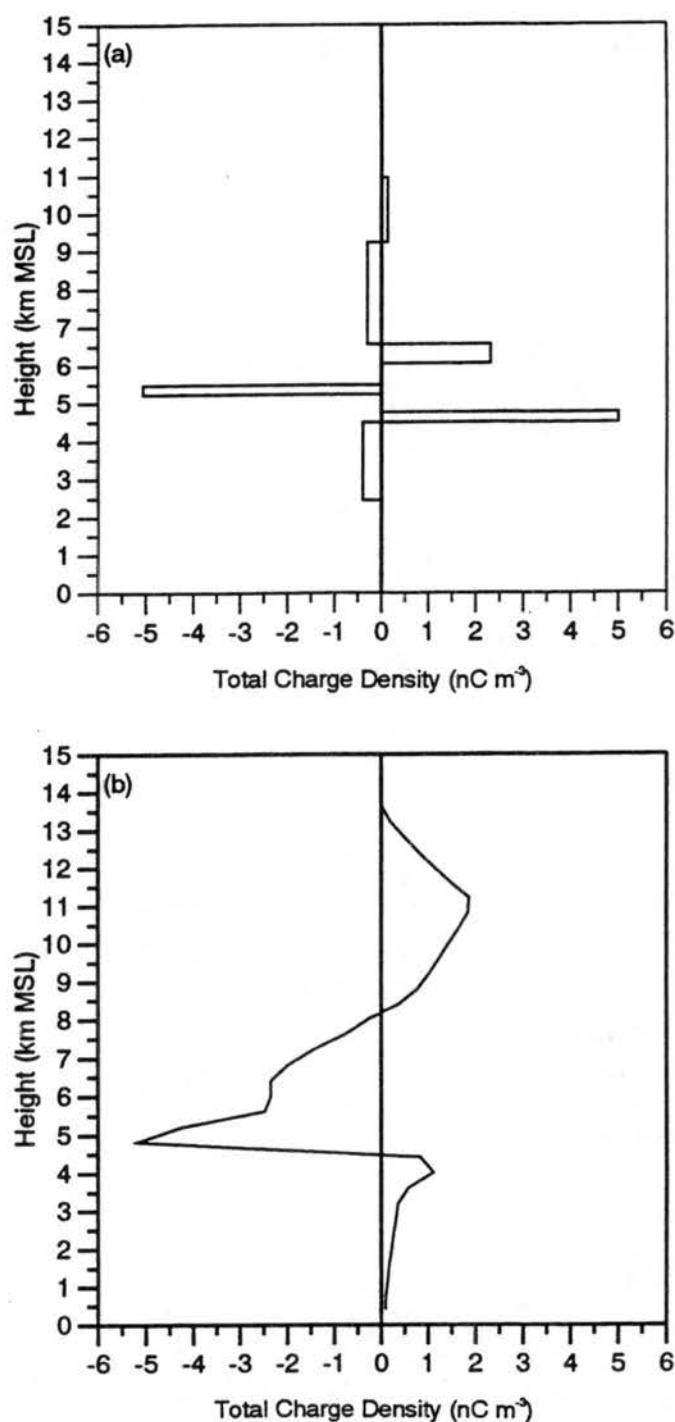


Fig. 6.21. Vertical charge density profile (nC m^{-3}) (a) from symmetric MCS observations (cf. Fig. 3.12), and (b) constructed through the model grid depicted in Fig. 6.18 at 180 minutes into the simulation and a location 35 km behind the convective line. This is a location similar to where the balloon-borne EFM ascended through the symmetric MCS discussed in Chapter 3.

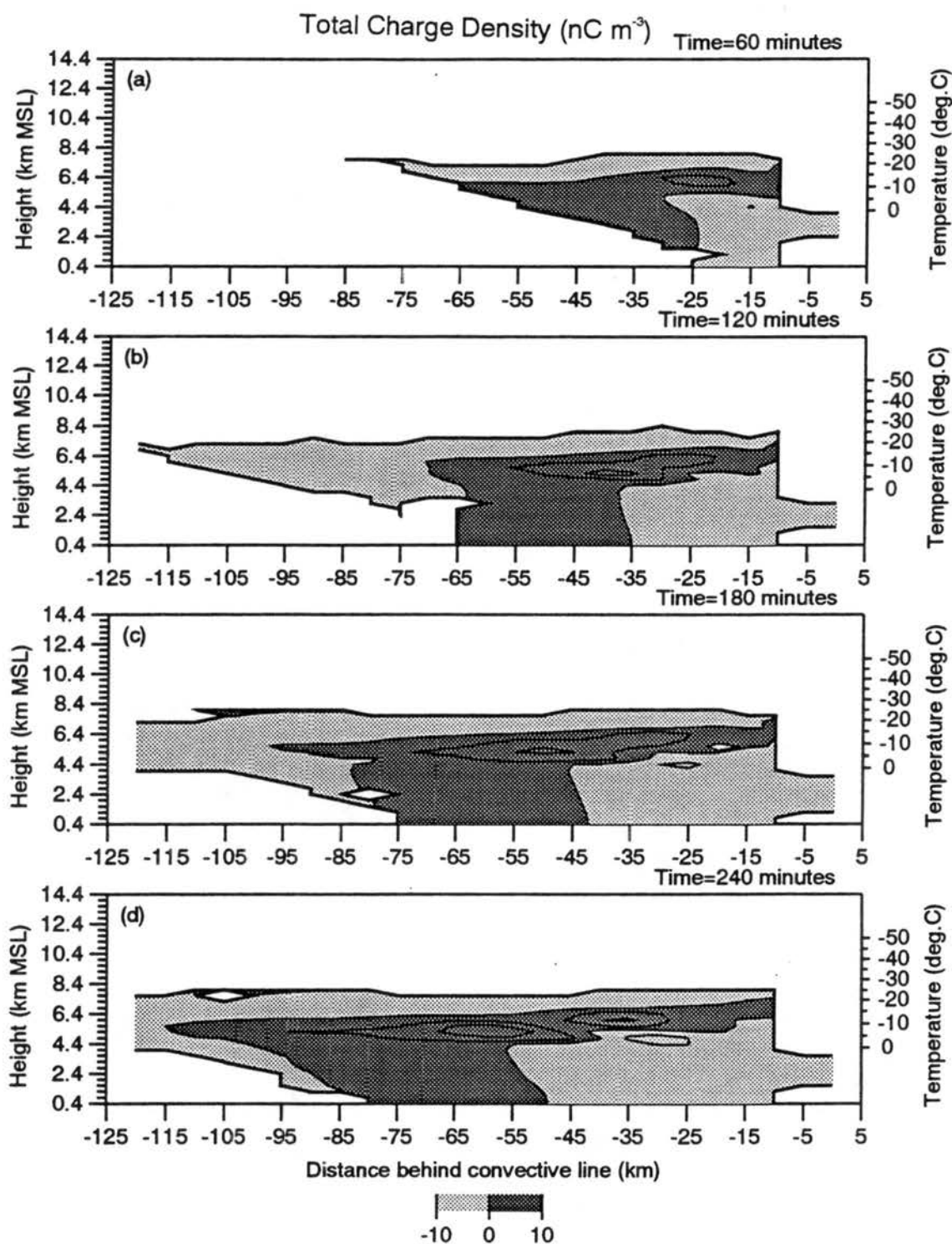


Fig. 6.22. Total charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

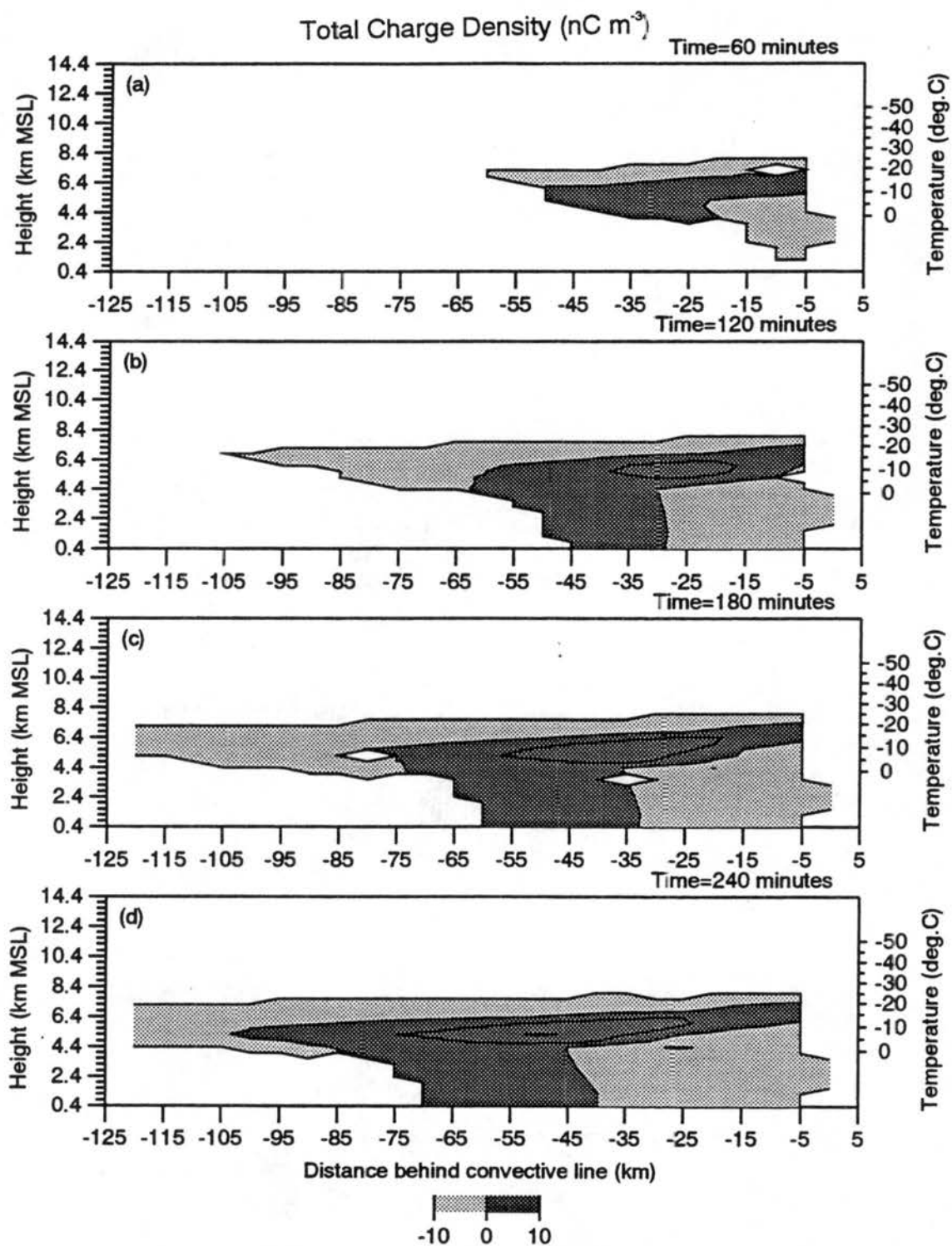


Fig. 6.23. Same as Fig. 6.22 except model output is averaged over two convective pulses.

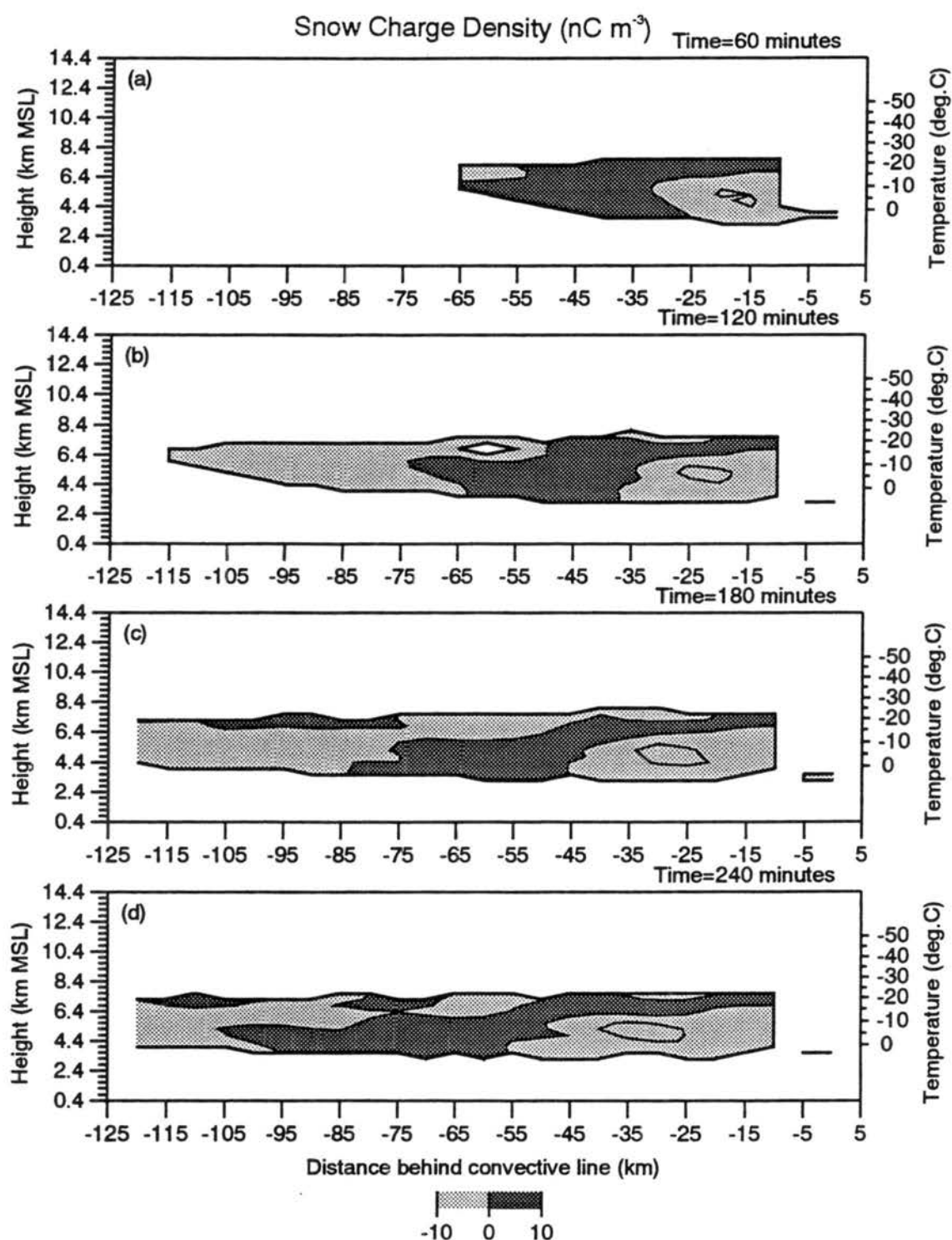


Fig. 6.24. Snow charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

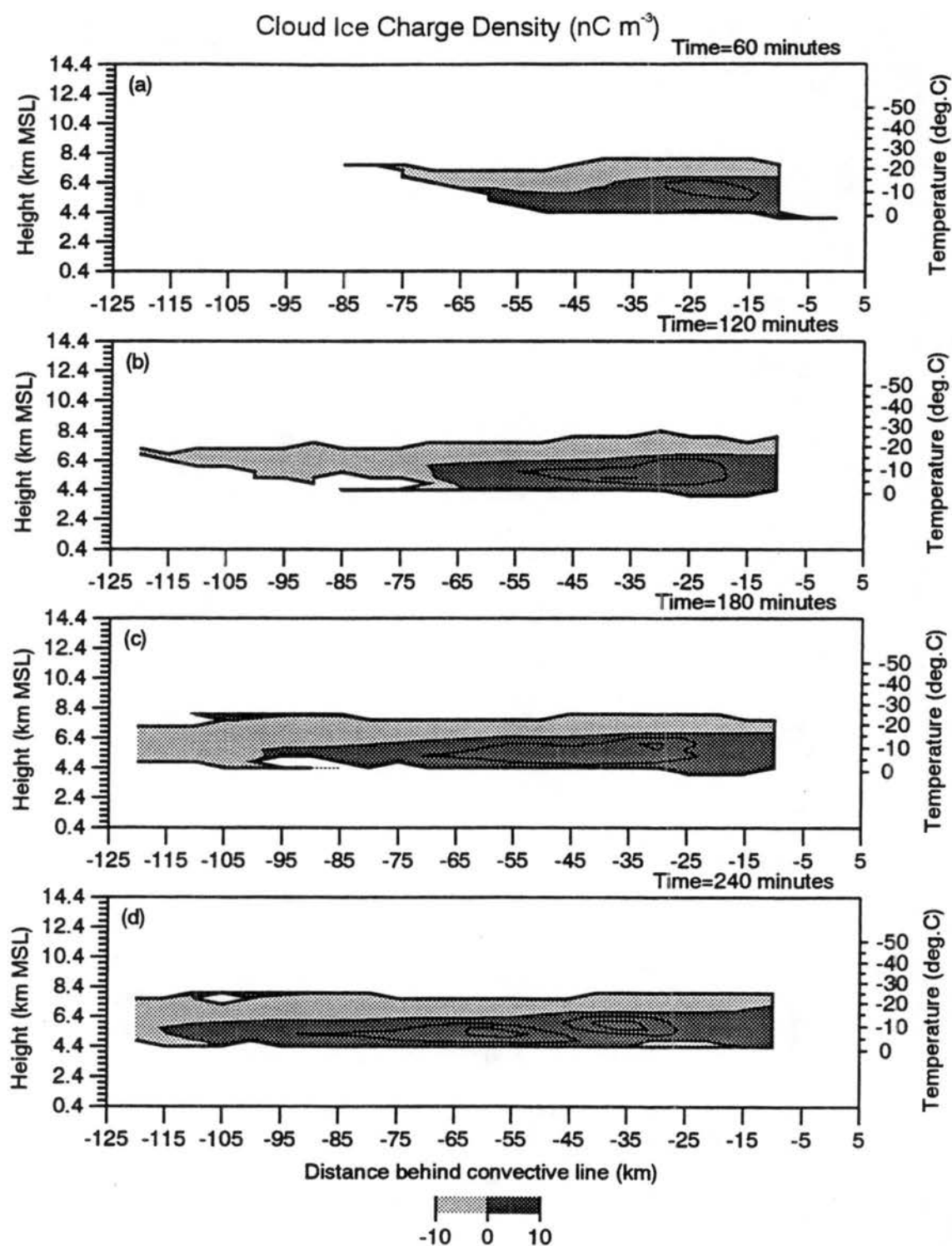


Fig. 6.25. Cloud ice charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

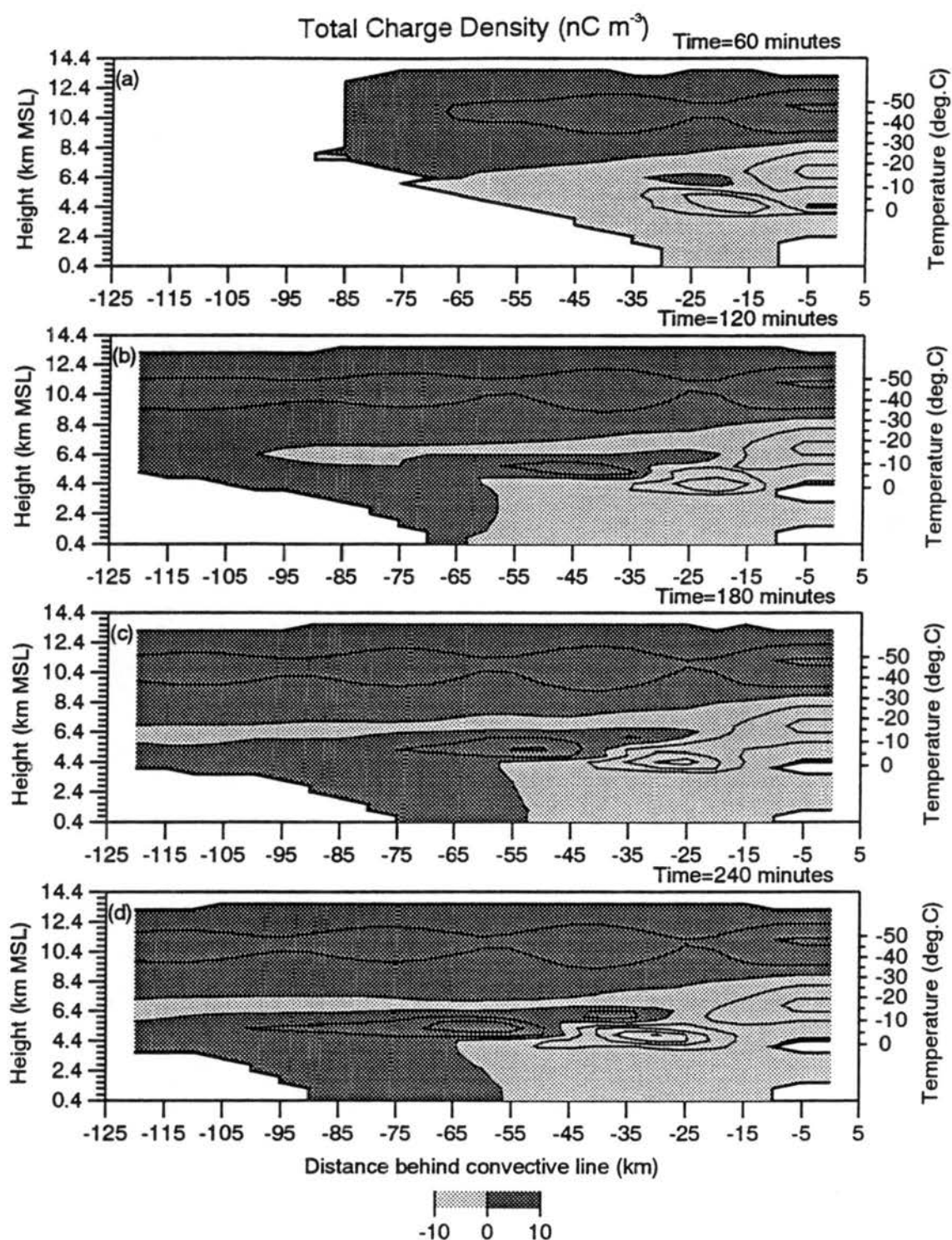


Fig. 6.26. Same as Fig. 6.22 except with the addition of charge advection.

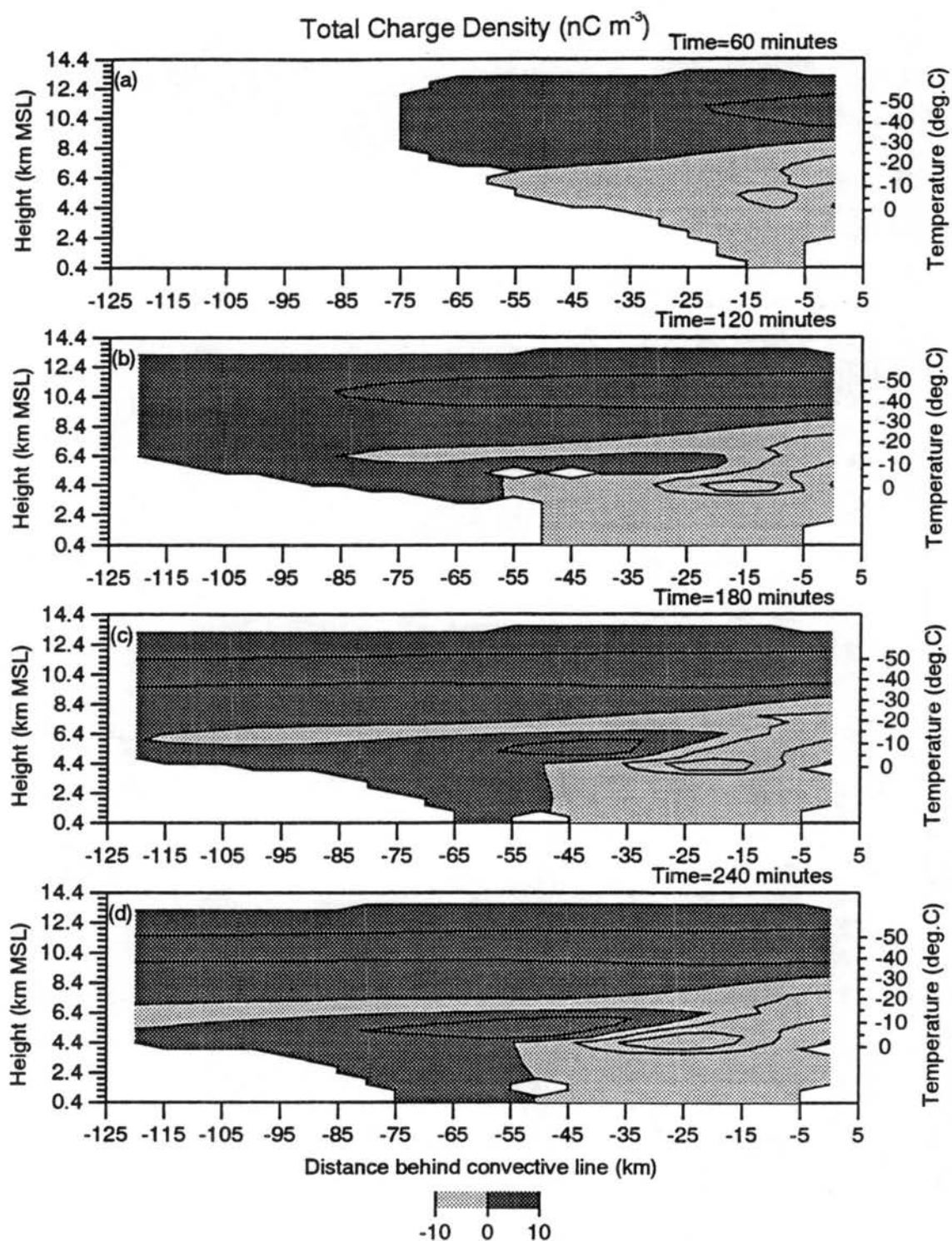


Fig. 6.27. Same as Fig. 6.26 except model output is averaged over two convective pulses.

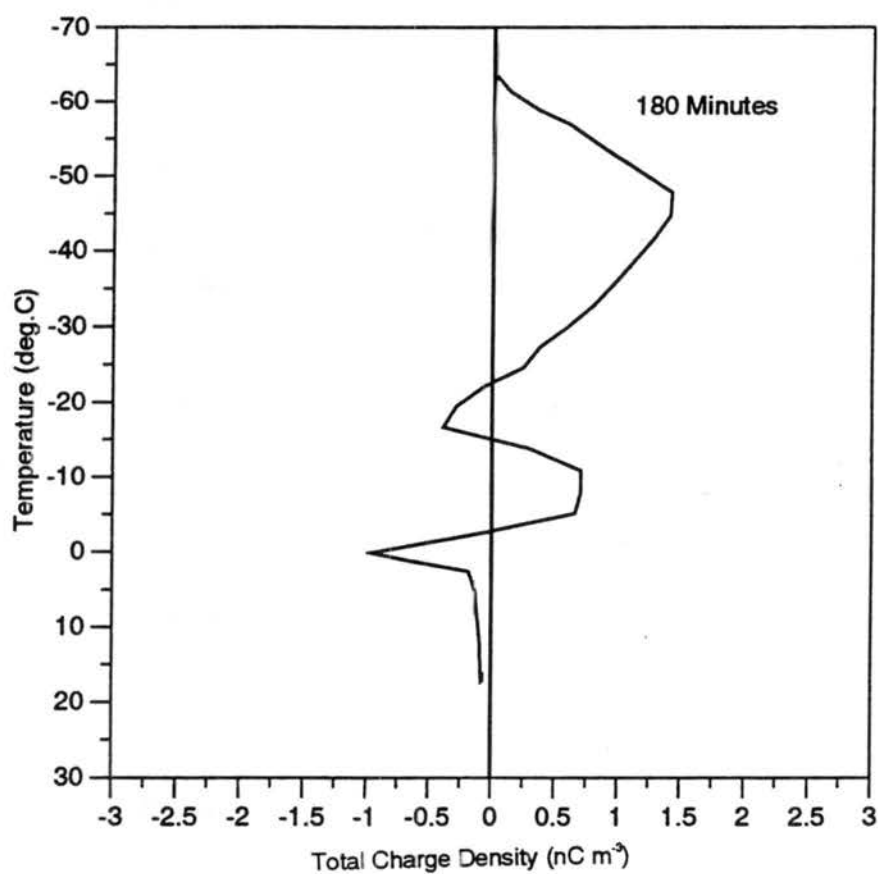


Fig. 6.28. Grid averaged total charge density (nC m^{-3}) for the simulation depicted in Fig. 6.26 at 180 minutes into the simulation.

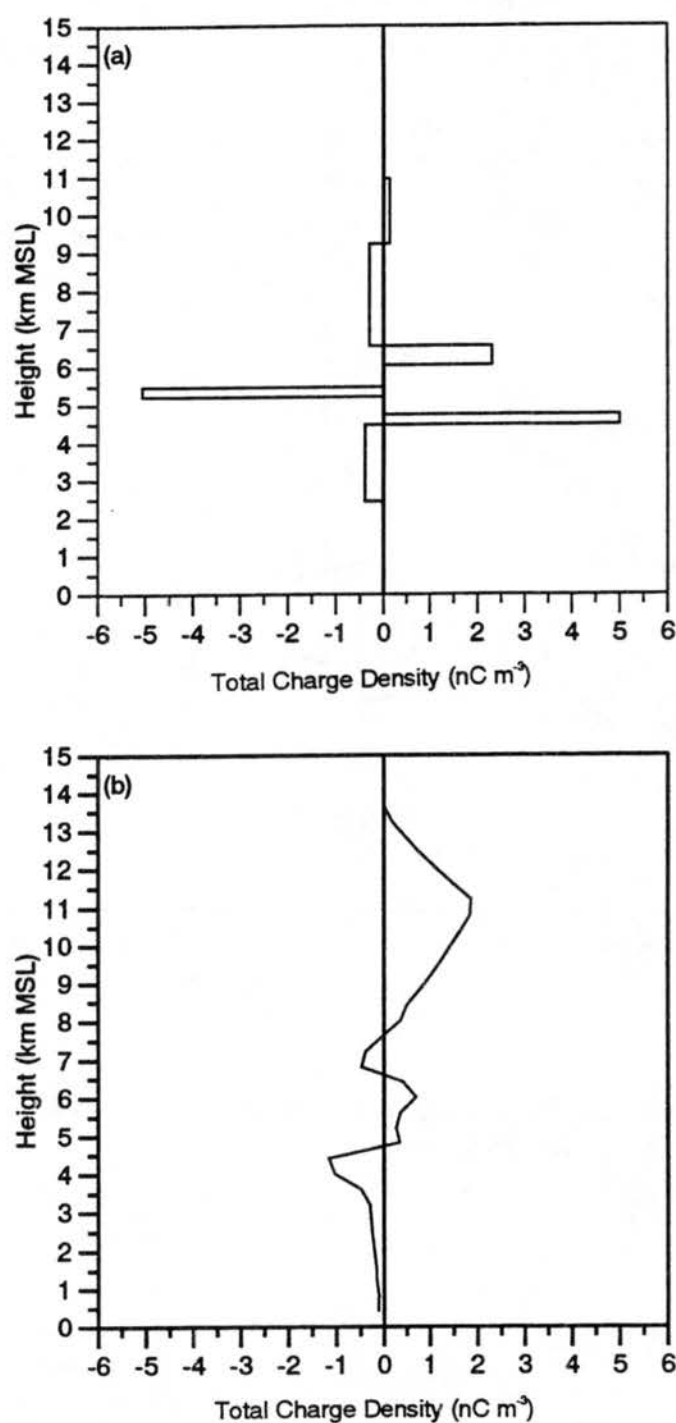


Fig. 6.29. Vertical charge density profile (nC m^{-3}) (a) from symmetric MCS observations (cf. Fig. 3.12), and (b) constructed through the model grid depicted in Fig. 6.18 at 180 minutes into the simulation and a location 35 km behind the convective line. This is a location similar to where the balloon-borne EFM ascended through the symmetric MCS discussed in Chapter 3.

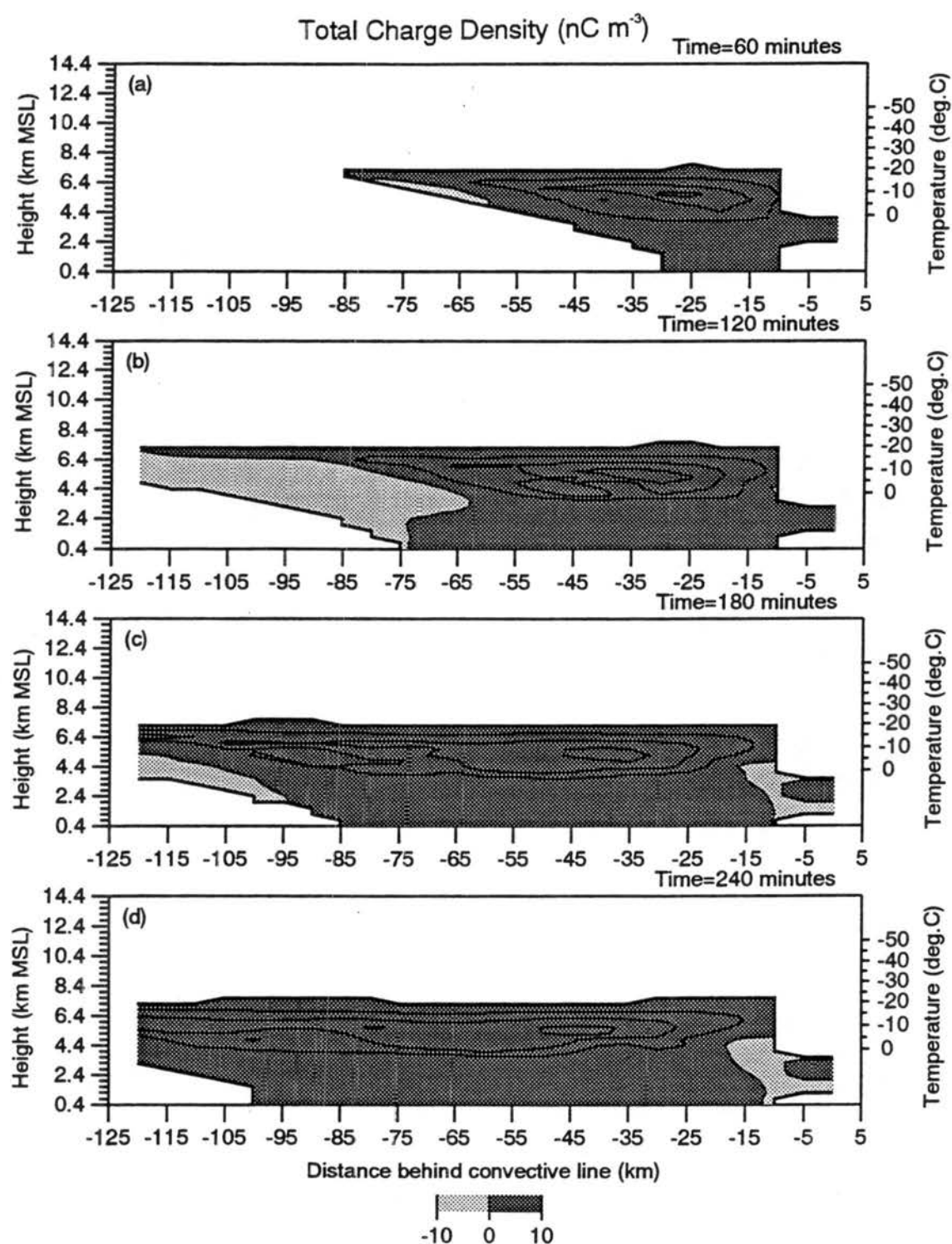


Fig. 6.30. Charge density magnitudes (nC m^{-3}) for the Dong and Hallett (1992) evaporation-condensation charging mechanism at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 0.01 nC m^{-3} .

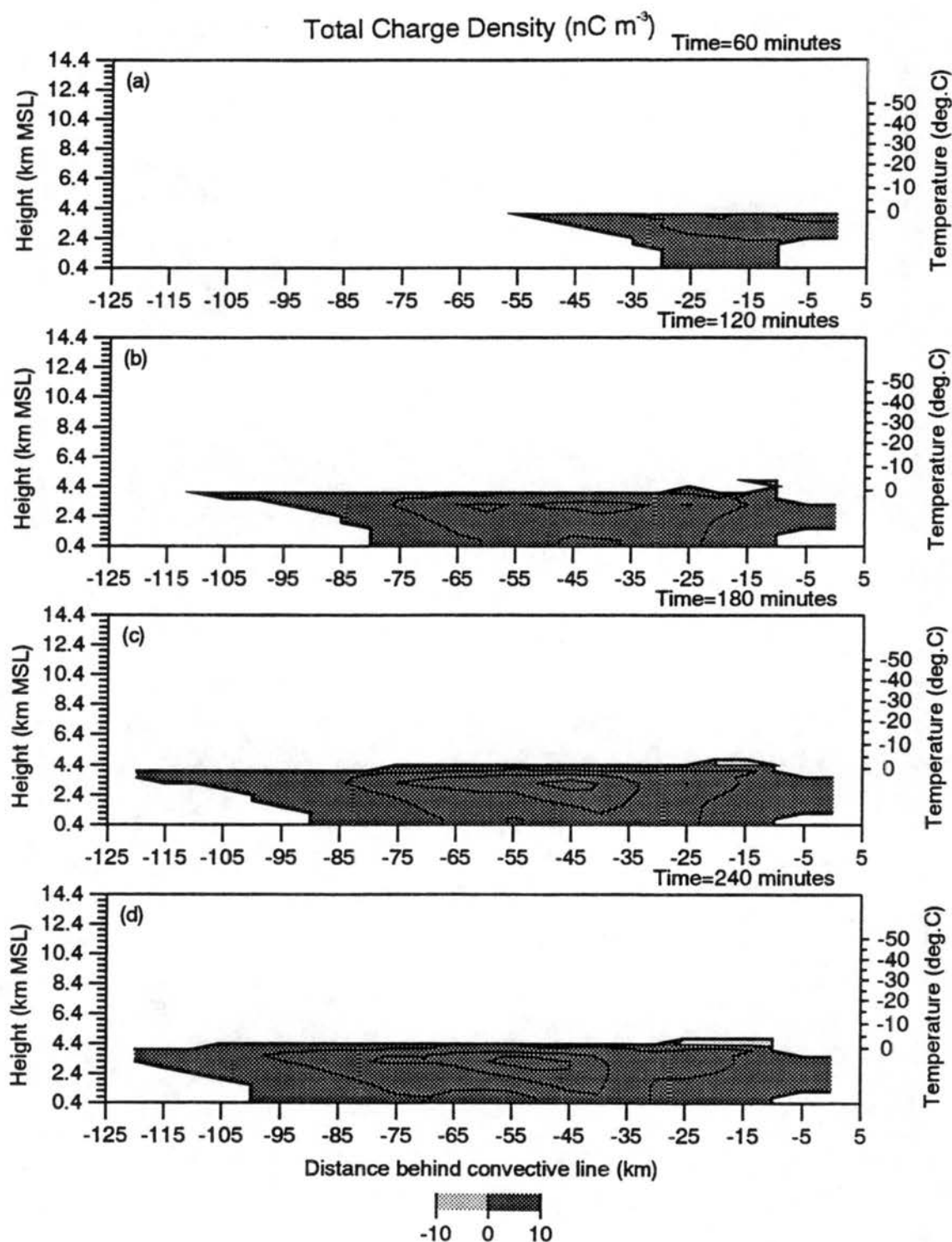


Fig. 6.31. Charge density magnitudes (nC m^{-3}) for the Drake (1968) melting charging mechanism at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 0.1 nC m^{-3} .

CHAPTER 7

MODEL SENSITIVITY TESTS

Initialization conditions used for the electrical control run simulations (presented in Chapter 6) were determined after comparisons with observational data sets. Nevertheless, given the general lack of fine-scale observations within symmetric MCSs and uncertainties associated with the parameterization of microphysical processes (e.g., Verlinde et al., 1990), considerable subjectivity is often involved in the initialization of mesoscale models. Therefore, we tested a variety of initial conditions and microphysical parameters with the goal of gaining insight into the charge budget's sensitivity to model input conditions. Since it is impossible to present a discussion of model sensitivity to all model parameters, our sensitivity studies focus on initial conditions that affect MCS evolution, as well as microphysical processes that appear to have the most significant impact on electrification processes. In addition, we test a charge advection profile that exhibits similarities to the structure derived from our symmetric MCS electric field meter profiles (cf. Fig. 3.12, 3.13).

7.1 Convective pulsing

Gallus (1993) investigated the importance of pulsing both heat and hydrometeors from the convective line. In the sensitivity test presented here, we examine the effect of using a constant advection rate. In this simulation, peak magnitudes of hydrometeor content and heating are reduced to assure that the same amount of water mass and heat enter the model domain as in the pulsed simulations. Furthermore, charge advection is turned off in order to explicitly investigate

the degree to which pulsing modifies the stratiform kinematic and microphysical environment that is responsible for in situ charge generation processes.

Horizontal and vertical velocities from this simulation (hereafter referred to as NP) are depicted by Figs. 7.1 and 7.2, respectively. When compared to the control run (hereafter referred to as CTL, cf. Figs. 6.2 and 6.4), it is apparent that the MCSs two-dimensional flow structure develops at a much slower rate in NP. In particular, the rear inflow jet does not exhibit any velocities $> 5 \text{ m s}^{-1}$ until 180 minutes and peak rear-to-front velocities remain below 10 m s^{-1} . This is in contrast to CTL where regions of rear-to-front velocities of $> 5 \text{ m s}^{-1}$ are quite extensive by 180 minutes. The upper-level front-to-rear airflow pattern is also much slower to develop. An extensive region of front-to-rear flow $> 20 \text{ m s}^{-1}$ does not appear until 240 minutes (compared to 120 minutes for CTL). Peak vertical velocities are also slightly smaller than those in CTL. These differences are similar to the findings of Gallus (1993) who concluded that pulsing the convective input results in an approximate 10% increase in horizontal airflow perturbations.

Figs. 7.3 through 7.5 depict the corresponding snow, cloud LWC, and ice concentration fields for the NP simulation. As can be seen by comparing the snow field to Fig. 6.7, many similarities exist between NP and the hourly averaged pulsed output from CTL. In particular, predicted snow mixing ratios are similar with peak mixing ratios increasing with time as additional water mass is added in the mesoscale updraft. Maximum predicted cloud LWCs are slightly larger than those produced in the pulsed simulations, eventually reaching magnitudes of approximately 0.45 g m^{-3} (though hourly averaged LWCs are in better agreement with observations). A possible reason why LWCs are larger in this simulation might be related to a decrease in collection associated with the reduced amplitude of the advected snow profile (discussed above).

The total charge density for the Takahashi (1978) and Saunders et al. (1991) non-inductive charging parameterizations is depicted by Figs. 7.6 and 7.7, respectively. Overall, the charge distribution produced using these two parameterizations are quite similar to those from CTL (cf. Figs. 6.14 and 6.22). Also, in both cases, the peak charge densities produced by NP are slightly larger than in CTL. This is likely due to the higher predicted LWCs through 180 minutes in NP. These differences, however, are quite small as the bulk charging rate in the MCS stratiform precipitation region (summed over hour three of the simulation) is 1.40 C h^{-1} with the Takahashi data and 0.28 C h^{-1} with the Saunders et al. data. This compares with 1.32 C h^{-1} and 0.33 C h^{-1} for CTL, respectively. Overall, this simulation indicates that convective pulsing has very little effect on the resultant charging rate and distribution.

7.2 Convective heating

Gallus (1993) found that including convective heating in the model (i.e., buffer zone) initialization was necessary to produce a realistic MCS kinematic structure. In this section, we explore the effect that variations in the convective heating initialization profile have on MCS electrical structure and evolution. In particular, we investigate variations in the peak heating rate. As in the previous sensitivity test, charge advection is not included in these simulations to explicitly investigate the degree to which convective heating modifies the stratiform kinematic and microphysical environment that is responsible for in situ charge generation processes. The effect of the convective heating profile on charge advection (through modification of the MCS two-dimensional flow structure), however, is also discussed.

Horizontal and vertical velocities from a simulation in which no convective heat is advected rearward (hereafter referred to as NH) are depicted by Figs. 7.8 and 7.9, respectively. Although no convective heating and simultaneous hydrometeor advection (as tested by Gallus, 1993) is somewhat unrealistic, we test this extreme case in order to determine the degree to which heat

advection from the convective line is responsible for MCS electrical development. As can be seen by comparing Figs. 6.2 with 7.8, the lack of convective heating significantly alters the MCSs two-dimensional kinematic structure. Unlike in CTL where rear-to-front flow spans the breadth of the stratiform precipitation region by 60 minutes, rear-to-front flow does not appear in NH until 180 minutes. The upper-level front-to-rear airflow regime is also much slower to develop. Whereas an extensive region of front-to-rear flow of $> 20 \text{ m s}^{-1}$ develops in CTL by 120 minutes, front-to-rear flow in NH varies little from the initialization wind profile throughout the first 240 minutes of the simulation. This reduction in front-to-rear flow also affects charge advection rates from the convective line (discussed later). Qualitatively, the vertical velocity structure produced by NH is also quite different though, in general, peak vertical velocities produced by NH are only slightly smaller than in CTL.

Figs. 7.10 through 7.12 depict the corresponding snow, cloud LWC, and ice concentration fields, respectively. As can be seen by comparing Figs. 6.6 and 7.10, peak mixing ratios in the snow category are noticeably smaller in NH than in CTL. Since the peak pulse amplitude of the advected snow profile is the same for each of these simulations, this suggests that water mass from the snow category is being depleted by sublimation at locations immediately behind the convective line. Cloud LWCs are also noticeably smaller in the NH simulation. Although LWCs of $> 0.4 \text{ g m}^{-3}$ are present at locations immediately behind the convective line, LWCs at distances greater than 15 km from the convective line are typically $< 0.15 \text{ g m}^{-3}$. This is in contrast to LWCs from CTL, which were occasionally $> 0.25 \text{ g m}^{-3}$ at similar locations.

The total charge density for the Takahashi (1978) and Saunders et al. (1991) non-inductive charging parameterizations is depicted by Figs. 7.13 and 7.14, respectively. As a result of the smaller snow mixing ratios and lower LWCs in NH, peak charge densities produced by NH are significantly smaller in CTL. The overall distribution of charge, however, is essentially

unchanged. When integrated over the stratiform precipitation region, the charging rate in NH is 1.00 C h^{-1} (compared to 1.32 C h^{-1} for CTL) for the Takahashi (1978) data and 0.16 C h^{-1} (compared to 0.33 C h^{-1} for CTL) for the Saunders et al. (1991) data. As noted earlier, the lack of convective heating in this simulation also affects the advection of charge from the convective line. A simulation that investigated this contribution indicated a reduction of charge advection by 17% (from 0.52 C h^{-1} to 0.43 C h^{-1}) when convective heating was neglected.

We also investigate the effect of increasing the peak heating rate in the convective initialization profile. Gallus (1993) noted that their rawinsonde-based heating profile probably underestimated the magnitude of heating within convective cores. The more recent study of Braun and Houze (1996) indicates that the peak heating rate in the convective line of another symmetric MCS was as high as 40°C h^{-1} (see their Fig. 21). Thus, we test this higher heating rate in our model by doubling the heating rates at all levels of our initialization profile (cf. Fig. 5.2). The result of this modification on the evolution of the MCS horizontal and vertical velocity structures is shown in Figs. 7.15 and 7.16, respectively. Interestingly, the results of this simulation are quite similar to those from NH (cf. Figs. 7.8 and 7.9). Most notably, rear-to-front flow does not develop until 120 minutes into the simulation, never exceeds 5 m s^{-1} , and never extends across the entire breadth of the stratiform precipitation region. At upper levels, unlike in CTL, front-to-rear flow speeds remain below 20 m s^{-1} . Microphysical and electrical characteristics of this simulation (not shown) also exhibited many similarities to NH.

From these simulations, we conclude that the inclusion of convective heating has a significant impact on MCS stratiform region evolution. Although all of the dynamic feedbacks associated with the choice of a convective heating rate are not evident, variations on either side of the 20°C h^{-1} maximum heating rate used in CTL, which produced the most realistic kinematic structure, resulted in significant modifications to the MCS flow structure. Simulations that

investigated the height of maximum heating (not shown) also demonstrated a sensitivity. Thus, it appears that the convective lines influence on the stratiform precipitation regions electrical development is not necessarily limited to advection of charge on hydrometeors. That is, contributions from the convective line in the form of heating can modify in situ thermodynamic and microphysical contributions which, in turn, can potentially result in modifications to the MCS's electrical structure and development. Although we were not able to model the more three-dimensional asymmetric MCS with this model, it is possible that decreased heat and hydrometeor advection late in an MCSs life cycle (at which time an asymmetric structure is often observed) might result in a simultaneous modification of the MCS stratiform precipitation region's thermodynamic, microphysical, and electrical structures. Unfortunately, a better understanding of these dynamic and microphysical feedbacks would require a more detailed numerical model than the one used in this study.

7.3 Collection efficiencies

Finally we investigate the effect that snow/ice collection efficiency has on the MCS electrical structure. As noted in Chapter 5, sensitivity tests using explicit microphysical code indicated that several factors often work to overestimate collection rates (by as much as order of magnitude). These include the assumption of continuous collection in the derivation of mass weighted fallspeeds, the choice of fallspeed coefficients, and the choice of precipitation hydrometeor category shape parameters. To correct for this overestimation, in CTL we used smaller (when compared to Gallus, 1993) collection efficiencies for all processes that involved the collection of monodisperse hydrometeor categories (e.g., cloud water, cloud ice) by exponential hydrometeor categories (e.g., snow/aggregate). In the sensitivity test presented here, we use the larger snow/ice collection efficiency ($E_{si} = 0.1$) that was used in Gallus (1993). To

accurately predict cloud LWCs, however, other collection efficiencies were maintained at their CTL values (see Appendix B).

Horizontal and vertical velocities from this simulation (hereafter referred to as CE) are depicted by Figs. 7.17 and 7.18, respectively. When compared against CTL (cf. Figs. 6.2 and 6.4), it is apparent that this modification has very little effect on the MCS kinematic structure and evolution. The MCS microphysical evolution, however, exhibits many differences. Figs. 7.19 through 7.21 depict the evolution of the snow, cloud LWC, and cloud ice fields, respectively. Fig. 7.22 depicts the average predicted ice mass of CE. Due to the collection of large amounts of cloud ice at locations immediately behind the convective line, maximum snow mixing ratios in CE are much larger and upper-level ice concentrations much smaller than in CTL. Though cloud ice is also collected at a higher rate at lower levels, it is also quickly replenished by fragment generation (see Chapter 5). Since cloud ice added by the fragmentation function is typically smaller than the ice removed by the collection process (whose mass has been enhanced with time by depositional growth), this process eventually results in an average ice crystal mass that is somewhat smaller than in CTL. Cloud LWCs are quite similar to those in CTL (cf. Fig. 6.8).

Figs. 7.23 and 7.24 depict the total charge density from CE for the Takahashi (1978) and Saunders et al. (1991) non-inductive parameterizations, respectively. Although the change is small, the increased collection efficiency in CE has a significant impact as maximum charge densities in CE are only approximately 30% of those from CTL. It might be recalled that the higher charge densities in CTL were primarily the result of a residual negative charge on the smaller ice crystals that essentially mask the opposite polarity of charge on the more rapidly falling snow category. Since these ice crystals are removed at a much faster rate in CE, and furthermore since the fragmentation typically contributes a net positive charge to the ice

category, the negative charge on the ice is significantly smaller in CE and charge densities are more similar to those indicated by the observations (cf. Tables 3.1-3.4). As noted earlier, an additional effect of the increased collection efficiency was to decrease the predicted ice crystal size. Thus, the Saunders et al. parameterization, which is dependent on predicted ice size, generates a smaller charge transfer per collision.

Despite these changes, the overall distribution of charge remains essentially unchanged from that generated by CTL. The Takahashi parameterization produces an inverted dipole structure with a sharp charge transition near the melting level. The Saunders et al. parameterization produces a positive layer of charge in the midlevels (though charge densities are $< 1 \text{ nC m}^{-3}$). When integrated over the stratiform precipitation region, the charging rate in CE is 0.96 C h^{-1} (compared to 1.32 C h^{-1} for CTL) for the Takahashi (1978) data and 0.20 C h^{-1} (compared to 0.33 C h^{-1} for CTL) for the Saunders et al. (1991) data. From this sensitivity test, it is apparent that collection (and separation) efficiency has a significant impact on model predicted charge densities and charge generation rates.

7.4 Charge advection

In this sensitivity study (hereafter referred to as CA), we test the advection of a charge profile that more closely resembles the charge structure analyzed from our symmetric MCS observations (cf. Figs. 3.12, 3.13 and Tables 3.1, 3.2). In simulations CA1 and CA2, we combine this new charge advection profile with non-inductive charge production by the Takahashi and Saunders et al. parameterizations, respectively. The charge initialization profile for CA is depicted by Fig. 7.25. Since very little particle charge data are available within convective cells, it is very difficult to determine how the total charge in this initialization profile should be distributed between the two (snow/aggregate and cloud ice) hydrometeor categories. Since the balloon-borne EFM data in the symmetric MCS indicated predominantly negative

charge in the upper levels, however, we have assigned negative charge to the ice category in this initialization profile. At lower levels, where EFM data indicated a complicated charge structure consisting of alternating layers of positive and negative charge, we primarily assign this charge to the snow category. In the lower levels, this charge profile exhibits many similarities to the non-updraft charge structures presented in Stolzenburg (1996).

The evolution of the total charge density in CA is depicted by Fig. 7.26. Though the charge depicted in Fig. 7.25 is initialized at slightly colder temperatures than indicated by the EFM observations, descent (at a rate of $0.9\text{--}1.2\text{ m s}^{-1}$, cf. Fig. 6.13) of the charge on the snow category as it crosses the transition zone results in the mid-level positive charge layer (which is centered at approximately -25°C in the initialization profile, cf. Fig. 7.25) being situated entirely below -15°C at a location 35 km behind the convective line. This temperature level for the positive charge layer is quite similar to an analogous layer of charge (at a similar system-relative location) in the EFM observations (cf. Figs. 3.12, 3.13). Subsequent descent of this charge region, however, quickly results in the dissipation of the layered charge structure. Since several studies have shown stratiform charge structure to be spatially extensive (e.g., Marshall and Rust, 1993; Stolzenburg et al., 1994), this suggests either that charge advection is insufficient to explain observed structures at locations > 50 km from the convective line or that the majority of the charge is carried on smaller ice particles (with correspondingly smaller fallspeeds). At upper levels where the majority of the charge resides on cloud ice, the initialization charge structure is essentially maintained across the breadth of the model grid. It is interesting to note that, though turbulent diffusion was included in this model simulation, very little mixing of the upper-level positive charge layer occurred as it advected across the model grid.

The evolution of total charge density in CA1 is depicted by Fig. 7.27 while Fig. 7.28 shows the vertical charge density profile derived from the second symmetric MCS electric field profile

and a vertical cross-section of charge density constructed through the CA1 model grid at a location 35 km behind the convective line (analogous to where the balloon-borne EFM ascended). Figs. 7.29 and 7.30, respectively, depict similar analyses for CA2. At locations > 25 km behind the convective line, the CA1 mixed-phase region charge structure exhibits marked similarities to that depicted by Fig. 6.18 (the simulation in which Takahashi non-inductive charging was combined with the advection of the much simpler tripole thunderstorm charge structure). Given the rather large charging rates associated with the Takahashi parameterization, charging at locations immediately behind the convective line in CA1 resulted in the rapid accumulation of negative charge on the cloud ice, thereby eliminating the positive charge layer in the advected profile by a location 25 km behind the convective line. In CA2, on the other hand, where non-inductive charging rates from Saunders et al. are much smaller than those in CA1, charging at locations immediately behind the convective line appears to enhance the advected charge profile. By 180 minutes, the combination of charge advection and non-inductive charging in CA2 produces a rather sharp charge transition of positive charge above negative (with a charge transition at approximately -8°C) that exhibited marked similarities to the observations (cf. Figs. 3.12, 3.13). Unfortunately, as noted earlier, it is very difficult to determine how the total charge in the initialization profile should be distributed between the two (snow/aggregate and cloud ice) hydrometeor categories. Therefore, a more complete investigation of model sensitivity to advected charge structure would require improved in situ observations of particle charge within MCS convective lines.

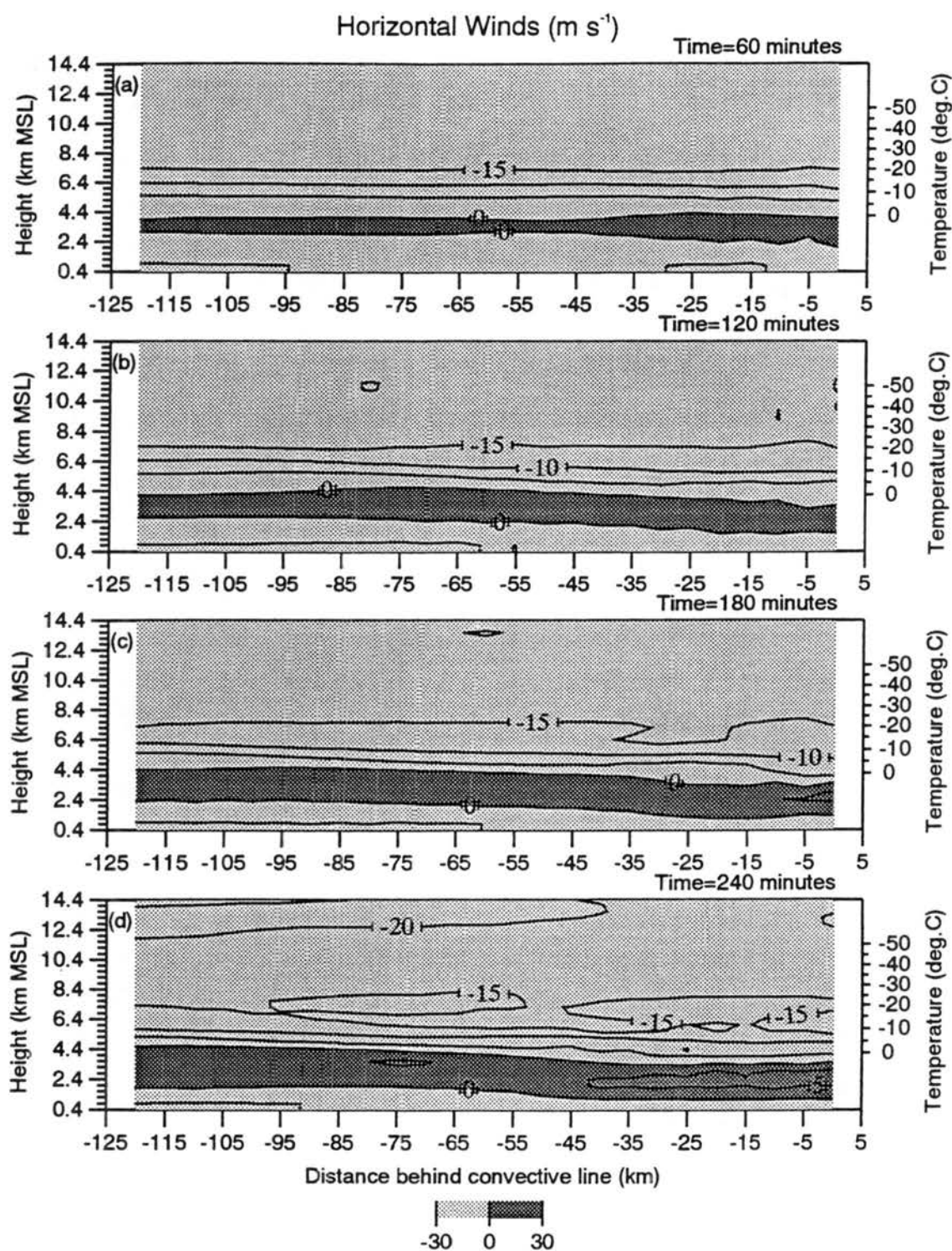


Fig. 7.1. Horizontal winds (m s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation. Light shading denotes front-to-rear flow while dark shading denotes rear-to-front flow. Contours are every 5 m s^{-1} .

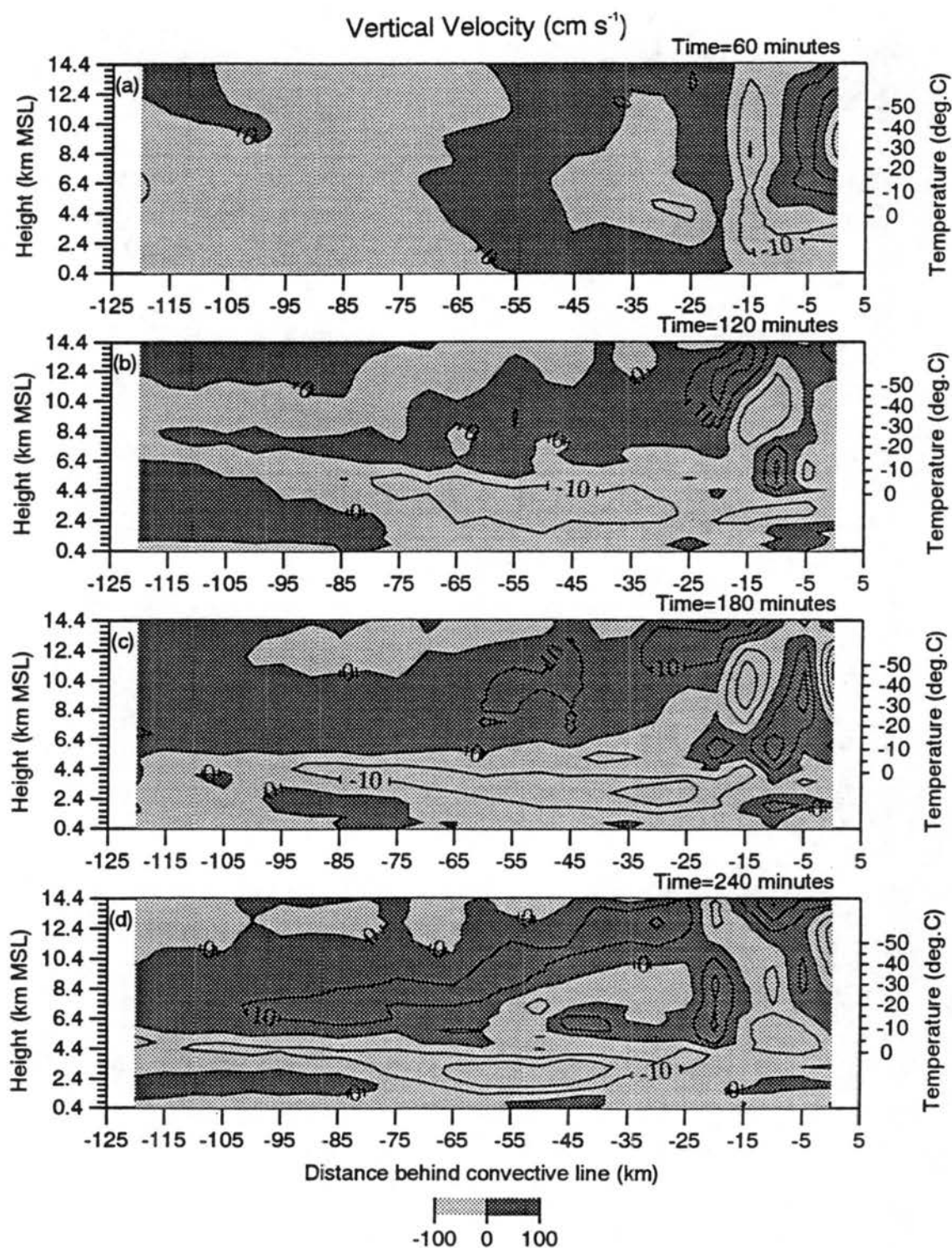


Fig. 7.2. Vertical velocities (cm s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation. Light shading denotes downward motion while dark shading denotes upward motion. Contours are every 10 cm s^{-1} .

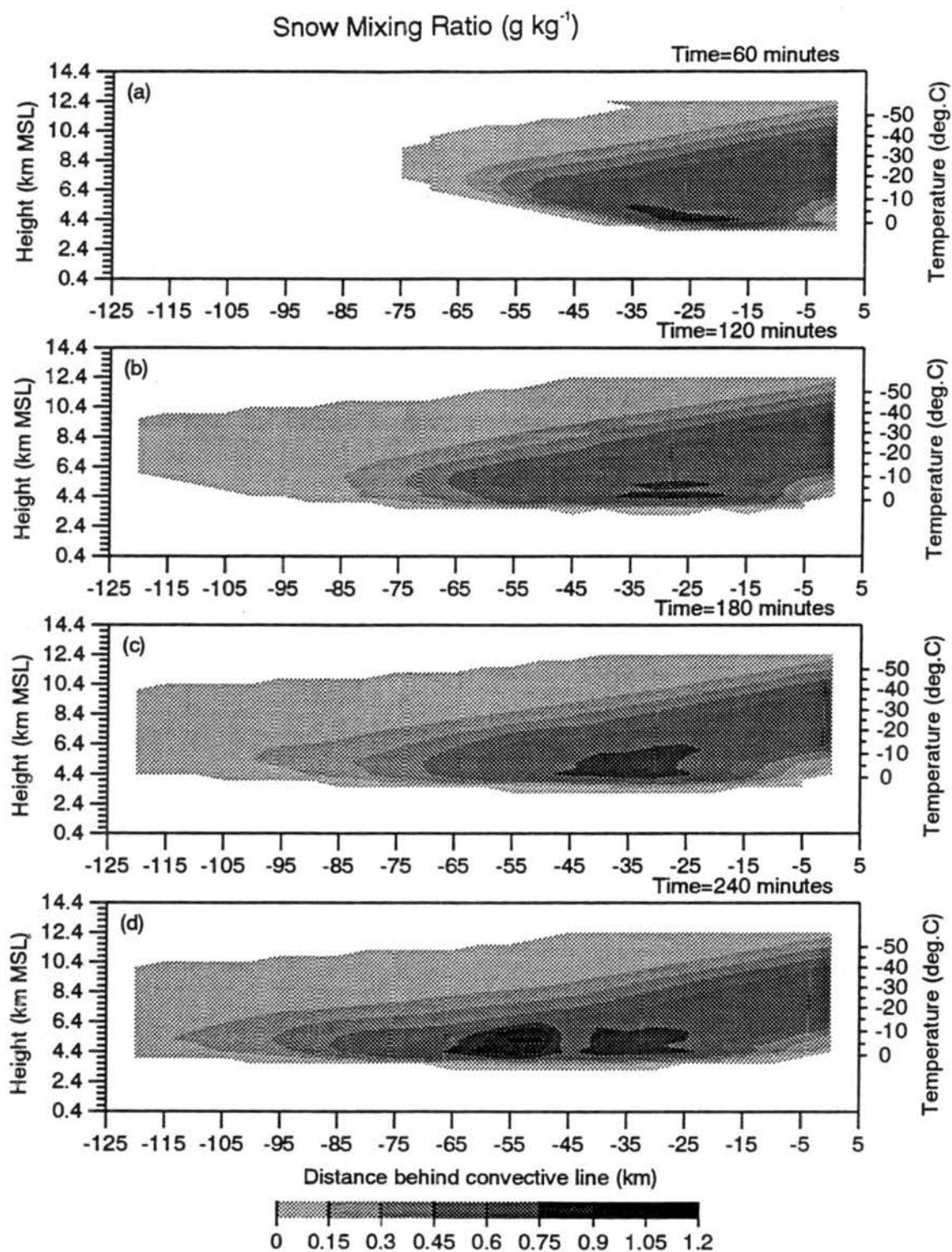


Fig. 7.3. Snow mixing ratios (g kg^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation. Shading denotes mixing ratio at 0.15 g kg^{-1} intervals (key at bottom).

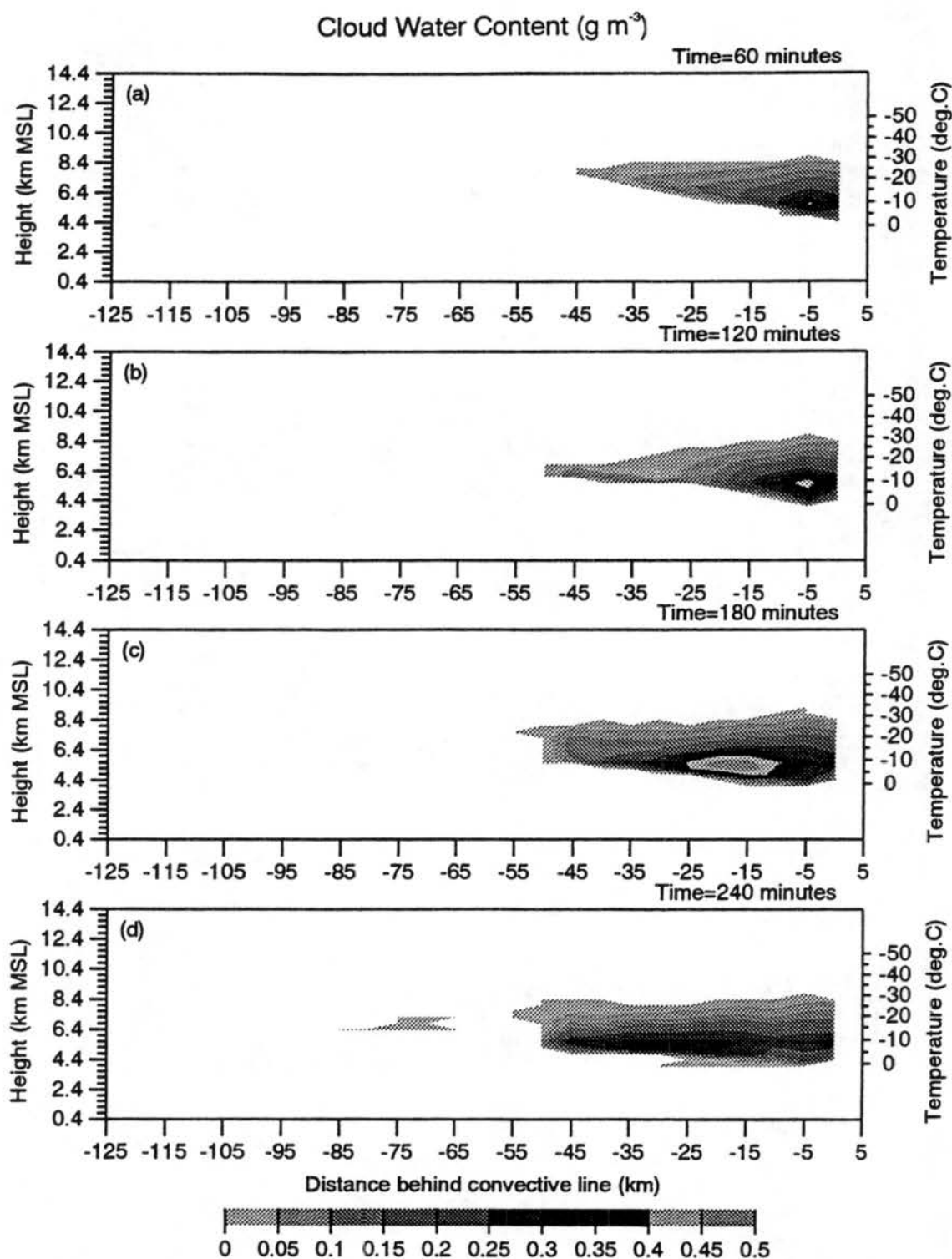


Fig. 7.4. Cloud water liquid water content (g m^{-3}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation. Shading denotes mixing ratio at 0.05 g kg^{-1} intervals (key at bottom).

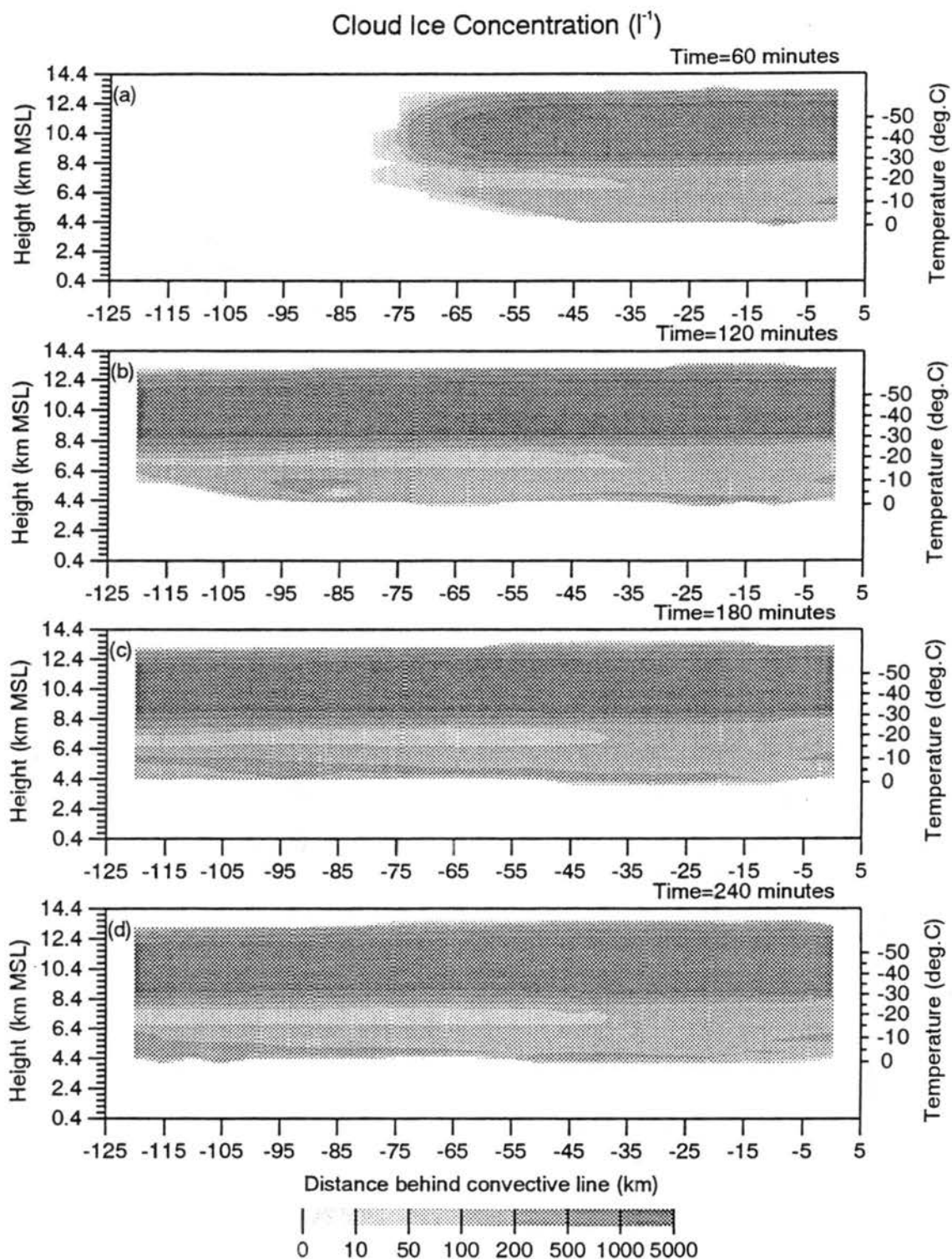


Fig. 7.5. Cloud ice concentration (l^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation. Shading denotes ice concentrations at variable intervals (key at bottom).

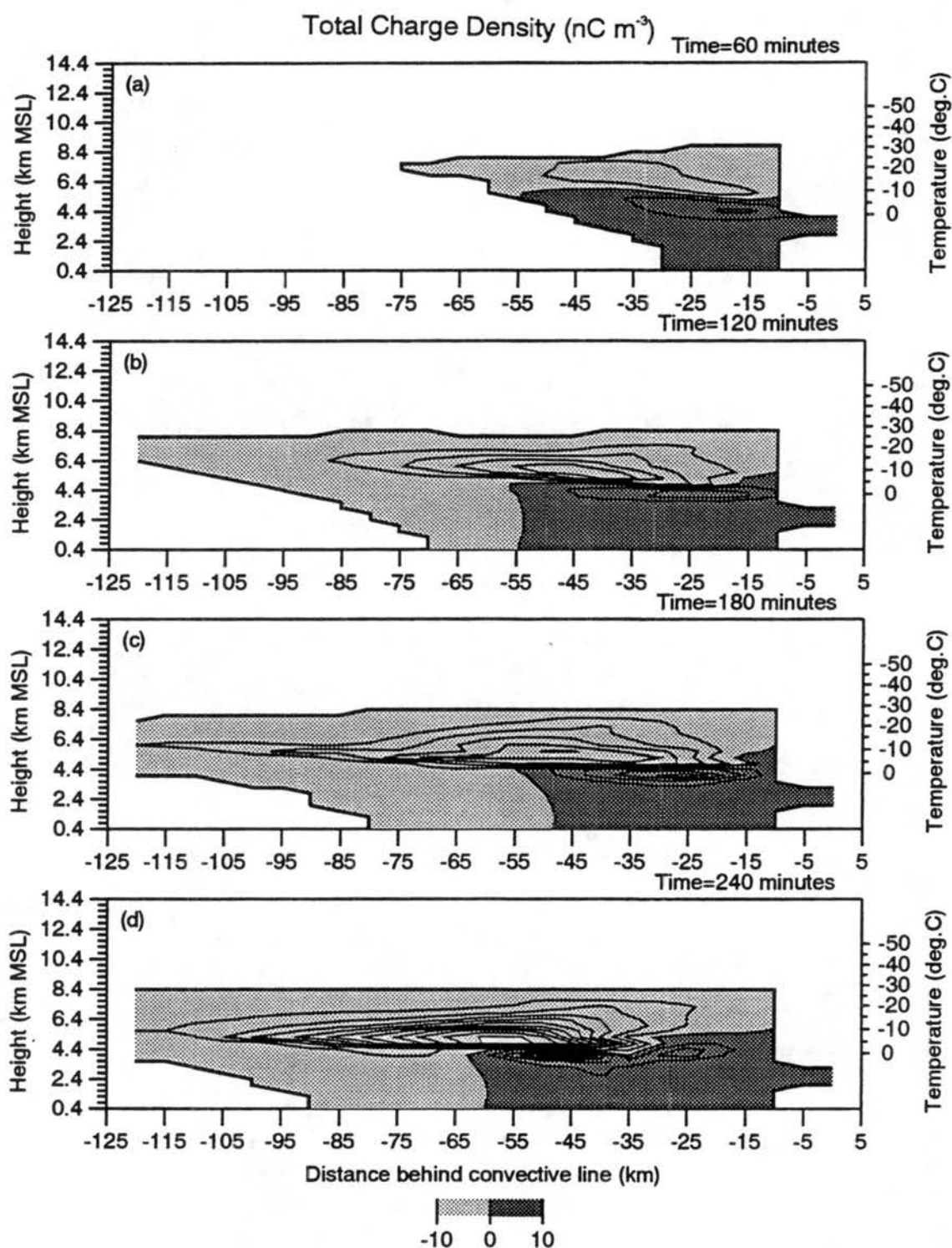


Fig. 7.6. Total charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

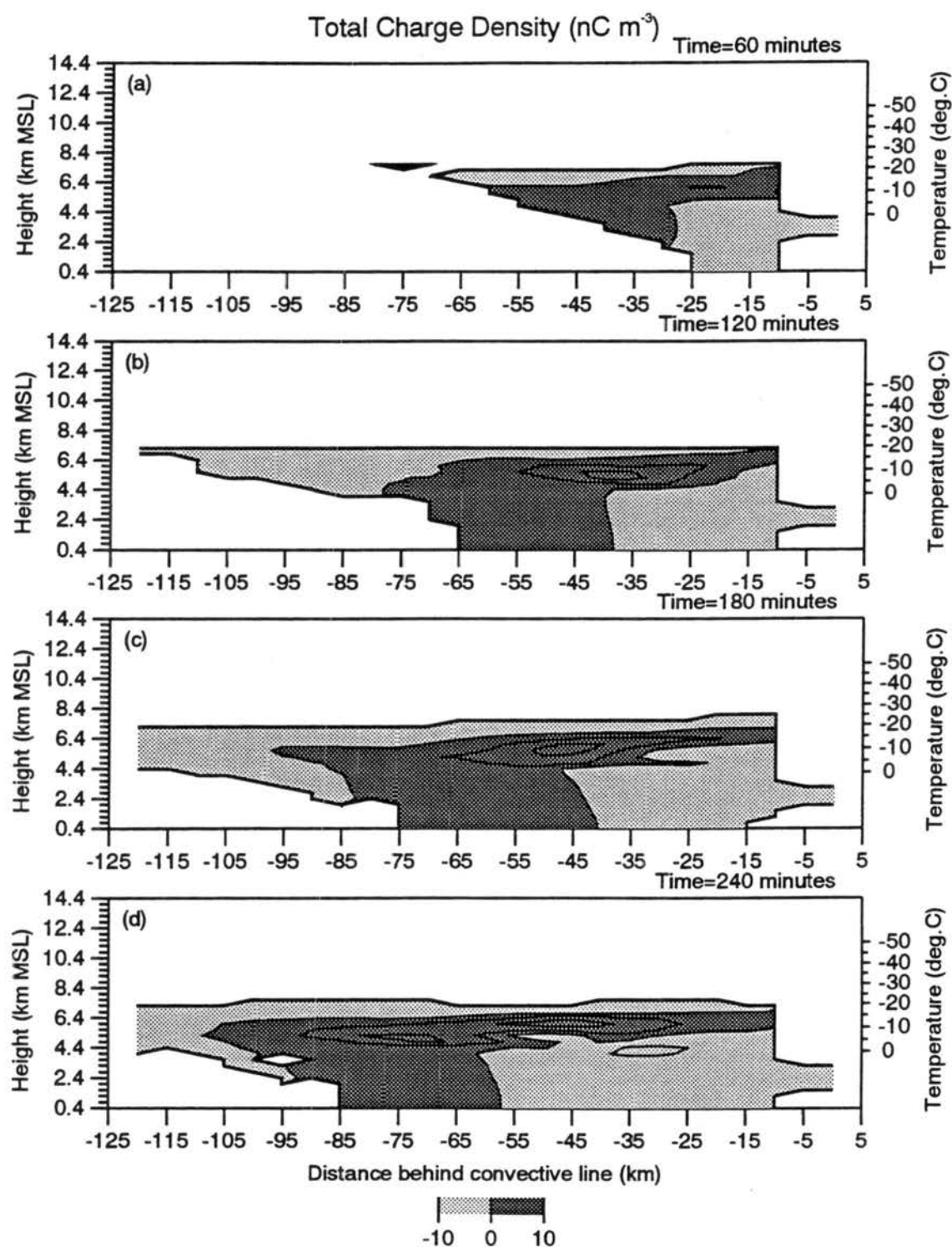


Fig. 7.7. Total charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no pulse (NP) model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

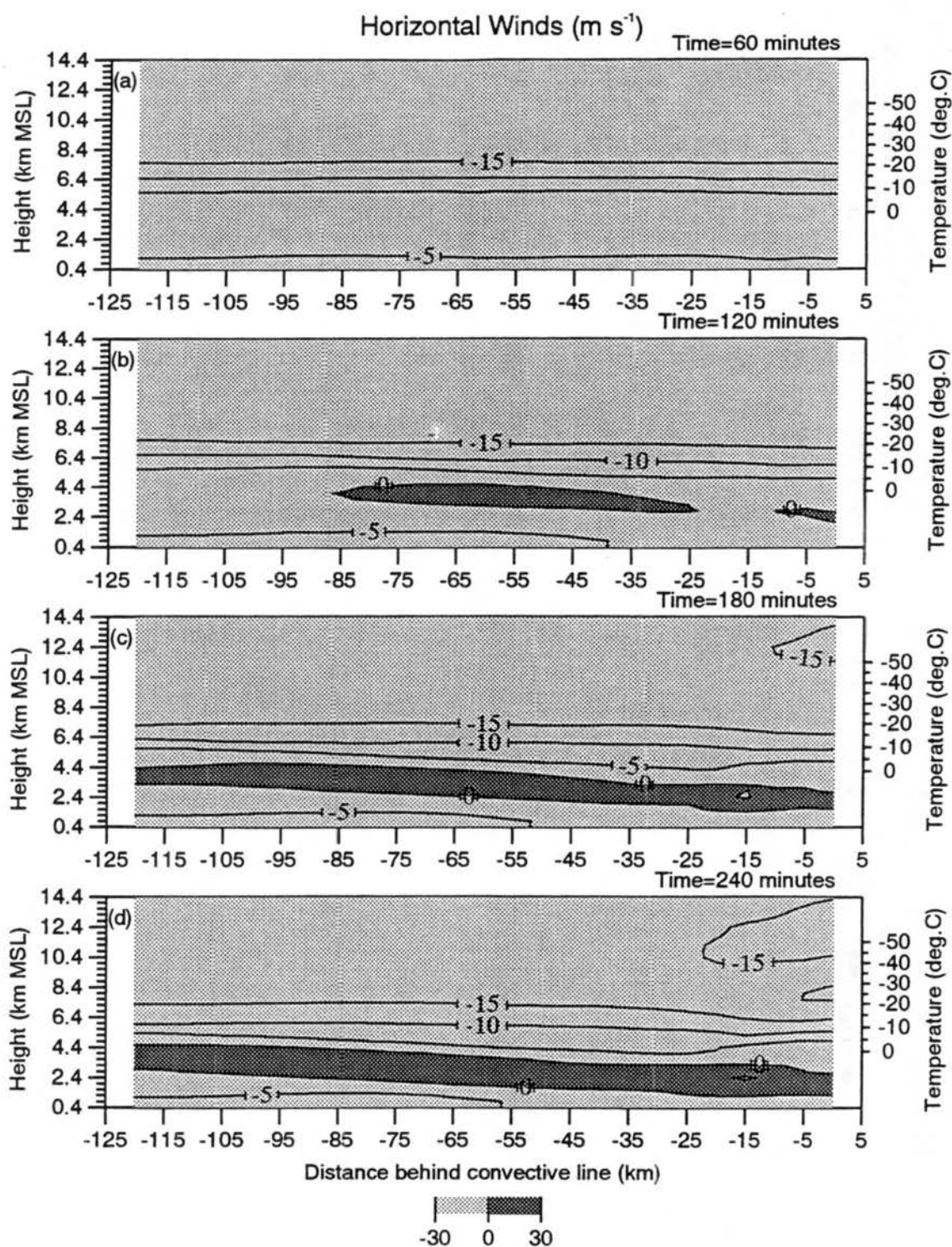


Fig. 7.8. Horizontal winds (m s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation. Light shading denotes front-to-rear flow while dark shading denotes rear-to-front flow. Contours are every 5 m s^{-1} .

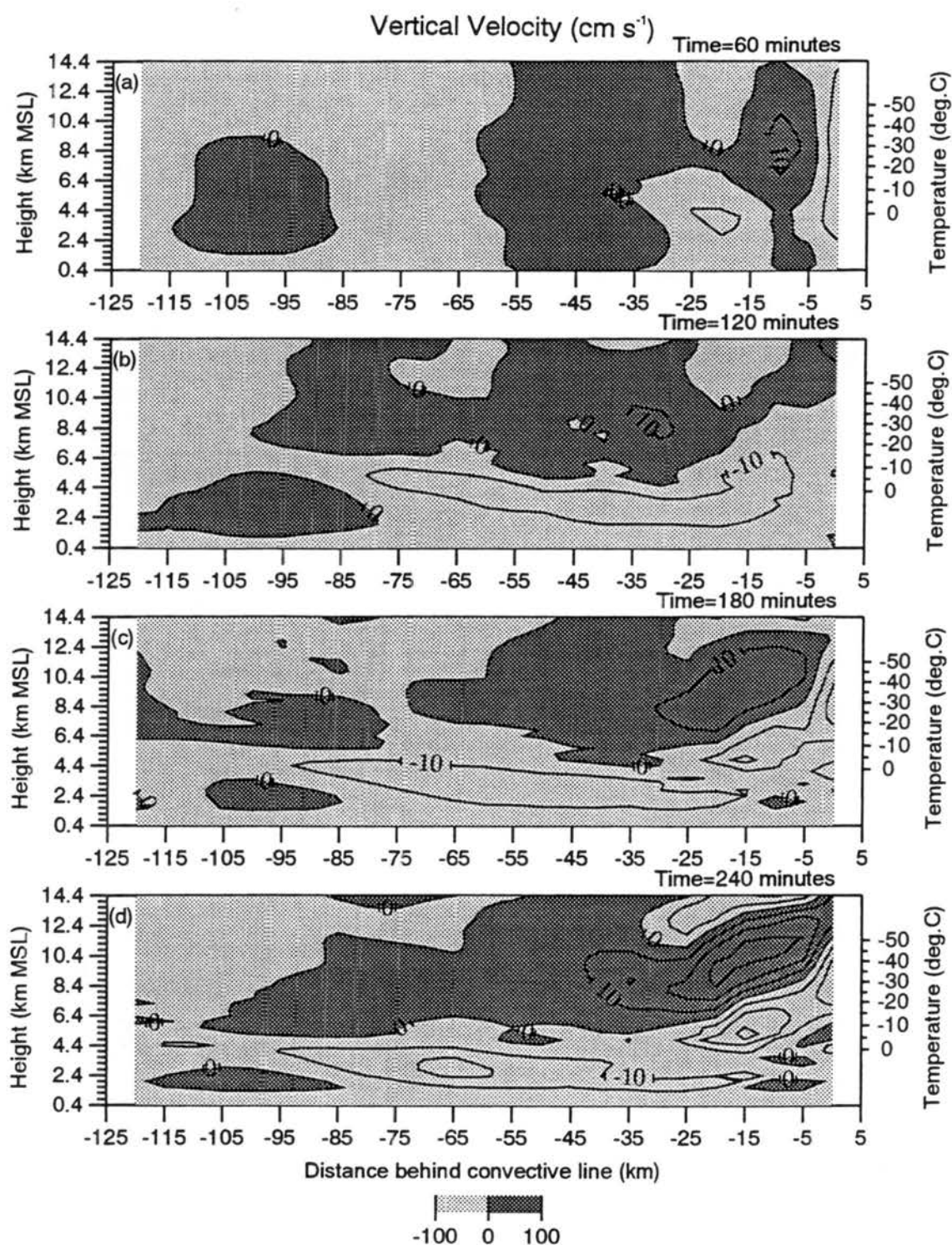


Fig. 7.9. Vertical velocities (cm s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation. Light shading denotes downward motion while dark shading denotes upward motion. Contours are every 10 cm s^{-1} .

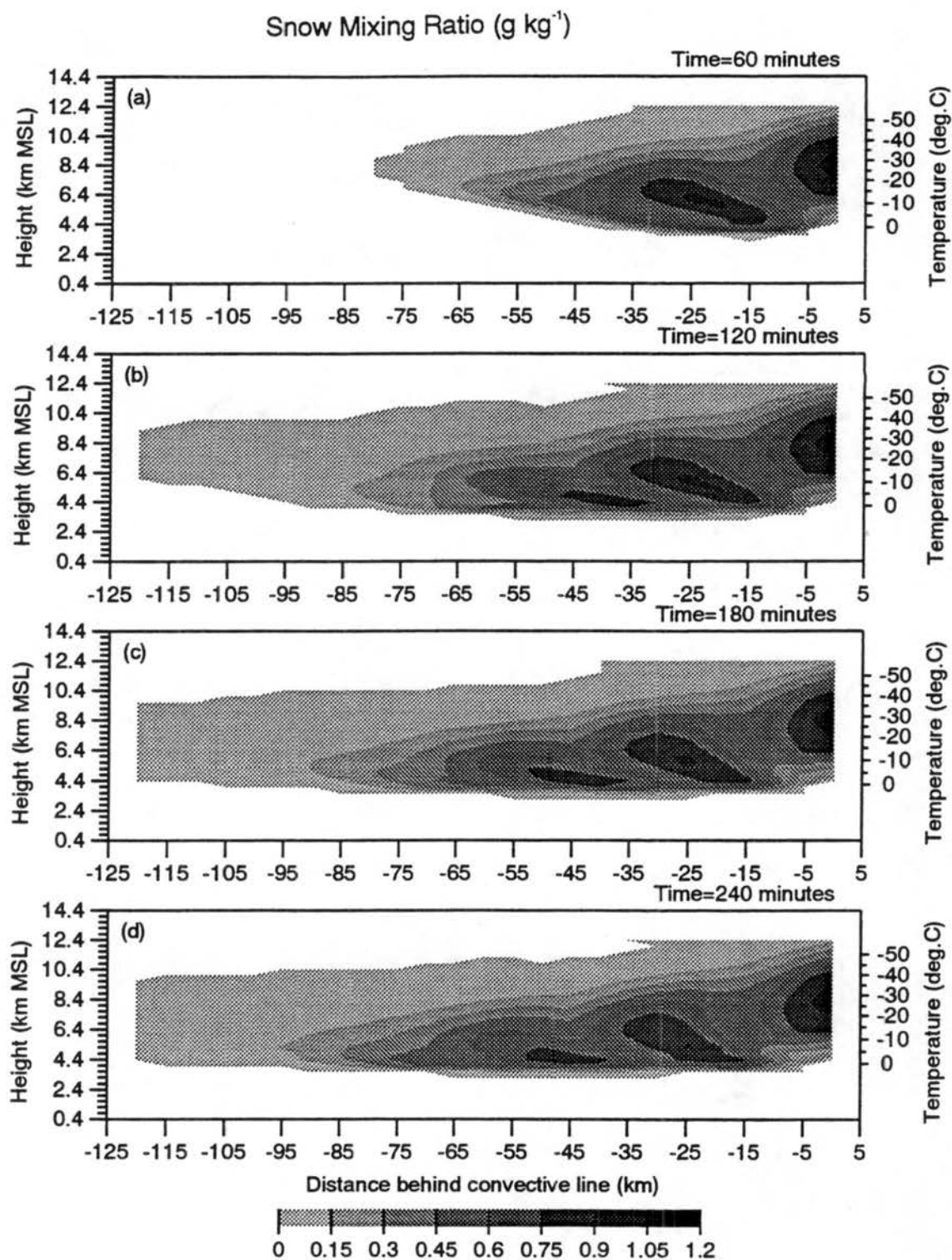


Fig. 7.10. Snow mixing ratios (g kg^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation. Shading denotes mixing ratio at 0.15 g kg^{-1} intervals (key at bottom).

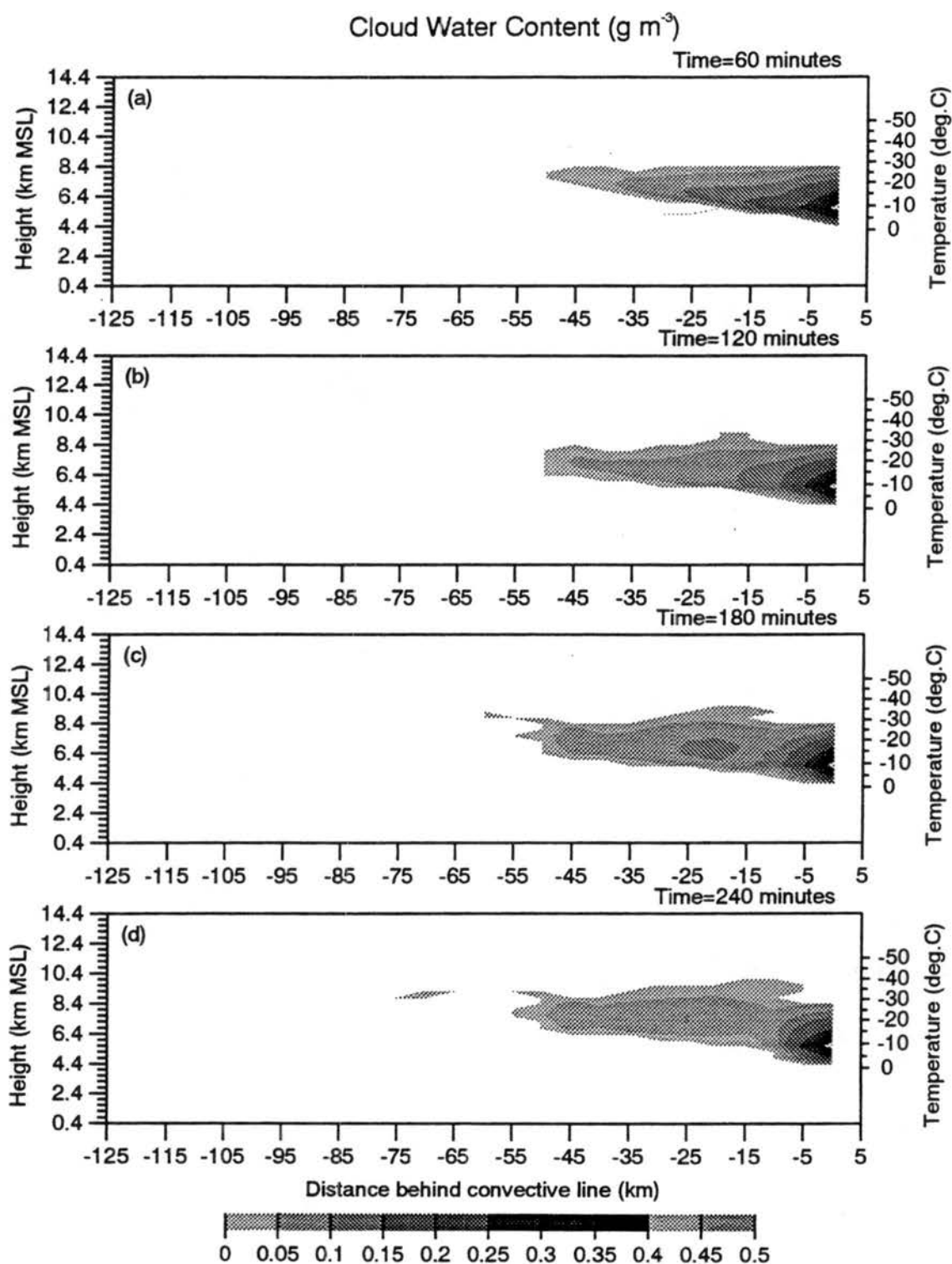


Fig. 7.11. Cloud water liquid water content (g m^{-3}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation. Shading denotes mixing ratio at 0.05 g kg^{-1} intervals (key at bottom).

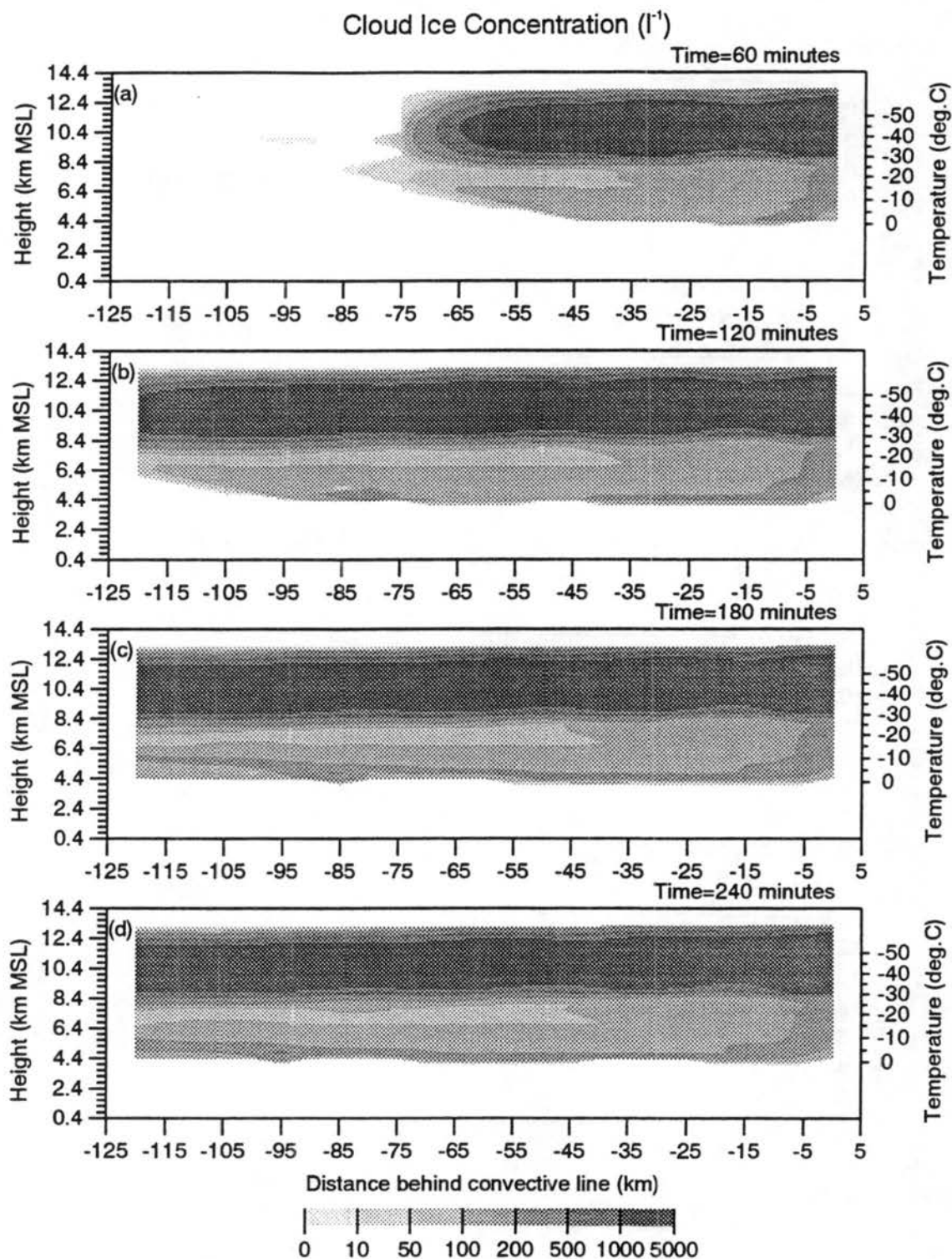


Fig. 7.12. Cloud ice concentration (l^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation. Shading denotes ice concentrations at variable intervals (key at bottom).

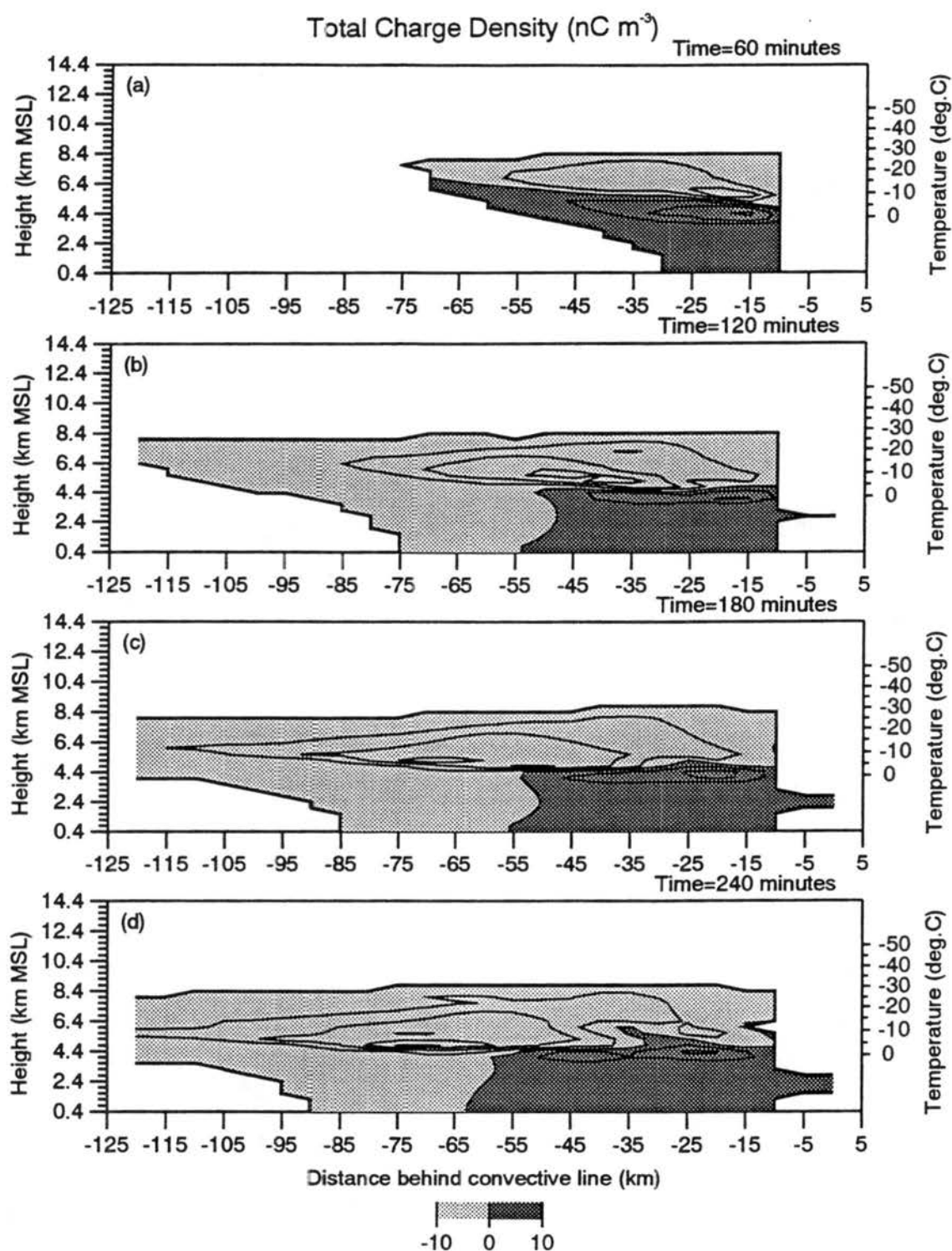


Fig. 7.13. Total charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

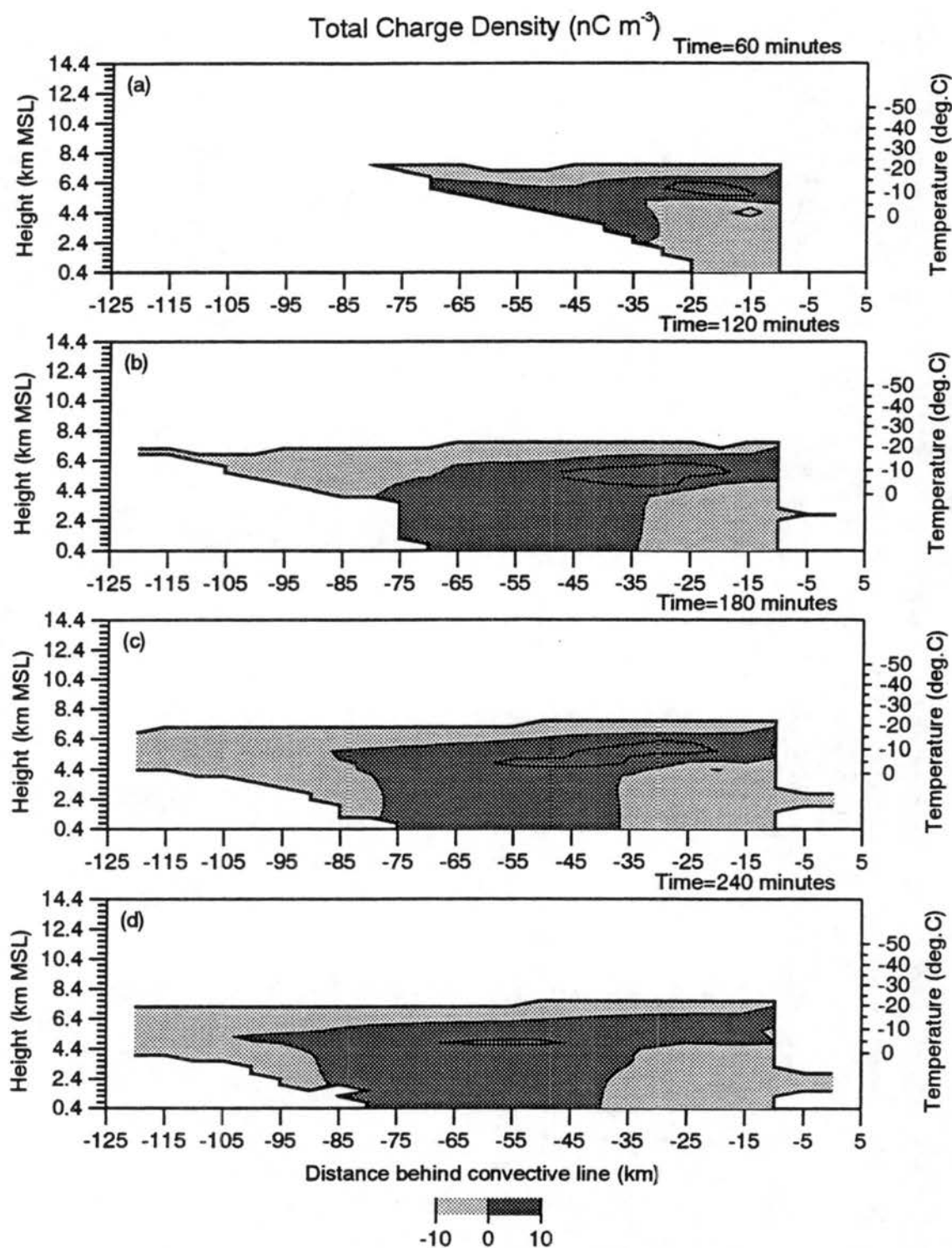


Fig. 7.14. Total charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the no heating (NH) model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

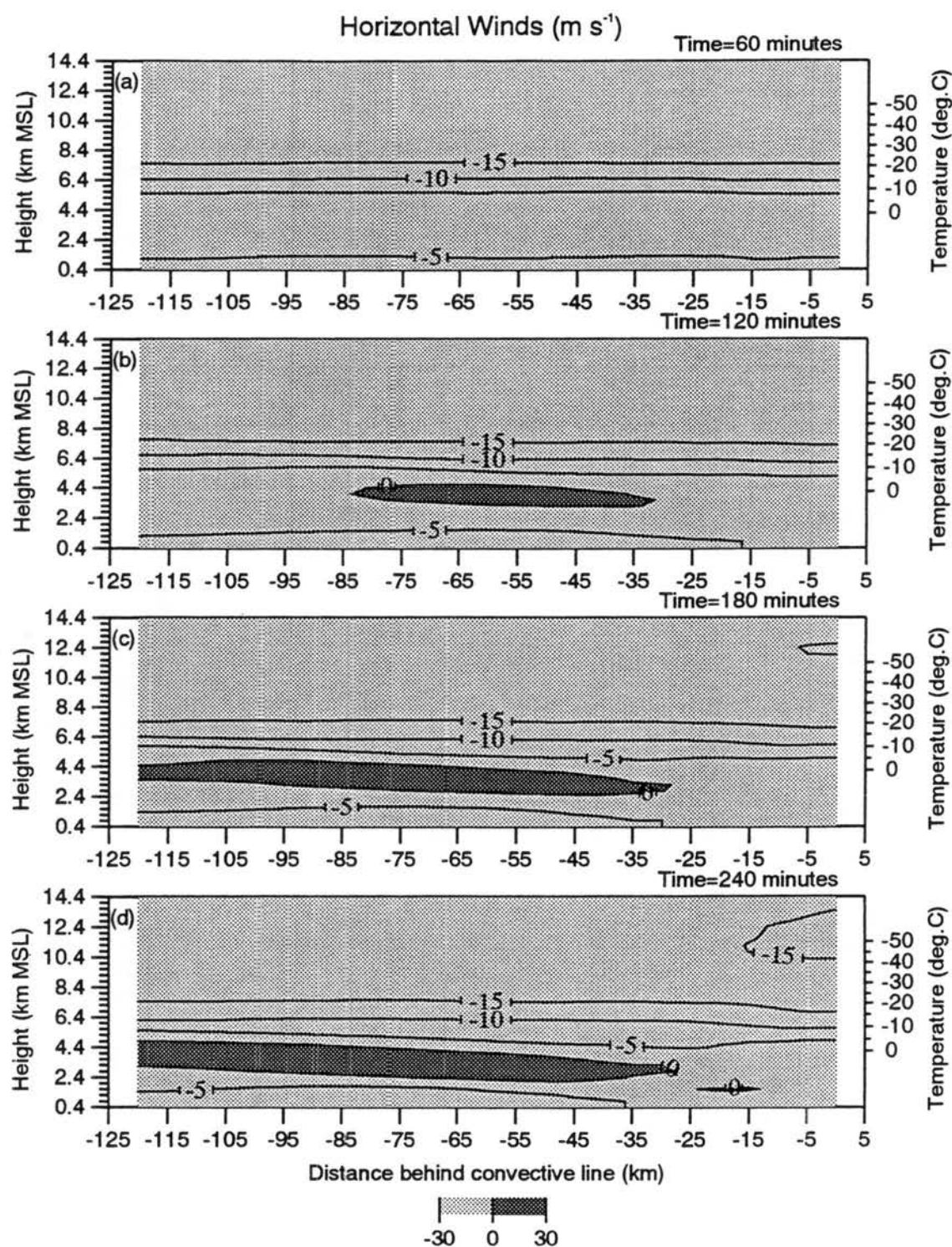


Fig. 7.15. Horizontal winds (m s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Peak convective heating used is $40^{\circ}\text{C h}^{-1}$. Light shading denotes front-to-rear flow while dark shading denotes rear-to-front flow. Contours are every 5 m s^{-1} .

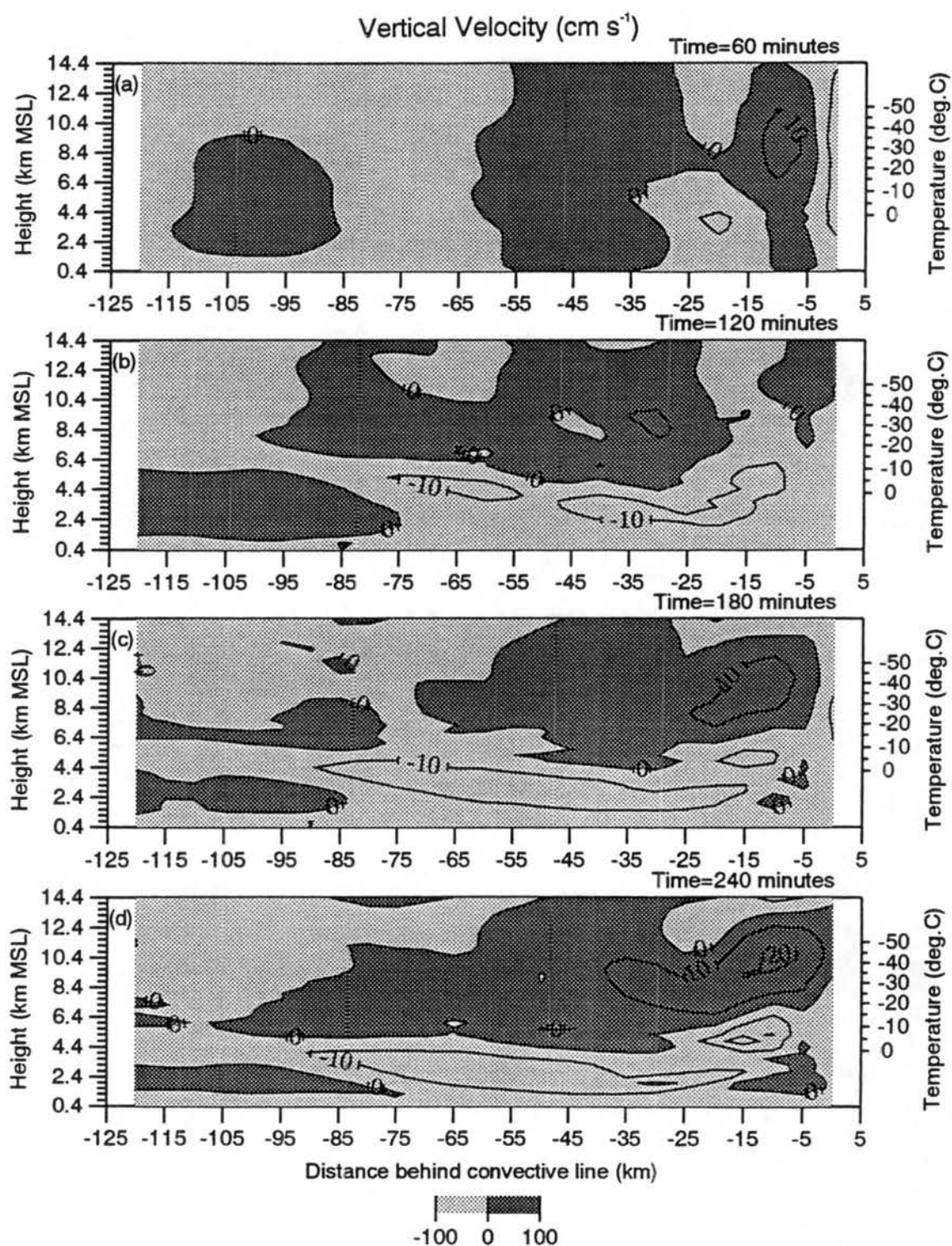


Fig. 7.16. Vertical velocities (cm s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Peak convective heating used is $40^{\circ}\text{C h}^{-1}$. Light shading denotes downward motion while dark shading denotes upward motion. Contours are every 10 cm s^{-1} .

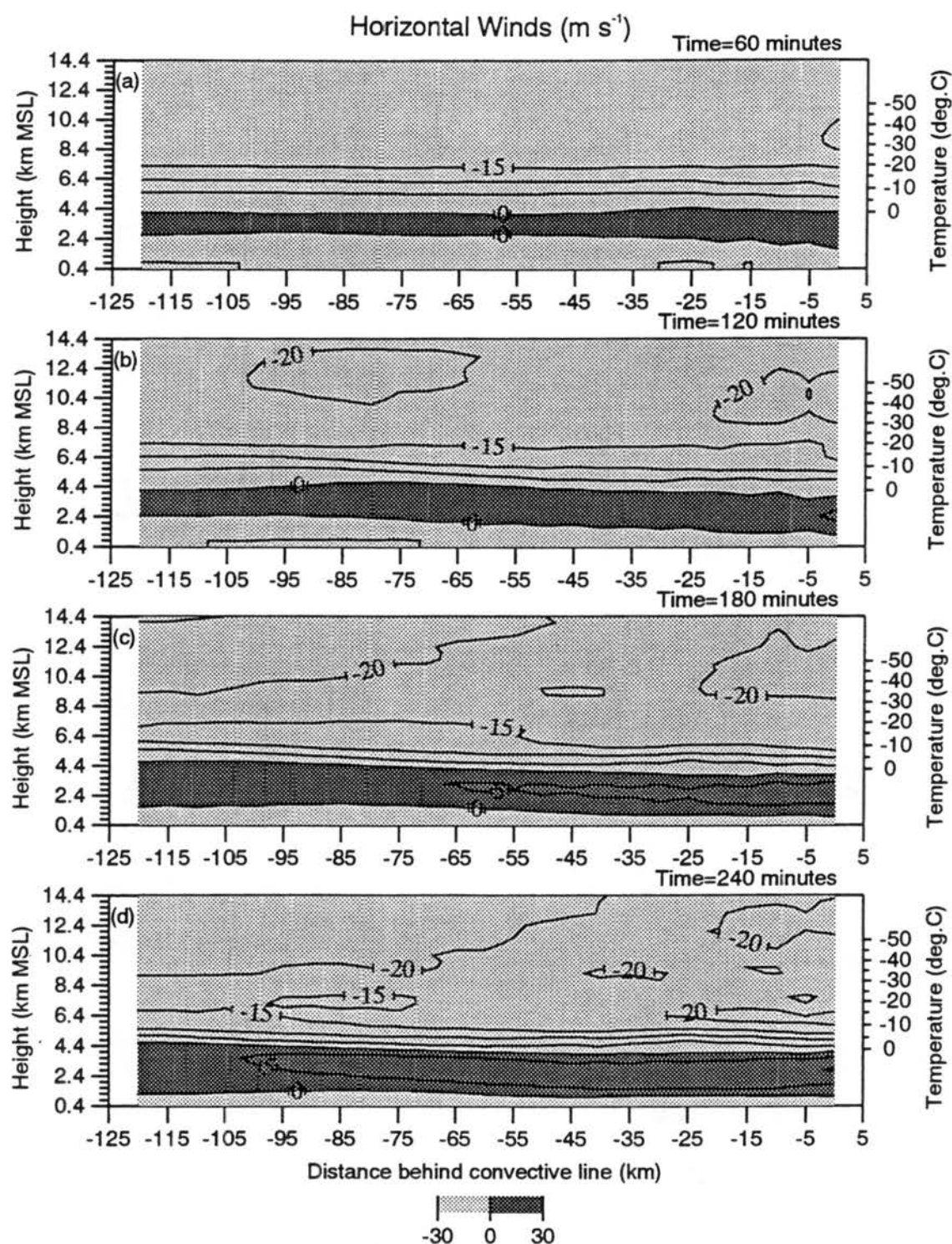


Fig. 7.17. Horizontal winds (m s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies. Light shading denotes front-to-rear flow while dark shading denotes rear-to-front flow. Contours are every 5 m s^{-1} .

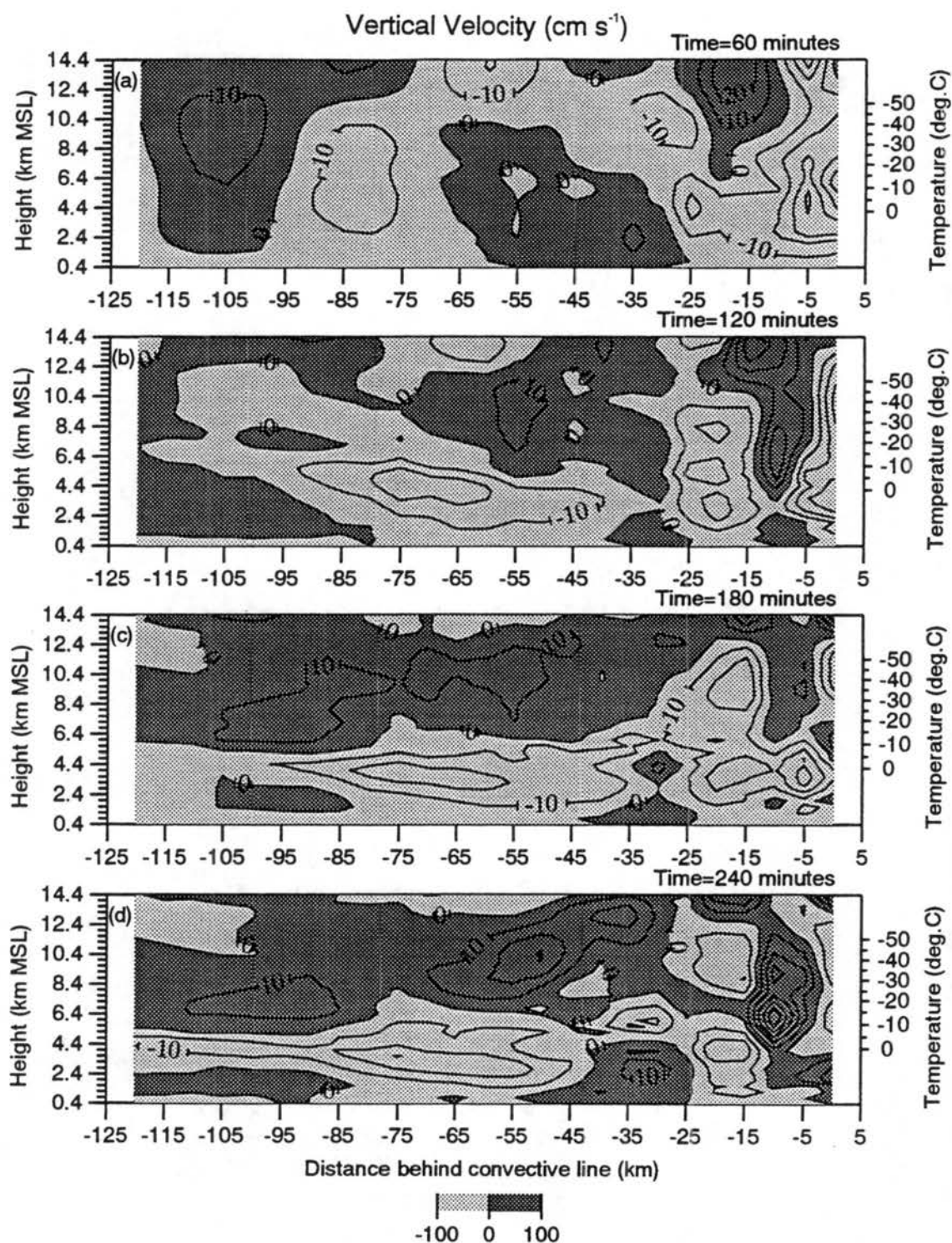


Fig. 7.18. Vertical velocities (cm s^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies. Light shading denotes downward motion while dark shading denotes upward motion. Contours are every 10 cm s^{-1} .

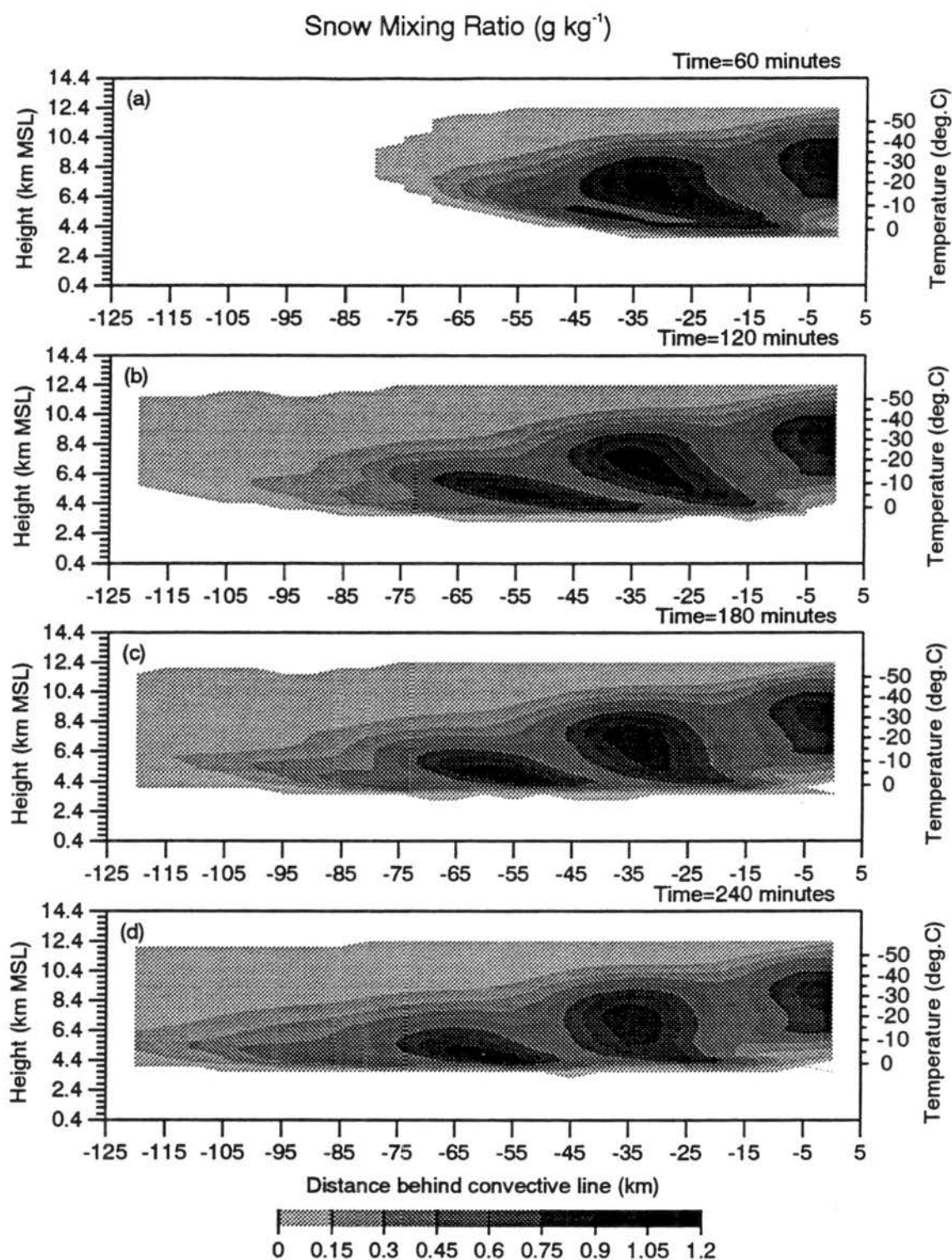


Fig. 7.19. Snow mixing ratios (g kg^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies. Shading denotes mixing ratio at 0.15 g kg^{-1} intervals (key at bottom).

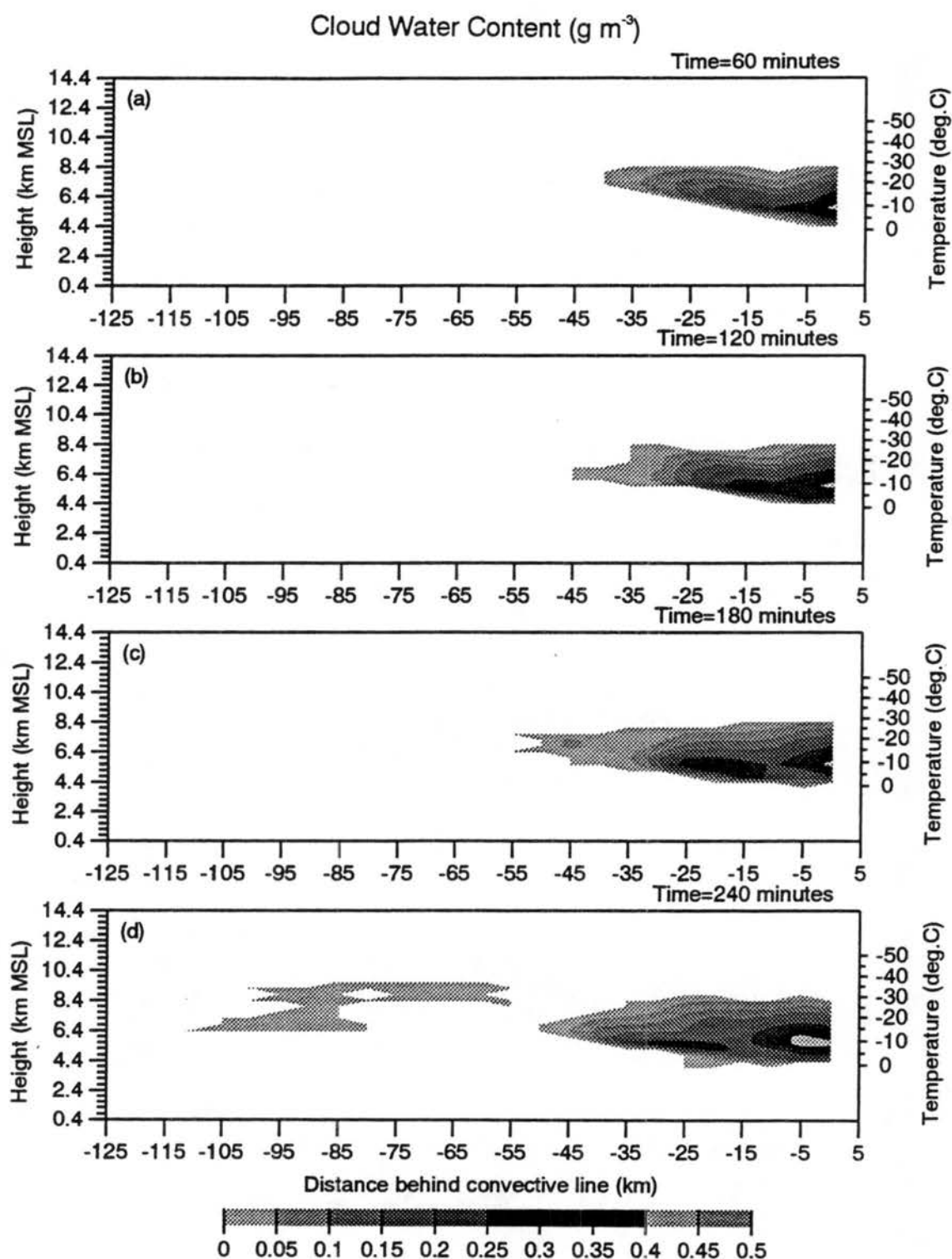


Fig. 7.20. Cloud water liquid water content (g m^{-3}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies. Shading denotes mixing ratio at 0.05 g kg^{-1} intervals (key at bottom).

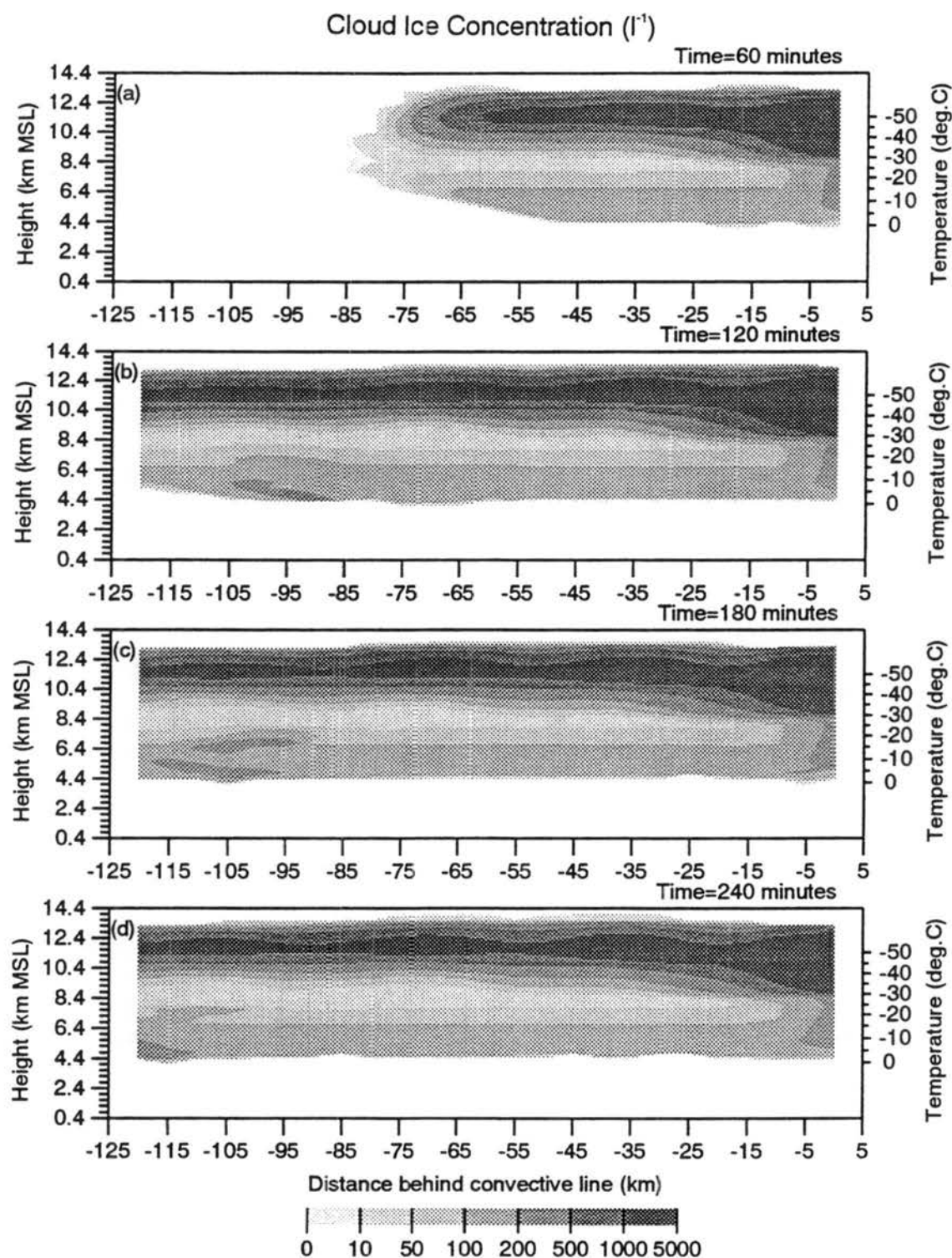


Fig. 7.21. Cloud ice concentration (l^{-1}) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies. Shading denotes ice concentrations at variable intervals (key at bottom).

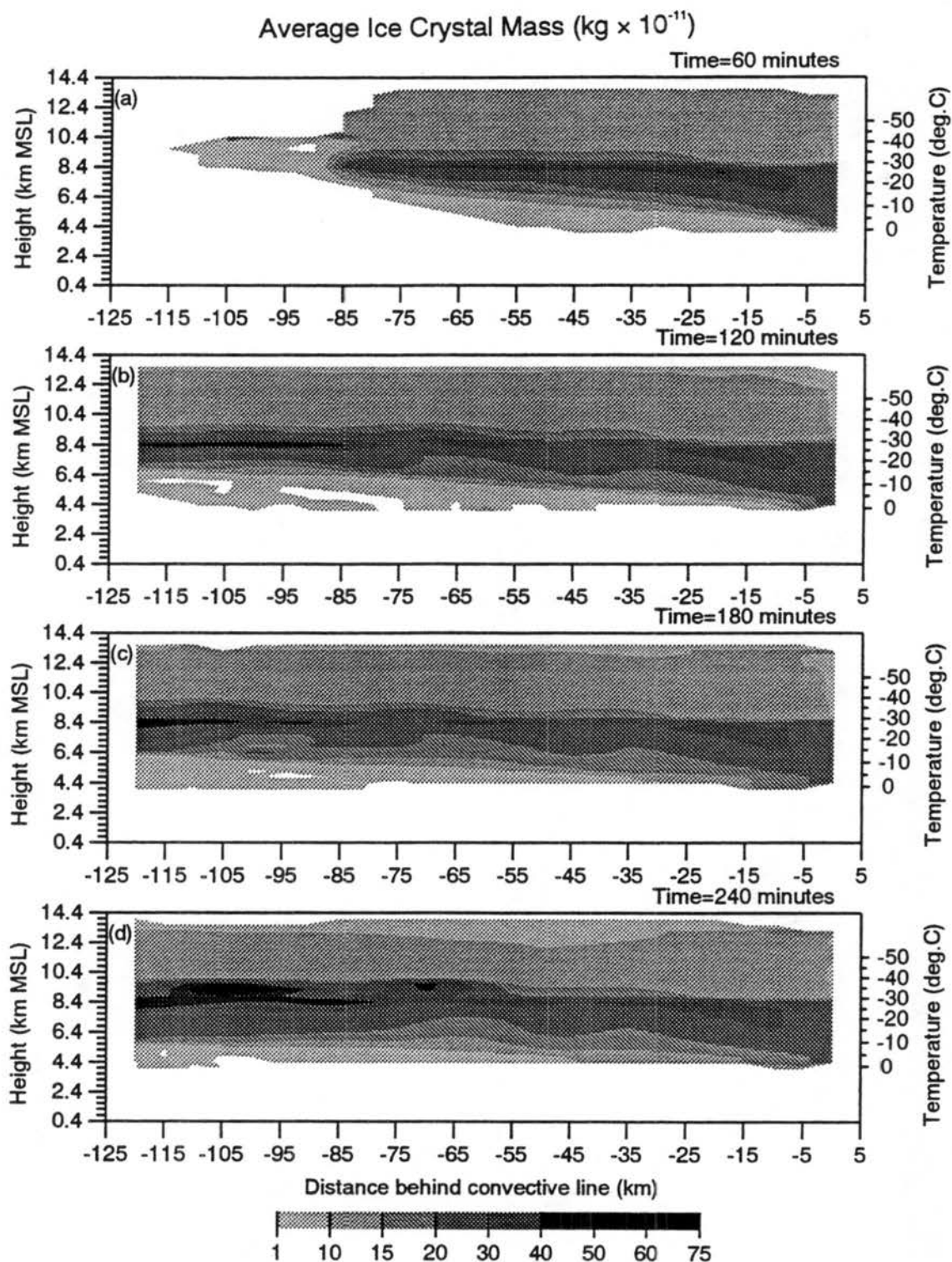


Fig. 7.22. Average ice mass ($\text{kg} \times 10^{-11}$) at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies. Shading denotes ice mass at variable intervals (key at bottom).

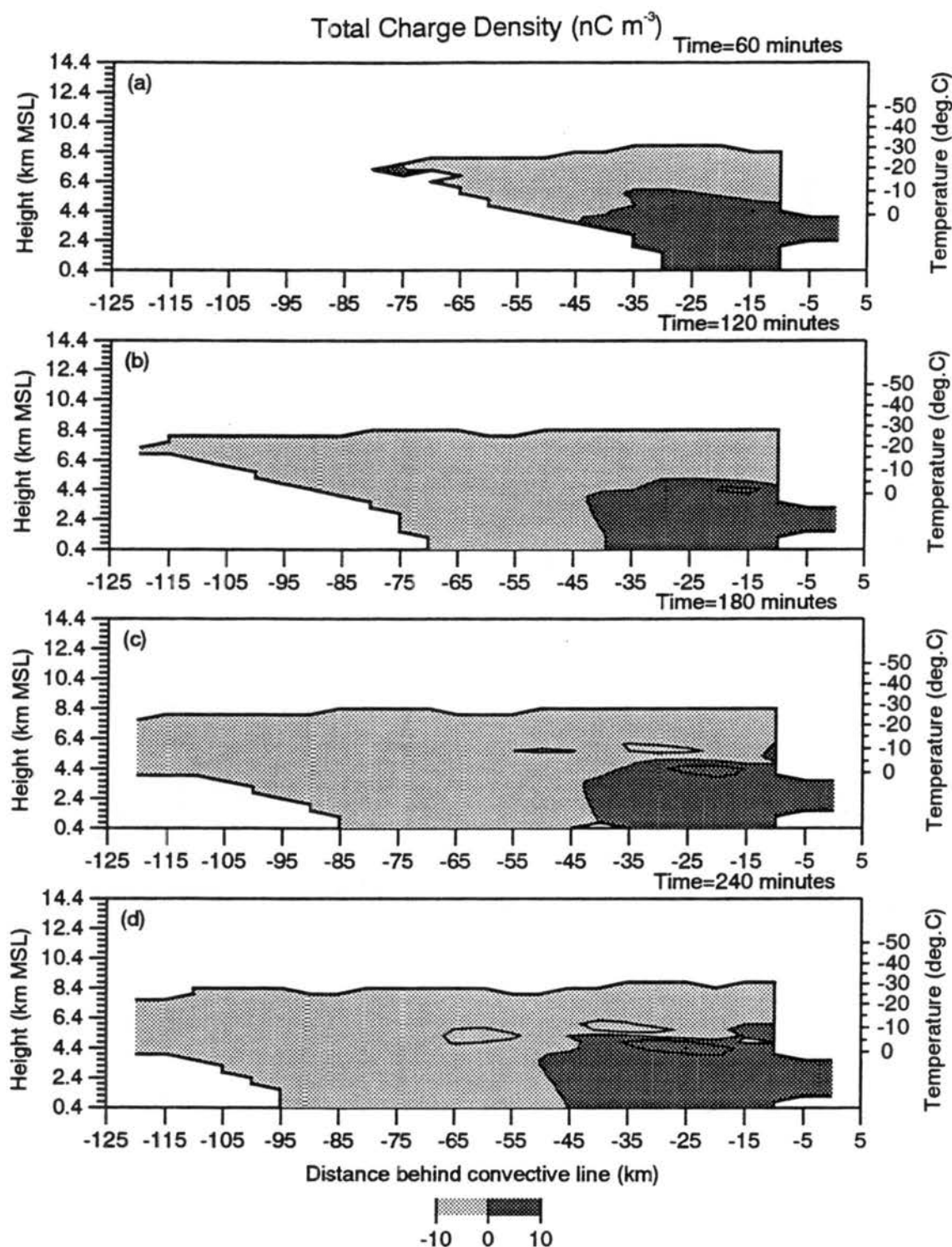


Fig. 7.23. Total charge density magnitudes (nC m^{-3}) for the Takahashi (1978) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

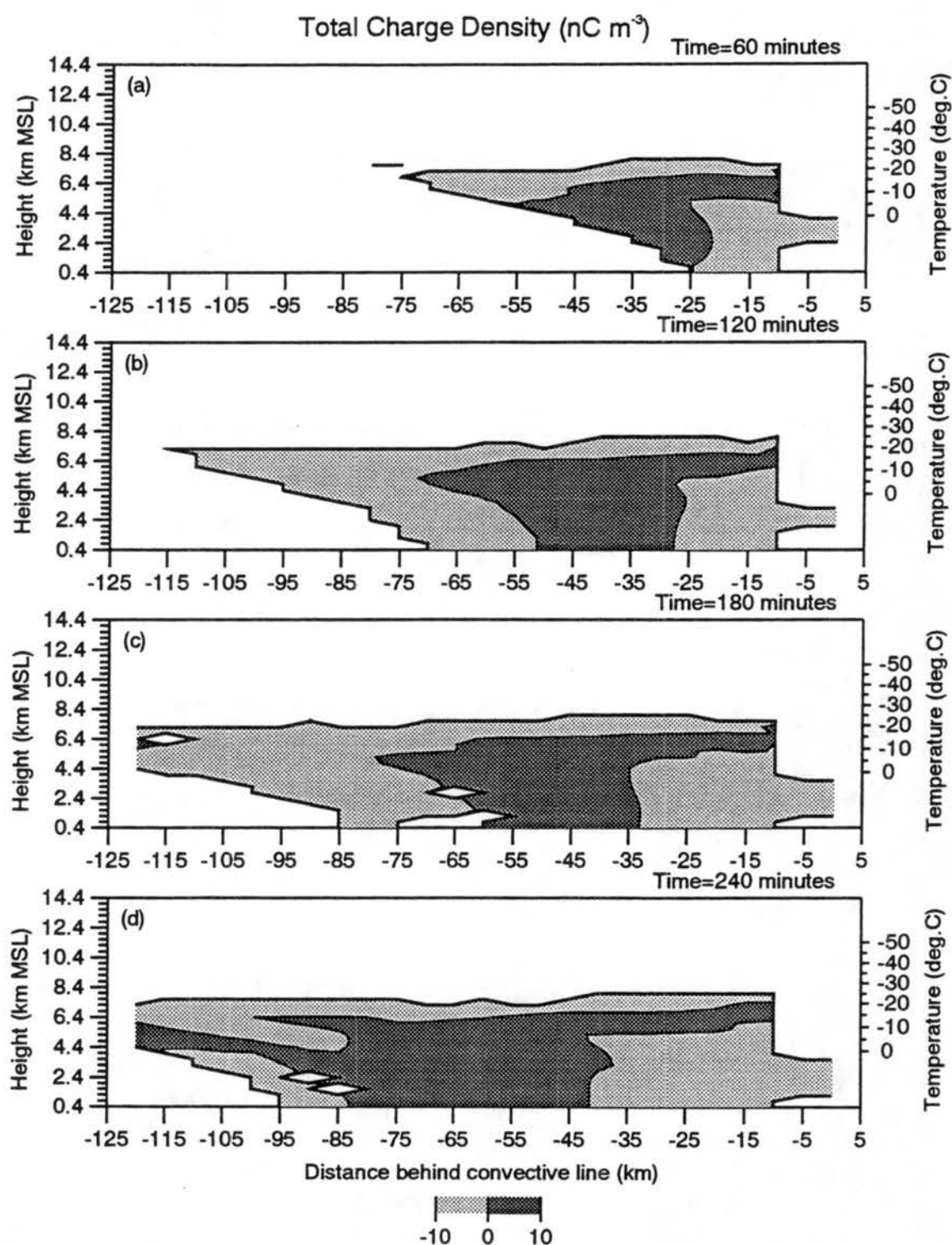


Fig. 7.24. Total charge density magnitudes (nC m^{-3}) for the Saunders et al. (1991) non-inductive charging model run at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation with increased collection efficiencies. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

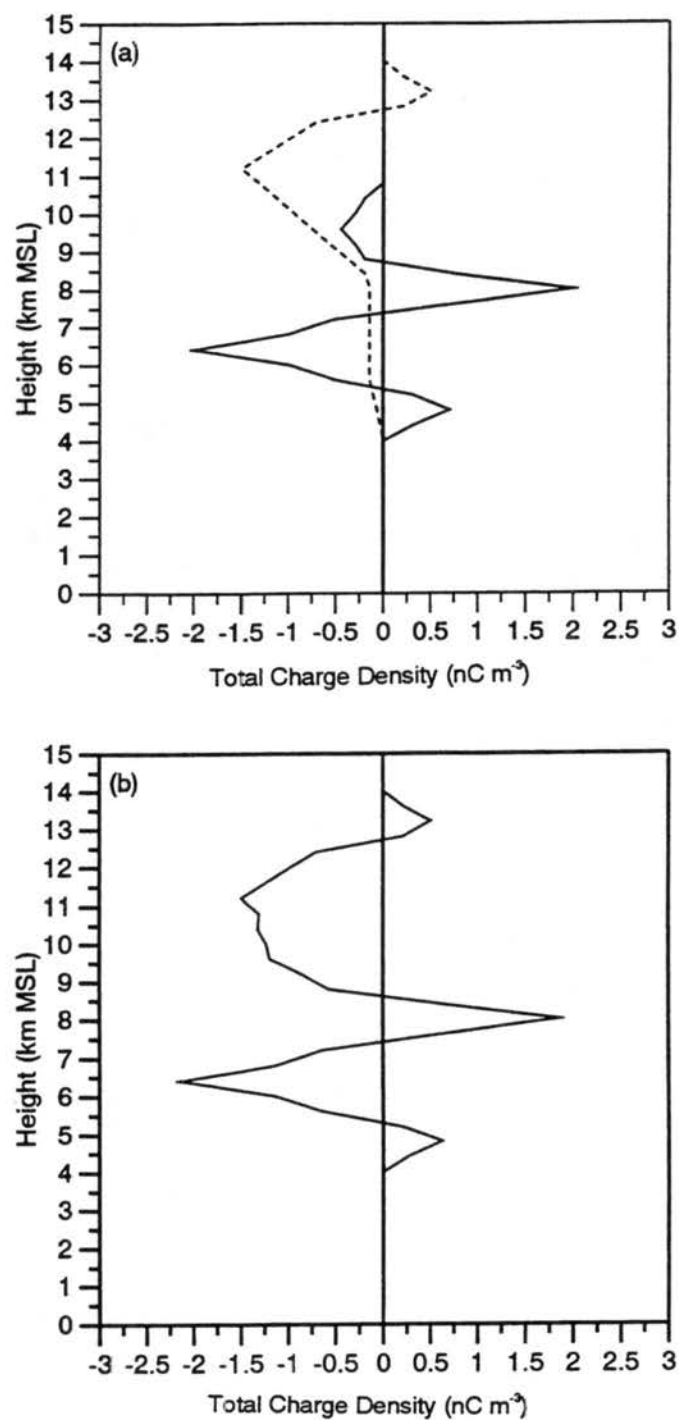


Fig. 7.25. Charge profiles used to initialize the buffer zone region for CA (a) Charge densities (nC m⁻³) of snow (solid line) and cloud ice (dashed line). (b) Total charge density (nC m⁻³).

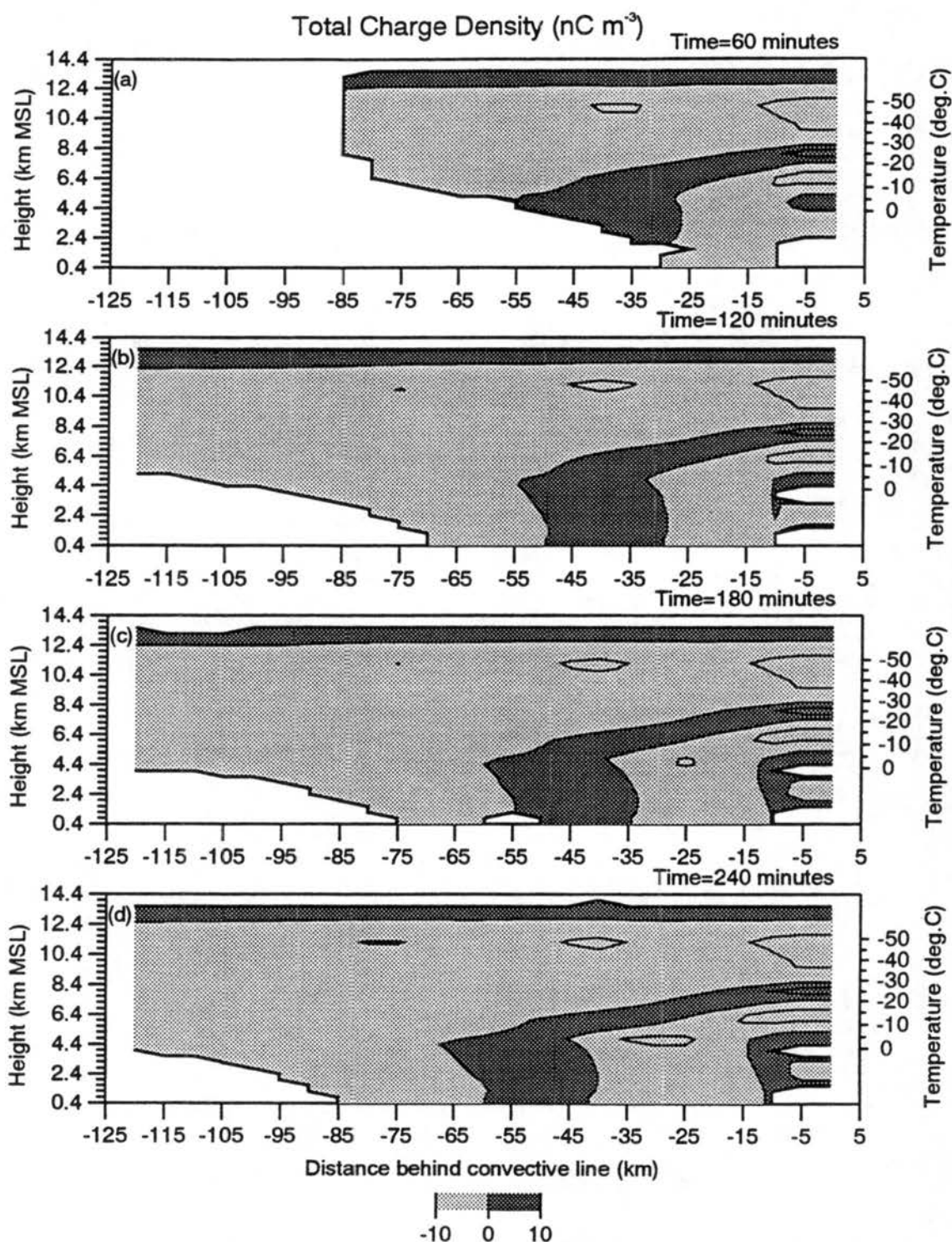


Fig. 7.26. Total charge density magnitudes (nC m^{-3}) for advection of the charge profile depicted in Fig. 7.25 at (a) 60 minutes, (b) 120 minutes, (c) 180 minutes, and (d) 240 minutes into the model simulation. Light shading denotes negative charge while dark shading denotes positive charge. Contours are every 1.0 nC m^{-3} .

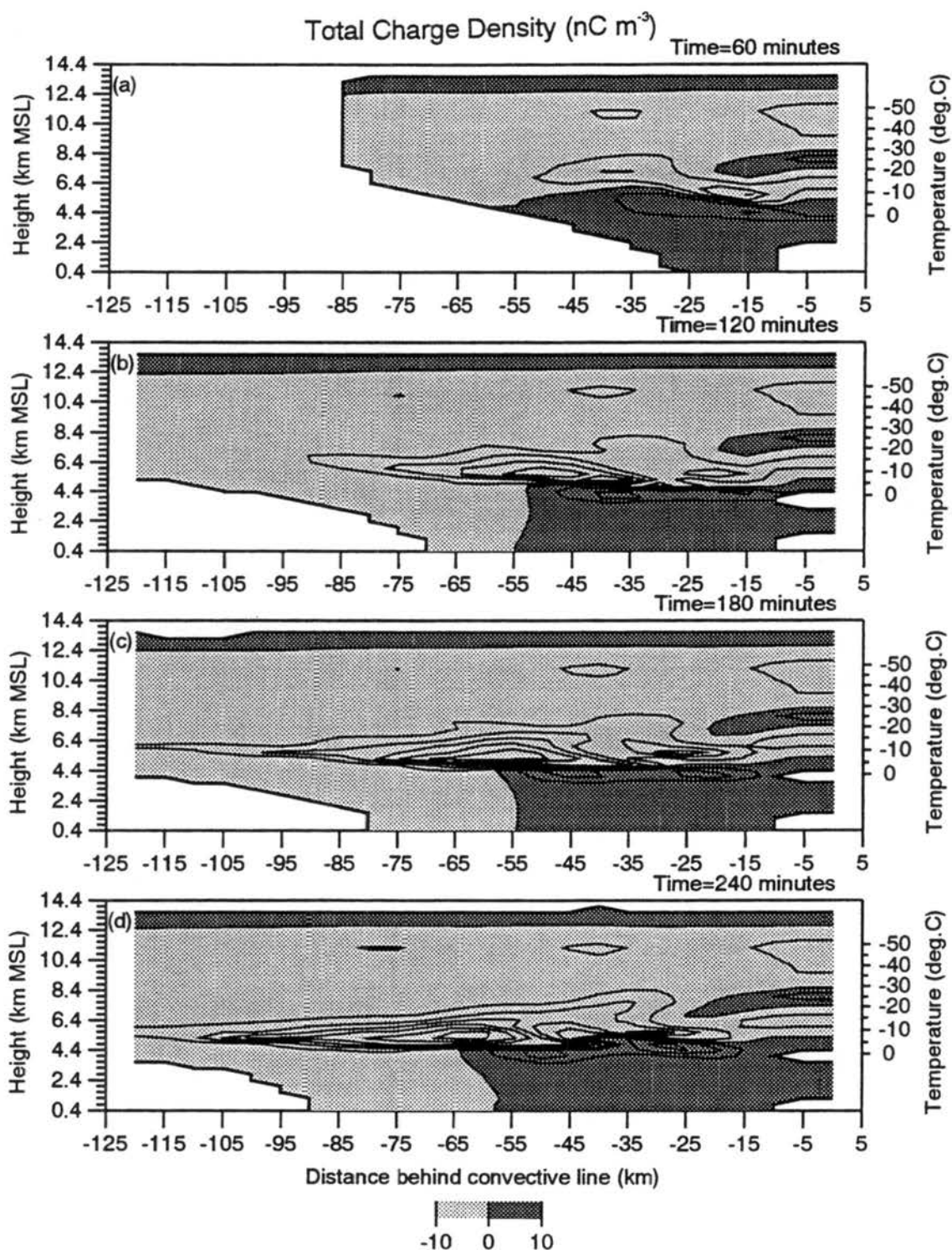


Fig. 7.27. Same as 7.26 except with Takahashi (1978) non-inductive charging included.

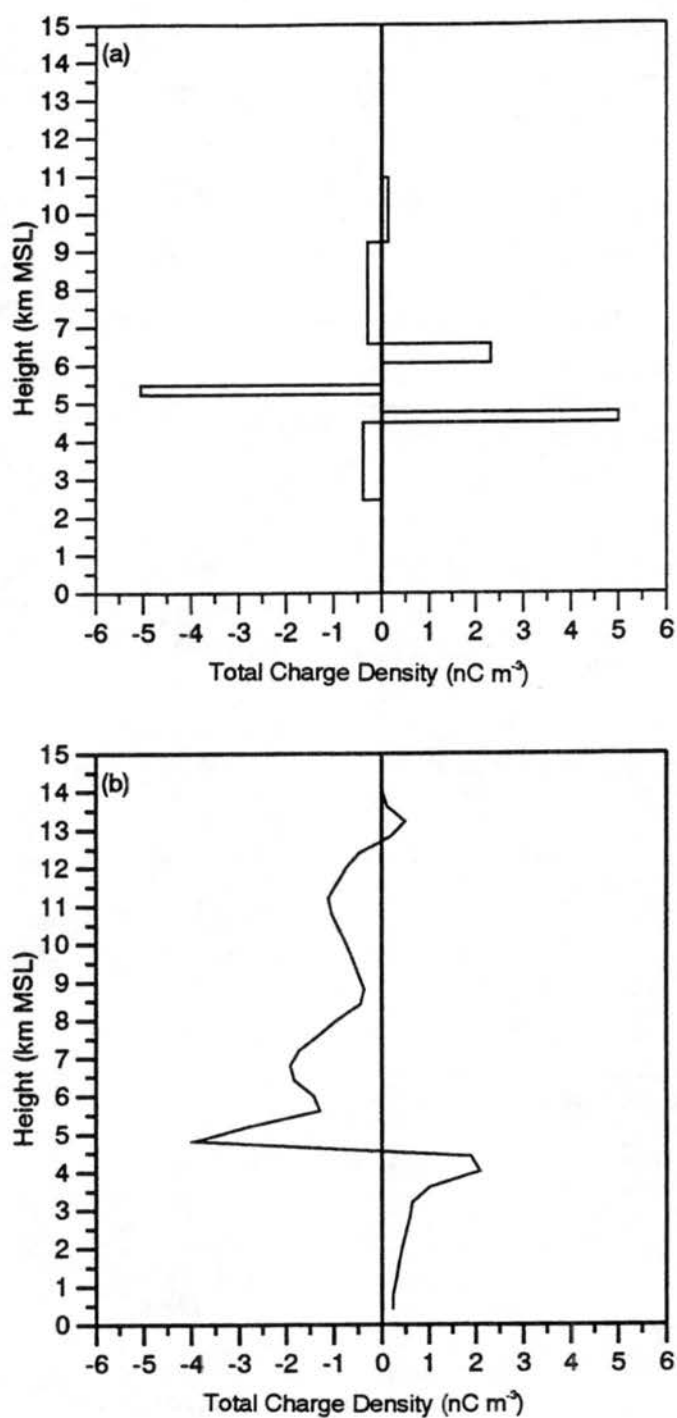


Fig. 7.28. Vertical charge density profile (nC m⁻³) (a) from symmetric MCS observations (cf. Fig. 3.12), and (b) constructed through the model grid depicted in Fig. 7.27 at 180 minutes into the simulation and a location 35 km behind the convective line. This is a location similar to where the balloon-borne EFM ascended through the symmetric MCS discussed in Chapter 3.

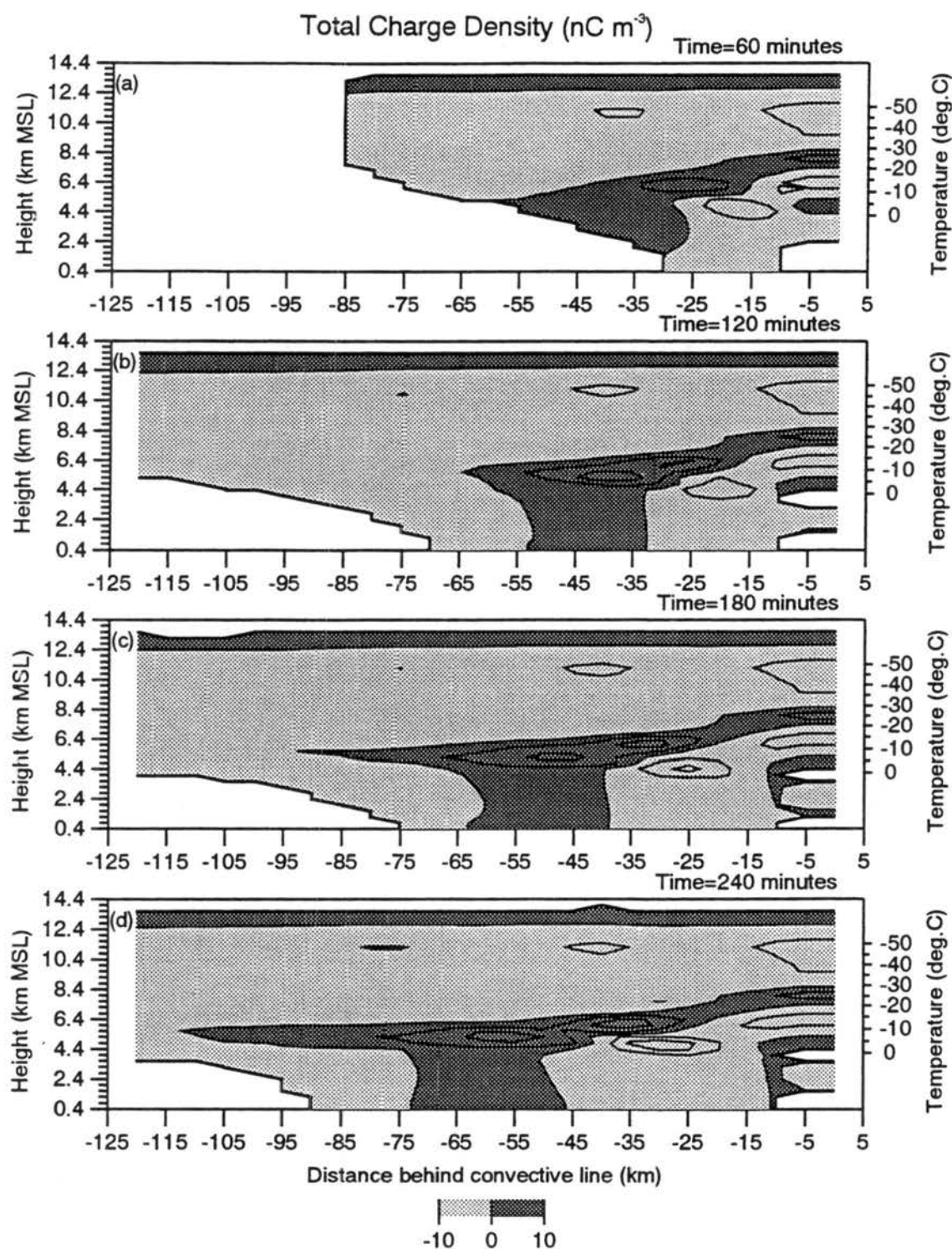


Fig. 7.29. Same as 7.26 except with Saunders et al. (1991) non-inductive charging included.

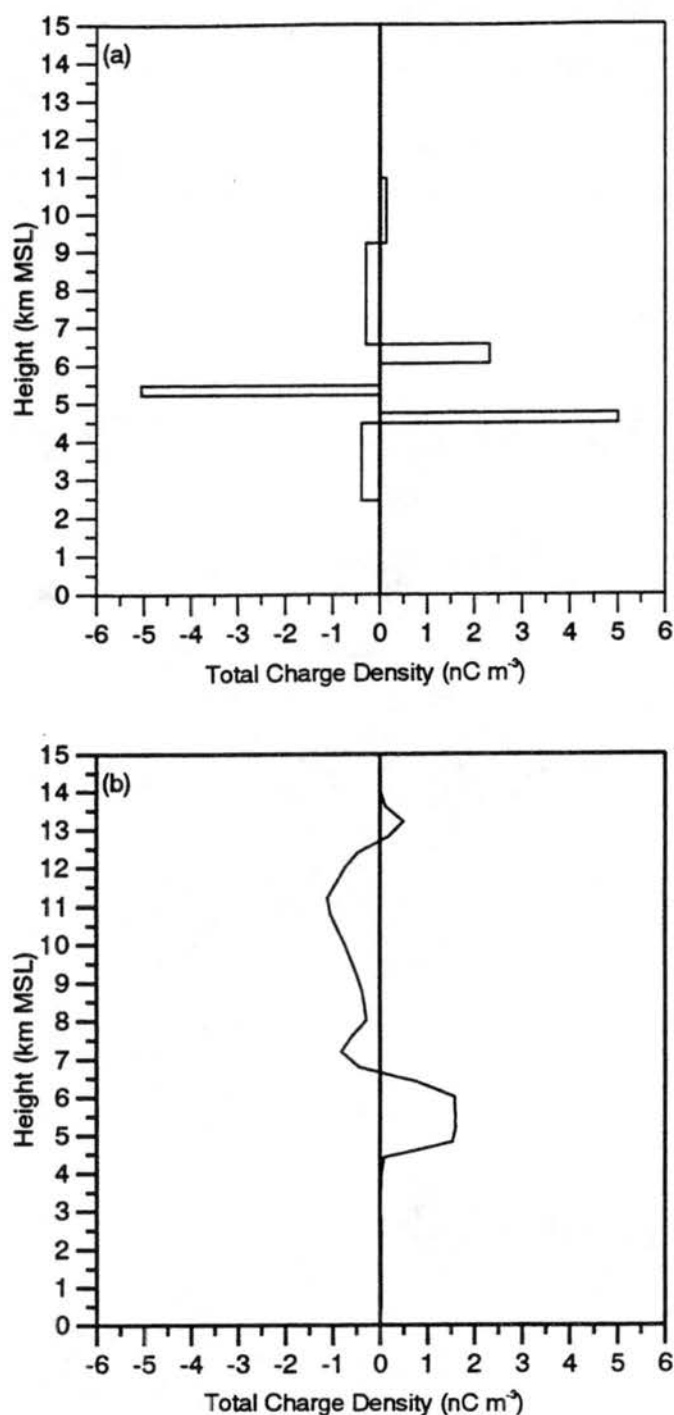


Fig. 7.30. Vertical charge density profile (nC m⁻³) (a) from symmetric MCS observations (cf. Fig. 3.12), and (b) constructed through the model grid depicted in Fig. 7.29 at 180 minutes into the simulation and a location 35 km behind the convective line. This is a location similar to where the balloon-borne EFM ascended through the symmetric MCS discussed in Chapter 3.

CHAPTER 8

SUMMARY AND CONCLUSIONS

8.1 Observations

In the observational component of this study, we examined the concurrent kinematic, microphysical, and electrical structures of two mesoscale convective systems. Multiple balloon-borne electric field observations in these Oklahoma MCSs revealed very different electrical structures. The symmetric MCS observed on 24 May 1991 had a rather complicated vertical structure consisting of five layers of charge while the asymmetric MCS observed on 8 May 1991 had a much simpler vertical structure consisting of three layers of charge. Peak electric fields were $> 100 \text{ kV m}^{-1}$ in the symmetric MCS and $> 80 \text{ kV m}^{-1}$ in the asymmetric MCS. In both cases, the sharpest charge transition and largest charge density magnitudes (derived using an approximation to Gauss' Law) were situated near the melting level. The symmetric MCS also had a sharp charge transition (negative charge above positive) at approximately -12°C ; this is a charge transition that was not present in the asymmetric MCS. The multiple profiles further suggest that stratiform electrical structures exhibited very little spatial and temporal variability at least over the course of the EFM observations, which spanned several hours.

Marshall and Rust (1993) suggested that the different electrical structures of these two MCSs might be related to differences in MCS flow structure, microphysical processes, or both. Our observational analyses revealed several notable differences between symmetric and asymmetric MCSs. First, particle trajectories computed from three-dimensional wind fields suggest that ice advected from the convective line of the asymmetric MCS follows a much more

line-parallel path than does ice advected from the convective line of the symmetric MCS. As a result, an ice particle's "residence time" in the transition zone downdraft is as much as three times longer in the asymmetric MCS than in the symmetric MCS. Several studies (e.g., Yeh et al., 1991; Braun and Houze, 1994) have indicated that mass growth due to both deposition and aggregation is suppressed in the transition zone downdraft. Given the dependence of most microphysical charging theories on the presence of cloud water and/or water saturated conditions, it is expected that non-inductive charging would be absent or greatly reduced in the transition zone. Second, upon reaching the MCS stratiform precipitation region, it appears that the advected ice in each system encountered a quite different kinematic and thermodynamic environment. That is, composited data from the NOAA P-3 tail-Doppler radar indicated that the symmetric MCSs mesoscale updraft was deepest and most intense near the leading edge of the stratiform precipitation region (the approximate location where the two primary electric field profiles were obtained). Furthermore, measurements of LWCs $> 0.2 \text{ g m}^{-3}$ and NOAA P-3 optical array probe observations of dendritic growth in this region both support the presence of widespread water saturated conditions. By comparison, the asymmetric MCS's composited vertical velocity structure exhibited a deep layer of descent, which would have acted to suppress supercooled liquid water. In situ aircraft observations indicated that water saturated conditions were much less widespread in the asymmetric MCS, consistent with extensive downward motion in the MCS stratiform region. It was argued that these contrasting kinematic and microphysical structures were at least partially responsible for the differences in the observed electrical structures.

When the levels of the charge transitions were compared to microphysical growth regimes, we found that the charge transition at -12°C in the symmetric MCS electric field profile was located immediately below the region where dendritic crystals are most favored (~ -12 to -18°C).

A charge transition at this level was not found in the asymmetric case, nor was supporting evidence for dendritic growth. Furthermore, we found that two charge transitions (at approximately -1 and -12°C) in the symmetric MCS electric field profile were collocated with levels of maximum aggregate size (e.g., Hobbs et al., 1974). Only the lower of these two charge transitions was found in the asymmetric MCS. From these comparisons, which qualitatively agree with other Marshall and Rust (1993) Type A (symmetric MCS) electric field profiles, we offer that gravitational separation of large and small hydrometeors carrying opposite polarities of charge might be responsible for at least some of the charge transitions revealed in the EFM soundings. Of course this argument assumes that charge generation processes are active at or near these levels. For example, numerical model simulations (summarized later) suggest that non-inductive charging (e.g., Takahashi, 1978; Saunders et al., 1991) can result in observed charge densities ($\sim 5 \text{ nC m}^{-3}$) in the stratiform precipitation region.

It is interesting to note the lower of these two levels, in addition to being at a temperature where large aggregates are commonly found (i.e., Hobbs et al., 1974), is also a region where observed small ice particle concentrations are quite large (i.e., Willis and Heymsfield, 1989). Therefore, if charge is generated by a fragmentation process (e.g., Avila and Caranti, 1994) or if opposite polarities of charge already exists on large and small hydrometeors, respectively, as a result of some other charging process (i.e., non-inductive charging), it is then possible that subsequent gravitational separation of the snow/aggregates and smaller ice crystals could result in a sharp charge transition. Other hypotheses for melting level charge transitions have been advanced by previous studies (e.g., Shepherd et al., 1996), but they do not typically account for charge transitions that occur at temperatures above the melting level. Unfortunately, from the observations we can only speculate on the source of charge in the MCS stratiform region. An

analysis of the relative contributions of various in situ charge generation mechanisms is addressed in more detail by the modeling component of this study, summarized next.

8.2 Modeling

In the modeling component of this study, we examined in situ charge production by non-inductive charging (Takahashi, 1978; Saunders et al., 1991). The relative contribution of charge advection from the convective line, evaporation-condensation charging (Dong and Hallett, 1992), and melting charging (Drake, 1968) were also investigated. The model control run (described in detail in Chapter 5) accurately reproduced many of the kinematic and microphysical features indicated by the observations. A rear-inflow jet developed at locations immediately below the melting level while at the same time front-to-rear flow aloft strengthened to $> 20 \text{ m s}^{-1}$. Vertical velocities produced by the model were as large as 50 cm s^{-1} (though time-averaged magnitudes were typically $< 20 \text{ cm s}^{-1}$), which compares well with composited aircraft observations. In general, the vertical velocity structure was characterized by a horizontally extensive region of ascent, particularly by 180 minutes into the simulation, overlying a deep region of subsidence. In the time-averaged model output, a narrow transition zone downdraft developed approximately 10 km behind the convective line. This is a feature that agrees well with our observations and those from previous studies (e.g., Biggerstaff and Houze, 1993). Due to the model's design, however, this transition zone downdraft is driven entirely by microphysical processes (and not mechanically forced) in our simulations.

In close agreement with observations, the model reflectivity structure exhibited maximum bright band reflectivities of approximately 45 dBZ by 180 minutes. The width of the transition zone, however, was consistently narrower than indicated by observations. It is possible that this difference is due to the use of a two-class ice scheme in the model. Maximum snow/aggregate mixing ratios were typically $< 1.2 \text{ g kg}^{-1}$ and cloud ice concentrations in the mixed-phase region

were maintained at $10\text{--}200\text{ l}^{-1}$ by fragmentation processes. As the mesoscale updraft strengthened and became more horizontally extensive, the snow and cloud ice categories experienced increased growth due to deposition. This is consistent with previous studies (e.g., Rutledge and Houze, 1987; Gallus and Johnson, 1991; Gallus and Johnson, 1995), which have indicated that in situ condensate production within the stratiform precipitation region is an important contributor to the MCSs water budget. Depositional growth in the mesoscale updraft also served to increase the average predicted ice crystal size in the model, which resulted in a larger charge transfer per collision with the Saunders et al. non-inductive charging parameterization (summarized below) at locations $> 20\text{ km}$ behind the convective line. At locations immediately above the melting level, however, cloud ice sizes were typically quite small as the ice sublimated in the mesoscale downdraft. In agreement with NOAA P-3 measurements, model cloud LWCs at locations $20\text{--}30\text{ km}$ behind the convective line were typically $0.2\text{--}0.3\text{ g m}^{-3}$.

As noted in Chapter 1, perhaps the most fundamental goal of the modeling component of this study was to investigate the relative contributions of 1) charge advection from the convective line, and 2) in situ charge generation resulting from microphysical processes within the stratiform region. These charging contributions were analyzed by constructing a charge budget that examined the charge flux through the model's right boundary and volume-integrated in situ charging rates over hour three of the model simulation. In this budget analysis, several different sizes of advected low-level ice crystal sizes were investigated. We found that the size of the advected ice crystals affected MCS charging in two ways: 1) by modifying the dynamical and microphysical evolution of the MCS, and 2) by modifying the charge transfer per collision (for the Saunders et al. parameterization). The budget analysis results indicated that, over the size range of ice crystals tested, non-inductive charging produced approximately 70% of the charge with the Takahashi parameterization and as much as 40% with the Saunders et al.

parameterization. It is interesting to note that the Takahashi-based charging result is consistent with the approximate contribution that mesoscale ascent (in situ condensate production) makes to the total stratiform precipitation (e.g., Rutledge and Houze, 1987). This suggests that stratiform electrification is strongly tied to stratiform precipitation processes.

Model simulations were also conducted to individually investigate charge advection from the convective line and local charge generation from in situ microphysical processes. A model simulation that investigated the advection of charge (using a simple thunderstorm tripole charge structure to initialize the model buffer zone) indicated that turbulent diffusion in the model formulation results in only minor changes to the advected charge distribution. Perhaps the most important result of the non-inductive charging simulations is that the model generated charge densities ($\sim 10 \text{ nC m}^{-3}$ with the Takahashi data and $\sim 3 \text{ nC m}^{-3}$ with the Saunders et al. data) that were quite similar in magnitude to those derived from balloon-borne electric field measurements ($\sim 5 \text{ nC m}^{-3}$). Furthermore, the model also indicates that, once these charge densities were achieved, the loss of charge resulting from fallspeed divergence became approximately equal to the rate of charge generation by the non-inductive process, consistent with the quasi-steady layered structure that is commonly seen in the observations (e.g., Marshall and Rust, 1993; Stolzenburg et al., 1994). Nevertheless, while the model has demonstrated that in situ charging is sufficient to explain charge generation in the stratiform regions lower levels, it is equally apparent that much of the charge observed at colder temperatures (i.e., $< -20^\circ\text{C}$) is more likely the result of charge advection from the convective line since cloud water, which is necessary for non-inductive charging, is quite rare at those colder temperatures. We now summarize and discuss the results from the various electrification simulations.

While charge densities produced by model simulations exhibit marked similarities to observations, considerably less agreement is found when the modeled charge *distribution* is

compared to observed charge structures. Some observed charge features, however, were reproduced quite well. In agreement with the charge profiles presented by Shepherd et al. (1996), the Takahashi data produced a sharp transition near the melting level. Shepherd et al. (1996) speculated that at least some charge transitions near 0°C may be the result of charging processes associated with melting (e.g., Simpson, 1909; Drake, 1968). In our numerical simulation, the melting level charge transition was produced when different polarity charge densities were carried by the cloud ice (negative) and snow (positive) at temperatures $< 0^{\circ}\text{C}$. Near the melting level, cloud ice was quickly removed by sublimation as it encountered the mesoscale downdraft (or melting and subsequent evaporation), thereby exposing the positive charge on the larger hydrometeors and creating a sharp charge transition. Subsequent melting and fallspeed divergence of rain resulted in a thin layer of positive charge situated immediately below the melting level. Given the quick removal of cloud ice at 0°C in the model parameterizations, it is likely that this process is more efficient at producing a charge layer in the numerical model than near the melting level of a real stratiform cloud. Nevertheless, this result does qualitatively agree with observations, which indicated that gravitational separation between oppositely charged large and small hydrometeors might occur near 0°C (see discussion above). Furthermore, though we do not have high quality NOAA P-3 optical array probe data through the melting level of the 24 May 1991 MCS, particle charge measurements by Bateman et al. (1995) indicated that as much as 90% of the charge at these levels resides on small (i.e., $< 800\ \mu\text{m}$) ice particles. Unfortunately, if this process were indeed active, we do not know the disposition of ions released by this sublimation/evaporation (or other processes such as corona and lightning) since our model does not include continuity equations for ions (due to computational constraints). Clearly, these model results should be interpreted with this deficiency in mind.

When the Saunders et al. data were used in the model, a layer of positive charge that qualitatively agreed with a similar layer in the symmetric MCS electric field profiles was produced in the mixed phase region (with the primary charge transition of negative above positive occurring between -12 and -15°C). Unlike with the Takahashi data, however, a sharp charge transition was not produced near the melting level. This was primarily due to the dependence of the Saunders et al. non-inductive parameterization on cloud ice size. Since locations immediately above the melting level were typically associated with subsidence, ice crystal sizes were generally quite small and, as a result, the amount of charge transfer per collision was also reduced according to the Saunders et al. charging formulations. Since Takahashi did not investigate the dependence of charge transfer on cloud ice size, we do not know if a similar, unaccounted for dependence biased the results of the Takahashi simulations (discussed above).

Since electric field observations indicated predominantly negative charge in the upper levels of the MCS stratiform region, it is not likely that the tripole thunderstorm structure (e.g., Williams, 1989) used to initialize the convective line was representative of the convective charge structure of the 24 May 1991 MCS. Sensitivity tests conducted with a more complicated charge initialization profile for the convective line (Fig. 7.25), which exhibited many similarities to the symmetric MCS stratiform charge structure derived in this study, indicated that the advected charge maintained its layered structure to a location approximately 35-45 km behind the convective line. When combined with non-inductive charging from Takahashi (1978), however, the advected charge profile was quickly overwhelmed by in situ charge generation within the stratiform region; when combined with non-inductive charging from Saunders et al. (1991), the intensity of the mid-level (~ -12 to -15°C) positive charge layer in the advective profile was enhanced at location approximately 35 km behind the convective line (the approximate location

where balloon-borne EFM observations were obtained in the symmetric MCS). Subsequent descent of this charge region, however, resulted in quick dissipation of the layered structure.

Evaporation-condensation charging (e.g., Dong and Hallett, 1992) and melting charging (e.g., Drake, 1968) were also examined. Both of these mechanisms have been discussed in previous studies (e.g., Marshall and Rust, 1993; Rutledge and Petersen, 1994; Stolzenburg et al., 1994) as possible sources of charge in MCS stratiform regions. The simulations wherein the evaporation-condensation mechanism was tested, however, produced charge densities that were two orders of magnitude smaller than indicated by the observations. Furthermore, charge transitions did not occur at observed temperature levels. Simulations that tested the melting mechanism of Drake (1968) produced positive charge densities that were $< 0.4 \text{ nC m}^{-3}$. Converting maximum snow mixing ratios to charging rates using the data presented by Drake would suggest that charge densities of $> 1.8 \text{ nC m}^{-3}$ are possible with this mechanism. However, rapid fallspeed divergence resulted in a significant reduction in charge density at locations immediately below the melting level. Furthermore, we note that the applicability of the mechanism discussed by Drake (1968) is uncertain in MCS stratiform precipitation regions since the predominant hydrometeor type (aggregates) is quite different from that (ice spheres) studied by Drake, as the physics of the melting process is not the same between the two hydrometeor types.

8.3 MCS structure and evolution

While the observational analysis indicates that the symmetric and asymmetric MCS had very different concomitant kinematic, microphysical, and electrical structures, it should be noted that the data obtained in each system were collected at different stages of their life cycles. That is, the majority of the data in the symmetric MCS were collected during the mature stage while the majority of the data in the asymmetric MCS were collected during the dissipative stage. Since

several studies (e.g., Rutledge et al., 1988; Blanchard, 1990; Loehrer and Johnson, 1995) have indicated that symmetric type MCSs often evolve into asymmetric MCSs later in their lifetime, it is possible that a similar evolution from a Type A to a Type B electrical structure (e.g., Marshall and Rust, 1993) might occur. That is, as a system evolves from a symmetric to asymmetric structure, the concurrent dynamical evolution of the MCS stratiform precipitation region, as influenced by heat and hydrometeor advection from the convective line, may be modified such that the electrical structure that results from charge advection and in situ generation process may also be modified. Though we are unable to explicitly model the more three-dimensional asymmetric system with our 2-D model, sensitivity tests clearly indicated that the MCS stratiform regions kinematic, microphysical, and electrical evolution was sensitive to convective influences other than hydrometeor and/or charge advection.

8.4 Conclusions and recommendations for future research

Here we summarize the primary conclusions. In the observational component of the study, they are:

- Ice particles detrained from the convective line of symmetric MCSs have much shorter “residence times” in the unfavorable growth/charging region associated with the transition zone downdraft.
- Upon entering the MCS stratiform region, ice particles in the symmetric MCS appear to enter kinematic/microphysical environments that are more conducive to in situ charging.
- Two primary charge transitions in the symmetric MCS occurred at levels where local maximum aggregational growth is commonly found. Only the lower charge transition occurred in the asymmetric MCS. Evidence for the upper-most aggregation layer in the asymmetric MCS was not found.

- The collocation of charge transition levels with aggregation maxima suggests that weak gravitational separation of charge might be responsible for some charge transitions in the stratiform region.
- Since symmetric MCSs occasionally evolve into asymmetric MCSs, observed charge structures may be related to MCS evolutionary stage.

The main conclusions of the modeling component of the study are:

- Both (Takahashi, 1978 and Saunders et al., 1991) non-inductive charging parameterizations, in a low liquid water content cloud, produce charge densities similar to those indicated by electric field meter observations.
- A charge budget analysis indicates that in situ charge production accounts for as much as 70% of stratiform charge. This result is consistent with the approximate contribution that mesoscale ascent (in situ condensate production) makes to the total stratiform precipitation, suggesting that stratiform electrification may be related to stratiform precipitation processes.
- Both non-inductive charging parameterizations reproduce some of the observed stratiform charge features. The best results were obtained through a combination of charge advection and non-inductive charging processes.
- Once observed charge densities are achieved, the sink of charge resulting from fallspeed divergence becomes approximately equal to the rate of charge generation by the non-inductive process. It is possible that this leads to the quasi-steady layered structure that is commonly seen in the observations.
- In charge advection simulations, turbulent diffusion results in only minor modifications to the MCSs charge structure.

The results from this study point to a need to obtain more comprehensive observations of MCS kinematic, microphysical, and electrical structures. Foremost, a more complete documentation of MCS microphysical structure is needed. The microphysical data presented here often contained gaps (resulting from instrumentation problems) that greatly inhibited our ability to interpret possible relationships between MCS microphysical and electrical structures. Furthermore, while the collection of several multiple-EFM data sets in recent years has provided much information on MCS electrification, conceptual models are still based on a relatively small number of samples. Therefore, more in situ electric field observations (along with measurements of particle charge) are needed to improve upon conceptual models and furthermore to provide a more comprehensive data set to which theoretical laboratory studies and numerical model output may be compared. The observations should strive to include multiple, simultaneous EFM ascents to better document variability in convective and stratiform charge structures within individual MCSs. Particle charge measurements need also be made within convective regions in order that more realistic charge advection profiles may be developed for numerical modeling studies similar to the one presented here. Finally, the origin and effect of lightning in MCS stratiform precipitation regions is poorly understood. Therefore, future field efforts should strive to include a three-dimensional mapping system (e.g., Richard, 1990) that would provide information on the origins of lightning discharges and their relationship to the MCS electrical structure.

The model simulations presented here reveal much information about the kinematic and in situ microphysical processes that account for MCS electrification. However, they also demonstrate a need to develop improved microphysical parameterizations. Unfortunately, our ability to improve upon parameterizations is often limited by the mathematics of using bulk microphysical parameterizations and, furthermore, by uncertainties in laboratory observations.

For example, laboratory studies typically address electrification that occurs when cloud ice collides with a graupel particle. Since graupel are rarely found in MCS stratiform region, we can only extrapolate the results of Takahashi (1978) and Saunders et al. (1991) to include collisions between snow and ice. A better understanding of charge separation mechanism in low LWC regimes is also required.

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APPENDIX A

MICROPHYSICAL PARAMETERIZATIONS

The simulations presented in this study use a two-class ice scheme similar to that presented by Rutledge and Hobbs (1983). Here we briefly discuss the microphysical parameterizations used for these simulations. In particular, we focus on modifications and additions that have been made to improve upon the prediction of cloud water and cloud ice, both of which are important to understanding electrification processes. A more comprehensive description of other mixing ratio parameterizations can be found in Rutledge and Hobbs (1983) while a discussion of a continuity equation for cloud ice concentration (and relative parameterizations) is presented in detail in Chapter 5.

A.1 Mass weighted fallspeeds

All hydrometeors in the precipitating fields (snow/aggregate and rain) are assumed to fall at their mass-weighted fallspeeds given by

$$\bar{V}_x = \frac{\int_0^{\infty} n_x(D_x) M(D_x) V_x(D_x) dD_x}{\int_0^{\infty} n_x(D_x) M(D_x) dD_x} \quad (A1)$$

where $n_x(D_x)$ is the number concentration (given by Eqn. 5.1), $M(D_x) = \pi / 6 (\rho_x D_x^3)$ the hydrometeor mass, and $V_x(D_x)$ the hydrometeor fallspeed.

As in Rutledge and Hobbs (1983), a polynomial fit to the experimental data of Gunn and Kinzer (1949) is used to determine fallspeeds for the rain category, i.e.,

$$V_r(D_r) = -0.267 + 51.5D_r - 102.25D_r^2 + 75.5D_r^3 \quad (\text{A2})$$

where D_r is in cm and V_r in m s^{-1} . Substituting Eqn. A2 into Eqn. A1 and integrating yields a mass weighted fallspeed for the rain category of

$$\bar{V}_r = \left(-0.267 + \frac{206.0}{\lambda_r} - \frac{2.045 \times 10^3}{\lambda_r^2} + \frac{9.06 \times 10^3}{\lambda_r^3} \right) \left(\frac{p_0}{p} \right)^{0.4} \quad (\text{A3})$$

where λ_r is given by Eqn. 5.2 and the factor $(p_0 / p)^{0.4}$ allows for the change in fallspeed with air pressure (e.g., Foote and DuToit, 1969).

The fallspeed for the snow category follows Locatelli and Hobbs (1974), i.e.,

$$V_s(D_s) = a_s D_s^{b_s} \left(\frac{p_0}{p} \right)^{0.4} \quad (\text{A4})$$

From substituting Eqn 5.1 and Eqn. A4 into Eqn. A1 and integrating, we get

$$\bar{V}_s = a_s \frac{\Gamma(4 + b_s)}{6\lambda_s^{b_s}} \left(\frac{p_0}{p} \right)^{0.4} \quad (\text{A5})$$

where λ_s is given by Eqn. 5.1. The values of a_s and b_s , as shown in Appendix B, differ from the values used by Rutledge and Hobbs (1983) because of corrections suggested by Potter (1991).

A.2 Sources and sinks of water continuity variables

As noted earlier, many of the mixing ratio parameterizations used in this study follow Rutledge and Hobbs (1983). As in Gallus (1993), these parameterizations are presented in this section with little detail. A more complete description, however, is provided for parameterizations (e.g., PINT, PDEPI, PSDEP, PCONV, PSFW, PSFI, PFRAG) that have been modified to improve upon the prediction of cloud water and cloud ice mixing ratios in the model.

1) CONDENSATION AND EVAPORATION OF CLOUD WATER

$$P_{COND} = \rho(r_v - r_{ws}) \left[\Delta t \left(1 + \frac{L_v^2 r_{ws}}{c_p R_w T^2} \right) \right]^{-1} \quad (A6)$$

2) AUTOCONVERSION OF CLOUD WATER

$$P_{RAUT} = k_r \rho(r_r - r_0) \quad (A7)$$

3) COLLECTION OF CLOUD WATER BY RAIN WATER

$$P_{RACW} = \frac{\pi}{4} \rho r_c E_{rc} N_{or} \left[\frac{a_{r0} \Gamma(3)}{\lambda_r^3} + \frac{a_{r1} \Gamma(4)}{\lambda_r^4} + \frac{a_{r2} \Gamma(5)}{\lambda_r^5} + \frac{a_{r3} \Gamma(6)}{\lambda_r^6} \right] \left(\frac{p_0}{p} \right)^{0.4} \quad (A8)$$

4) EVAPORATION OF RAIN WATER

$$P_{REVP} = \frac{2\pi N_{or} (S_w - 1)}{A_w + B_w} \left[\frac{0.78}{\lambda_r^2} + 0.31 \frac{(a_r \rho / \mu)}{\lambda_r^3} \Gamma(3) \left(\frac{p_0}{p} \right)^{0.2} \right] \quad (A9)$$

where A_w and B_w are given by

$$A_w = \frac{L_v}{K_a T} \left(\frac{L_v M_w}{R^* T} - 1 \right) \quad (A10)$$

and

$$B_w = \frac{R^* T}{\chi M_w e_{sw}} \quad (A11)$$

5) INITIATION OF CLOUD ICE

The initiation of cloud ice is addressed in Chapter 5 (see Eqns. 5.7-5.12).

6) DEPOSITIONAL GROWTH OF CLOUD ICE AND SNOW

Model sensitivity tests indicated that depositional growth of the cloud ice and snow/aggregate categories, when treated separately, resulted in cloud LWCs that were much lower than indicated by observations. Thus, depositional growth in this study is computed by

adding the cloud ice and snow/aggregate mixing ratios together, assuming that the combined ice category has an exponential distribution, and computing the bulk deposition rate by

$$PDEP = \frac{4(S_i - 1)N_{os}}{A_i + B_i} \left[\frac{0.65}{\lambda_{si}^2} + 0.44 \left(\frac{a_s \rho}{\mu} \right)^{0.5} \frac{\Gamma(b_s / 2 + 5 / 2)}{\lambda_{si}^{b_s/2+5/2}} \right] \left(\frac{p_0}{p} \right)^{0.2} \quad (A12)$$

where λ_{si} is the slope of the combined cloud ice and snow/aggregate categories and A_i and B_i are given by

$$A_i = \frac{L_s}{K_a T} \left(\frac{L_s M_w}{R^* T} - 1 \right) \quad (A13)$$

and

$$B_i = \frac{R^* T}{\chi M_w e_{si}} \quad (A14)$$

The depositional growth of the cloud ice (PDEPI) and snow (PSDEP) categories, respectively, are then given by

$$PDEPI = PDEP \left(\frac{\sigma_i}{\sigma_i + \sigma_s} \right) \quad (A15)$$

and

$$PSDEP = PDEP \left(\frac{\sigma_s}{\sigma_i + \sigma_s} \right) \quad (A16)$$

where σ_i and σ_s represented the total surface areas of the ice and snow categories (cf. Eqns. 5.27 and 5.28), respectively.

7) BERGERON PROCESS

To further improve cloud LWC prediction in the model, a Bergeron process parameterization following Lin et al. (1983) was added to the model microphysics. In this parameterization, two terms (PSFW and PSFI) describe the rates at which cloud water and cloud ice, respectively,

transform to snow by deposition and riming (modified here to based on the growth of the predicted ice size). Following Lin et al. (1983), the equations for these growth rates are written as

$$\text{PSFW} = N_i \left(a_{b1} m_i a_{b2} + \pi E_{ic} \rho r_c (D_i / 2)^2 \bar{V}_i \right) \quad (\text{A17})$$

and

$$\text{PSFI} = \frac{r_c}{\Delta t_1} \quad (\text{A18})$$

where the constants a_{b1} and a_{b2} are taken from Koenig (1971), and the temperature dependent time scale Δt_1 is discussed in Hsie et al. (1980). A more complete description of this parameterization can be found in Lin et al. (1983).

8) CONVERSION OF CLOUD ICE TO SNOW

The autoconversion of cloud ice to snow has been modified to account for predicted ice sizes.

That is,

$$\text{PCONV} = \rho \left(\frac{r_i - r_{i\max}}{\Delta t} \right) \quad (\text{A19})$$

where $r_{i\max} = (M_{\max} N_i / \rho)$ and N_i is the predicted ice concentration as determined by Eqn. 5.5.

9) COLLECTION OF CLOUD ICE BY SNOW

$$\text{PSACI} = \frac{\pi}{4} \rho a_s r_i E_{si} N_{os} \frac{\Gamma(b_s + 3)}{\lambda_s^{b_s+3}} \left(\frac{p_0}{p} \right)^{0.4} \quad (\text{A20})$$

10) COLLECTION OF CLOUD WATER BY SNOW

$$\text{PSACW} = \frac{\pi}{4} \rho a_s r_c E_{sc} N_{os} \frac{\Gamma(b_s + 3)}{\lambda_s^{b_s+3}} \left(\frac{p_0}{p} \right)^{0.4} \quad (\text{A21})$$

11) MELTING OF SNOW

$$PSMLT = -\frac{2\pi N_{os}}{L_f} K_a (T - T_0) \left[\frac{0.65}{\lambda_s^2} + 0.44 \left(\frac{a_s \rho}{\mu} \right)^{0.5} \frac{\Gamma(b_s / 2 + 5 / 2)}{\lambda_s^{b_s/2+5/2}} \right] \left(\frac{p_0}{p} \right)^{0.2} \quad (A22)$$

12) EVAPORATION OF MELTING SNOW

$$PMLTEV = \frac{4(S_w - 1)N_{os}}{A_w + B_w} \left[\frac{0.65}{\lambda_s^2} + 0.44 \left(\frac{a_s \rho}{\mu} \right)^{0.5} \frac{\Gamma(b_s / 2 + 5 / 2)}{\lambda_s^{b_s/2+5/2}} \right] \left(\frac{p_0}{p} \right)^{0.2} \quad (A23)$$

13) MELTING OF CLOUD ICE

$$PSMLTI = \frac{\rho r_i}{\Delta t} \quad (A24)$$

for $T \geq 0^\circ \text{C}$ only.

14) ICE MULTIPLICATION

Ice multiplication through fragmentation is addressed in Chapter 5 (see Eqns. 5.13 and 5.14).

A.3 Source terms for the water continuity variables

The source terms for the five water continuity variables (see Eqns. 5.3 and 5.4) are given by

1) WATER VAPOR

$$\frac{dr_v}{dt} = -[P_{COND} + P_{REVP} + P_{SDEP} + P_{MLTEV}(T \geq 0^\circ \text{C}) + P_{DEPI} + P_{INT}] \quad (A25)$$

2) CLOUD WATER

$$\frac{dr_c}{dt} = P_{COND} + PSMLTI(T \geq 0^\circ \text{C}) - P_{RAUT} - P_{RACW} - P_{SACW} - P_{SFW} \quad (A26)$$

3) CLOUD ICE

$$\frac{dr_i}{dt} = P_{DEPI} - PSMLTI(T \geq 0^\circ \text{C}) - P_{SACI} - P_{CONV} + P_{INT} + P_{FRAG} - P_{SFI} \quad (A27)$$

4) RAIN

$$\frac{dr_r}{dt} = \text{PREVP} + \text{PRAUT} + \text{PRACW} - \text{PSMLT}(T \geq 0^\circ\text{C}) + \text{PSACW}(T \geq 0^\circ\text{C}) \quad (\text{A28})$$

5) SNOW

$$\begin{aligned} \frac{dr_s}{dt} = & \text{PSDEP} = \text{PMLTEV}(T \geq 0^\circ\text{C}) + \text{PSACI} + \text{PSMLT}(T \geq 0^\circ\text{C}) + \text{PSACW}(T \leq 0^\circ\text{C}) \\ & + \text{PCONV} - \text{PFRAG} + \text{PSFW} + \text{PSFI} \end{aligned} \quad (\text{A29})$$

while the source term for the total microphysical diabatic heating used in the thermodynamic energy equation (see Gallus, 1993) is given by

$$\begin{aligned} S_h = & \frac{L_v}{c_p} [\text{PCOND} + \text{PREVP} + \text{PMLTEV}] + \frac{L_s}{c_p} [\text{PINT} + \text{PDEPI} + \text{PSDEP}] \\ & + \frac{L_f}{c_p} [\text{PSMLT} - \text{PSMLTI} + \text{PSACW}(T \leq 0^\circ\text{C})] \end{aligned} \quad (\text{A30})$$

APPENDIX B

LIST OF SYMBOLS

Symbols	Description	Value	units
a	Coefficient for non-inductive charging		
a_{b1}	Coefficient for Bergeron process		$\text{kg}^{(1-a_{b2})} \text{s}^{-1}$
a_{b2}	Coefficient for Bergeron process		
A_w	Thermodynamic term		m s kg^{-1}
A_i	Thermodynamic term		m s kg^{-1}
a_s	Constant in fallspeed relation for snow	12.37	$\text{m}^{(1-b_s)} \text{s}^{-1}$
a_r	Constant in linear fallspeed relation for rain	3×10^3	s^{-1}
a_{r0}	Coefficient in polynomial fallspeed relation for rain	-0.267	m s^{-1}
a_{r1}	Coefficient in polynomial fallspeed relation for rain	5.15×10^3	s^{-1}
a_{r2}	Coefficient in polynomial fallspeed relation for rain	-1.0225×10^6	$\text{m}^{-1} \text{s}^{-1}$
a_{r3}	Coefficient in polynomial fallspeed relation for rain	7.55×10^7	$\text{m}^{-2} \text{s}^{-1}$
a_{q0}	Coefficient in polynomial for depositional charging	-1.286	
a_{q1}	Coefficient in polynomial for depositional charging	-0.4477	
a_{q2}	Coefficient in polynomial for depositional charging	-0.0025	
a_{q3}	Coefficient in polynomial for depositional charging	0.0008	
B_w	Thermodynamic term		m s kg^{-1}
B_i	Thermodynamic term		m s kg^{-1}
b	Coefficient for non-inductive charging		
b_s	Fallspeed exponent for snow	0.42	
c_p	Specific heat of air at constant pressure	1004.0	$\text{J kg}^{-1} \text{K}^{-1}$
c_w	Specific heat of liquid water at 0°C	4218.0	$\text{J kg}^{-1} \text{K}^{-1}$
D_i	Ice crystal diameter		m
D_m	Characteristic mean diameter of precipitation		m
D_r	Raindrop diameter		m
D_s	Snow diameter		m
E	Electric field magnitude		V m^{-1}
E_{ic}	Cloud ice/cloud water collection efficiency	0.1	
E_{rc}	Rain/cloud water collection efficiency	0.01	
E_{ri}	Rain/cloud ice collection efficiency	1.0	
E_{sc}	Snow/cloud water collection efficiency	0.01	
E_{si}	Snow/cloud ice collection efficiency	0.01	
E_{sr}	Snow/rain collection efficiency	1.0	
E_x	Horizontal electric field magnitude		V m^{-1}
E_z	Vertical electric field magnitude		V m^{-1}

Symbols	Description	Value	units
e_{si}	Saturation vapor pressure for ice		$N m^{-2}$
e_{sw}	Saturation vapor pressure for water		$N m^{-2}$
F_q	Charging rate for depositional mechanism		$C m^{-2}$
k_r	Rate coefficient for autoconversion	0.001	s^{-1}
k_{frag}	Rate coefficient for fragmentation	0.5×10^{-5}	s^{-1}
K_a	Thermal conductivity of air	2.43×10^{-2}	$J m^{-1} s^{-1} K^{-1}$
K_m	Eddy exchange coefficient		$m^2 s^{-1}$
L_f	Latent heat of fusion	3.337×10^5	$J kg^{-1}$
L_s	Latent heat of sublimation	2.834×10^6	$J kg^{-1}$
L_v	Latent heat of vaporization	2.5×10^6	$J kg^{-1}$
m_i	Mass of ice crystal		kg
M_{frag}	Mass of an ice fragment from fragmentation function	1.0×10^{-10}	kg
M_{max}	Maximum mass of cloud ice crystal	6.0×10^{-10}	kg
M_o	Mass of an initiated ice crystal	1.0×10^{-12}	kg
M_w	Molecular weight of water	18.016	
n_c	Number concentration of nucleated ice crystals		m^{-3}
n_{c_1}	Number concentration of nucleated ice crystals ($T_c \geq -5^\circ C$)		m^{-3}
n_{c_2}	Number concentration of nucleated ice crystals ($T_c < -5^\circ C$)		m^{-3}
n_{∞_1}	Constant in ice nucleation ($T_c \geq -5^\circ C$)	50.0	m^{-3}
n_{∞_2}	Constant in ice nucleation ($T_c < -5^\circ C$)	1000.0	m^{-3}
N_i	Number concentration of ice crystals		m^{-3}
N_α	Intercept value in raindrop size distribution	8.0×10^6	m^{-4}
N_α	Intercept value in snow size distribution	4.0×10^6	m^{-4}
p	Pressure		$N m^{-2}$
p_0	Constant in empirical relation	10^5	$N m^{-2}$
NINT	Initiation of cloud ice		m^{-3}
PCOND	Condensation of water vapor		$kg m^{-3} s^{-1}$
PCONV	Conversion of cloud ice to snow		$kg m^{-3} s^{-1}$
PDEP	Depositional growth of snow/cloud ice		$kg m^{-3} s^{-1}$
PDEPI	Depositional growth of ice		$kg m^{-3} s^{-1}$
PFRAG	Fragmentation of snow to create cloud ice		$kg m^{-3} s^{-1}$
PIACR	Collection of rain by cloud ice		$kg m^{-3} s^{-1}$
PINT	Initiation of cloud ice		$kg m^{-3} s^{-1}$
PMLTEV	Evaporation of melting snow		$kg m^{-3} s^{-1}$
PRACI	Collection of cloud ice by rain		$kg m^{-3} s^{-1}$
PRACS	Collection of snow by rain		$kg m^{-3} s^{-1}$
PRACW	Collection of cloud water by rain		$kg m^{-3} s^{-1}$
PRAUT	Autoconversion of cloud water		$kg m^{-3} s^{-1}$
PREVP	Evaporation of rain		$kg m^{-3} s^{-1}$
PSACI	Collection of cloud ice by snow		$kg m^{-3} s^{-1}$
PSACR	Collection of rain by snow		$kg m^{-3} s^{-1}$
PSACW	Collection of cloud water by snow		$kg m^{-3} s^{-1}$
PSDEP	Depositional growth of snow		$kg m^{-3} s^{-1}$

Symbols	Description	Value	units
PSFI	Conversion of cloud ice to snow in the Bergeron process		$\text{kg m}^{-3} \text{s}^{-1}$
PSFW	Conversion of cloud water to snow in the Bergeron process		$\text{kg m}^{-3} \text{s}^{-1}$
PSMLT	Melting of snow		$\text{kg m}^{-3} \text{s}^{-1}$
PSMLTI	Melting of cloud ice		$\text{kg m}^{-3} \text{s}^{-1}$
PWACS	Collection of snow by cloud water		$\text{kg m}^{-3} \text{s}^{-1}$
q'	Non-inductive charge transfer		C
Q_c	Charge density of cloud water		C m^{-3}
Q_i	Charge density of cloud ice		C m^{-3}
Q_m	Charging rate for melting mechanism		C m^{-3}
Q_r	Charge density of rain		C m^{-3}
Q_s	Charge density of snow		C m^{-3}
Q_T	Total charge density		C m^{-3}
r_c	Mixing ratio of cloud water		kg kg^{-1}
r_{frag}	Fragmentation mixing ratio threshold for snow		kg kg^{-1}
r_i	Mixing ratio of cloud ice		kg kg^{-1}
r_{is}	Ice saturation mixing ratio		kg kg^{-1}
r_{imax}	Mixing ratio threshold for PCONV		kg kg^{-1}
r_r	Mixing ratio of rain		kg kg^{-1}
r_s	Mixing ratio of snow		kg kg^{-1}
r_v	Mixing ratio of water vapor		kg kg^{-1}
r_{ws}	Water saturation mixing ratio		kg kg^{-1}
r_0	Mixing ratio threshold for PRAUT	0.001	kg kg^{-1}
R^*	Universal gas constant	8314.0	$\text{J kmol}^{-1} \text{K}^{-1}$
R_w	Gas constant for water vapor	461.0	$\text{J kg}^{-1} \text{K}^{-1}$
S_c	Schmidt number	0.6	
S_h	Diabatic heating term		$\text{K kg m}^{-3} \text{s}^{-1}$
S_i	Saturation ratio with respect to ice		
S_w	Saturation ratio with respect to water		
T	Temperature		K
T_c	Air temperature		$^{\circ}\text{C}$
T_0	Reference temperature	273.16	K
u	Horizontal x-velocity		m s^{-1}
$\bar{V}$	Mass-weighted fallspeed of precipitation		m s^{-1}
$\bar{V}_i$	Average fallspeed of cloud ice		m s^{-1}
$\bar{V}_r$	Mass-weighted fallspeed of rain		m s^{-1}
$\bar{V}_s$	Mass-weighted fallspeed of snow		m s^{-1}
w	Vertical z-velocity		m s^{-1}
x	Horizontal distance		m
z	Vertical distance		m
α	Specific volume of air		$\text{m}^3 \text{kg}^{-1}$
α_0	Reference specific volume		$\text{m}^3 \text{kg}^{-1}$
α_1	First exponent in ice nucleation ($T_c \geq -5^{\circ}\text{C}$)	4.5	
α_2	First exponent in ice nucleation ($T_c < -5^{\circ}\text{C}$)	12.96	
β_1	Second exponent in ice nucleation ($T_c \geq -5^{\circ}\text{C}$)	0.6	K^{-1}

Symbols	Description	Value	units
β_2	Second exponent in ice nucleation ($T_c < -5^\circ\text{C}$)	0.639	
δq	Non-inductive charge transfer		C
ϵ_0	Permittivity of free space	8.85×10^{-12}	$\text{C}^2 \text{N}^{-1} \text{m}^{-2}$
Γ	Gamma function		
ρ	Air density		kg m^{-3}
ρ_L	Density of water	1000.0	kg m^{-3}
ρ_s	Density of snow	200.0	kg m^{-3}
λ_r	Slope of raindrop size distribution		m^{-1}
λ_s	Slope of snow size distribution		m^{-1}
λ_{si}	Slope of snow/cloud ice size distribution		m^{-1}
χ	Diffusivity of water vapor in air	2.26×10^{-5}	$\text{m}^2 \text{s}^{-1}$
μ	Dynamic viscosity of air	1.718×10^{-5}	$\text{kg m}^{-1} \text{s}^{-1}$
Φ	Electrostatic potential		V
α_i	Total surface area of ice		m^2
α_s	Total surface area of snow		m^2
Δt	Time increment	15.0	s
Δx	Horizontal spatial increment	5000.0	m
Δz	Vertical spatial increment	400.0	m
Δt_1	Time scale for Bergeron process		s