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**MULTI-SCALE PROCESS STUDIES IN THE TROPICS:
RESULTS FROM LIGHTNING OBSERVATIONS**

by Walter A. Petersen



Steven A. Rutledge, P.I.

**Colorado
State
University**

**DEPARTMENT OF
ATMOSPHERIC SCIENCE**

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Walter A. Petersen

Department of Atmospheric Science

Colorado State University

Fort Collins, CO 80523-1371

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ABSTRACT

MULTI-SCALE PROCESS STUDIES IN THE TROPICS: RESULTS FROM LIGHTNING OBSERVATIONS

Cloud-to-ground (CG) lightning and meteorological observations collected in the tropics were analyzed to address the following question: What do observations of lightning tell us about processes occurring over *multiple scales* in the tropical atmosphere? An emphasis was placed on the analysis of observations collected over the western Pacific warm-pool during TOGA COARE.

Large-scale observations from COARE suggest that the occurrence of lightning over the western Pacific Ocean is a sensitive function of *both* the magnitude and vertical distribution of convective available potential energy (CAPE). Small variations in the marine boundary layer humidity were highly correlated to variations in the CAPE and/or boundary-layer θ_w . In turn, small increases ($O[0.5^\circ \text{C}]$) in the boundary-layer θ_w were associated with disproportionate increases in lightning activity. The diurnal cycle of CG lightning exhibited a pronounced maximum (minimum) around 2 a.m. (12 p.m.) local-time. Diurnal cycles of CAPE, convective and total precipitation exhibited similar diurnal cycles, but were weaker in amplitude.

Over cloud-scales, upward-building 30 dBZ reflectivity cores extended to elevations colder than -10°C in lightning-producing tropical oceanic convection. Additionally, mean updraft strengths (when observed) in several lightning-producing cases exceeded 6 m s^{-1} near the -10°C

level. These observations support the hypothesis that updraft magnitudes between the 0°C and -10°C levels in tropical convection must exceed the terminal fall-speed of millimeter sized liquid and frozen drops in order to provide the requisite hydrometeor mass to electrification processes in the cold regions of the cloud.

To investigate the coupling between cloud-scale electrification, kinematics, microphysics, and the large-scale thermodynamic environment, a one-dimensional cloud-model with a four-class bulk-microphysical ice scheme and a parameterization for non-inductive charging processes, was used to simulate tropical convection. In the cloud-simulations, convective heating profiles associated with lightning (non-lightning) producing convection were associated with a more pronounced upper-level (low-level) heating peak and an increased (decreased) contribution by ice-processes to the total surface rainfall. Since the rainfall process and lightning production become increasingly more correlated as contributions from the ice-phase to the total rainfall increase, we investigated the correlation between rainfall and lightning over large spatial and temporal scales for many different rainfall regimes. The results indicate that CG lightning flash density and rainfall are well correlated in warm-season rainfall regimes where highly electrified convection is prolific. In certain situations, it may be possible to use CG-lightning flash density to diagnose warm-season monthly rainfall totals, or differentiate between rainfall regimes.

Walter A. Petersen
Department of Atmospheric Science
Colorado State University
Fort Collins CO. 80523
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CHAPTER 1

INTRODUCTION

The processes leading to cloud electrification and lightning span temporal and spatial scales separated by many orders of magnitude. These scales encompass dimensions from that of individual electrical charges to that of convective clouds and/or deep convective cloud systems. In turn, the development of deep convective cloud systems is modulated to a large extent by atmospheric processes occurring over scales much larger than that associated with an individual cumulonimbus cloud. This “multi-scale” interaction can be viewed as a system (Fig. 1.1) comprised of a) “large-scale” (outer box; Fig. 1.1) thermodynamic and dynamical “inputs” such as convective available potential energy (CAPE) and shear, which enable the formation and maintenance of deep convective clouds; 2) cloud-scale “processing” (inner box; Fig. 1.1), associated with the conversion of large scale energy inputs to kinetic, potential, thermodynamic and electrical energies of the cumulonimbus clouds and manifested in the form of vertical drafts, significant quantities of precipitation, and often lightning; and 3) an “output” stage associated with transports of water (e.g., rainfall, water vapor), electrical charge (e.g., lightning, conduction currents), heating (e.g., latent and sensible heat transports) and momentum from the convection to the environment. The primary question addressed by this research is: *What do cloud-electricity observations, in particular observations of lightning in the tropics, tell us about processes occurring over a range of scales in the “system” (i.e., large-scales to cloud-scales in*

the tropical atmosphere; O[1000-10 km])? Addressing this question is timely and particularly relevant to “Global-Change” research (e.g., Goodman and Christian, 1993), given that lightning observation is rapidly approaching a global scale with the advent of combined spaceborne (e.g., the TRMM¹ Lightning Imaging Sensor [LIS], and the Optical Transient Detector [OTD]; Christian et al., 1996) and ground based lightning detection networks (e.g., VLF sferics networks; Lee et al., 1986; National Lightning Detection Network, Cummins, 1996). Prior to addressing this question, it is instructive to consider some of the characteristics associated with each component of the “system” shown in Fig. 1.1.

1.1. Large-scale thermodynamic “inputs” as related to cloud-electrification

Large-scale² inputs to an electrified cloud “system” can be subdivided into two primary categories; thermodynamic and dynamical inputs. Thermodynamic inputs include processes that increase or decrease the vertical distribution and/or magnitude of Convective Available Potential Energy (CAPE) of the environment (the amount of thermodynamic energy available for conversion to kinetic energy in a cloud updraft). The CAPE, for a hypothetical air parcel in the sub-cloud layer, is defined as the column integrated thermal buoyancy acquired by the parcel when the parcel is displaced adiabatically from its level of free convection (LFC) to its equilibrium level (EL). If a pseudoadiabatic process is assumed (all condensed water immediately exits the parcel during ascent), the CAPE can be defined as:

$$CAPE = \int_{p_{EL}}^{p_{LFC}} R_d (T_{vp} - T_{ve}) d \ln p \quad (1.1)$$

¹ TRMM (Tropical Rainfall Measuring Mission; Simpson et al., 1988)

² Here the term “large-scale” will be applied rather loosely to imply spatial or temporal scales much larger than that associated with individual cloud systems.

where p = pressure; p_{LFC} = pressure at the LFC, p_{EL} = pressure at the equilibrium level; R_d = gas constant for dry air, T_p = virtual temperature of the parcel; T_e = virtual temperature of the environment. If the integrand in Eq. (1.1) is integrated from the parcel's level of origin to the LFC, the value thus obtained represents a type of parcel energy deficit referred to as the Convective Inhibition (CIN). The CIN is defined as the energy required to accelerate the parcel through the LFC, thereby enabling the parcel to realize its CAPE.

Thermal buoyancy drives both dry and moist convection. Therefore, one might expect the value of the CAPE ($\propto w^2$) to be correlated to the updraft magnitude and intensity of convection. However, a correspondence between CAPE and updraft/convective intensity is often not observed, contrary to what parcel theory implies. This is because 1) CIN in the environment (e.g., a "cap" or inversion in the sounding) is sometimes too large for the parcel to overcome, no matter what triggering mechanism is employed, hence high CAPE values often occur with little or no convection; 2) when convection does occur, diabatic processes (e.g., heat and moisture exchanges during entrainment) modify the parcel as it ascends, rendering the parcel ascent non-adiabatic; and 3) water loading reduces total updraft buoyancy and hence the updraft speed; the assumption that all of the condensed water falls out of the updraft is also not perfect.

Though parcel theory is an imperfect representation of the thermodynamic process leading to moist convection, positive correlation's between the CAPE and convective intensity in the tropics *have* been noted in previous observational studies. Indeed, Williams et al. (1992) and Rutledge et al. (1992) demonstrated that variations in the CAPE exerted a strong influence on the vertical development and electrical intensity (using total lightning flash rates; Fig. 1.2) of tropical continental convection observed over the maritime continent region of tropical northern Australia during the DUNDEE³. The observations presented by Rutledge et al. and Williams et al.

³ DUNDEE (Down Under Doppler and Electricity Experiment; Rutledge et al., 1992)

suggested that during periods dominated by tropical continental “break” period convection and associated high CAPE, intense, highly electrified convection occurred. This convection was denoted by vertically oriented radar reflectivity contours, often with the 40 dBZ contour extending well above the -10°C level in the cloud (e.g., Fig. 1.3a). Conversely, during passage of the monsoon trough (associated with a moist oceanic-like airmass), CAPE was relatively low and the convection, though often deep (Fig. 1.3b), exhibited much weaker vertical development (e.g., considerably lower radar reflectivity above the freezing level) and produced little if any lightning. Given the marked contrast in electrical activity observed as a function of CAPE, the relationship between CAPE, convective intensity and cloud-electrification, was further explored by Rutledge et al. (1992) using scaling relationships between appropriate variables.

The scaling arguments employed by Rutledge et al. assume that particle-scale charge separation occurs predominantly in the mixed phase region (0°C to -40°C) of convective clouds via collisions between precipitation sized ice particles (diameter $O[1\text{ mm}]$) and much smaller cloud ice particles. This assumption is based on laboratory results which suggest that in collisions between riming graupel particles and small ice crystals, negative charge is transferred to the graupel, and positive charge to the smaller ice-crystals (e.g., non-inductive charging [NIC]; Reynolds et al., 1957; Takahashi, 1978; Williams and Lhermitte, 1983; Williams 1985; Saunders et al., 1991). Following the particle-scale charge separation, gravitational sedimentation of the negatively charged graupel and coincident ascent of the positively charged cloud ice in the updraft provides cloud-scale separation of electrical charge (Fig. 1.4). Further, it was assumed that charge separation occurred near a particle “balance” level (level at which updraft velocity w is equal to terminal fall speed of the ice particles; V_t), where negative charge would accumulate on precipitation-sized ice particles (e.g., Williams and Lhermitte, 1983). In this scenario, larger convective updrafts (supported by a higher CAPE) provide for more

precipitation ice mass via accretional growth (and hence more charging) in the mixed phase region of the convective cloud.

Hence, the scaling relationship between CAPE and cloud-electrical intensity proposed by Rutledge et al. is based on the following reasoning: At the particle balance level $w=V_t$, also $CAPE \propto w^2$ and $V_t \propto D^{0.5}$ (D =diameter; for precipitation sized particles falling at their terminal fallspeed), hence $CAPE \propto D$. Therefore, the particle mass (M) $\propto CAPE^3$. This scaling argument can be further extended to include trends in radar reflectivity (Z), since in a Rayleigh scattering regime, $Z \propto D^6$ and therefore $Z \propto CAPE^6$. Therefore, small changes in the value of CAPE are likely to be associated with large changes in the precipitation ice mass, radar reflectivity, and more efficient non-inductive charging of the cloud. Given the sensitivities of updraft strength, reflectivity, precipitation-ice mass and electrification to small changes in CAPE, it is important to consider processes that can alter the CAPE.

CAPE can be modified by altering the moist entropy of sub-cloud layer parcels (e.g., fluxes of latent or sensible heat) or by altering the virtual temperature profile of the atmospheric column with respect to a parcel (e.g., adiabatic cooling of the column through forced lifting, diurnal cycle of radiative flux divergence etc.). Xu and Emanuel (1989), Williams and Renno (1991), and Mapes and Houze (1993) demonstrated that CAPE in the tropical atmosphere is a sensitive function of surface thermodynamic variables such as wet-bulb potential temperature (θ_w), or alternatively equivalent potential temperature (θ_e). For a variety of tropical locations, Williams and Renno determined that an increase of only 1°C in θ_w (assumed to be a result of increased surface temperature) produced increases in CAPE on the order of 1000 J kg⁻¹. Mapes and Houze found a similar relationship between CAPE and surface temperature. Additionally, Mapes and Houze demonstrated that CAPE is extremely sensitive to changes in moisture content, with a 6% change in surface relative humidity resulting in a 1000 J kg⁻¹ increase in CAPE.

The sensitivity of CAPE and/or convection to variables such as θ_w is also manifested in observations of lightning flash rate over monthly, seasonal and annual cycles. Williams (1992, 1994, 1996) used monthly observations of θ_w and total lightning flash counts from different tropical and mid-latitude *continental* locations around the globe to demonstrate a highly non-linear sensitivity of lightning activity to small changes in θ_w (and by proxy, CAPE). The sensitivity was largest for tropical locations (Fig. 1.5) where a 1°C increase in θ_w produced a 600% increase in the amount of lightning activity (100-200% for mid-latitudes). Additionally, Williams (1992, 1994) demonstrated that small perturbations ($O[1^\circ\text{C}]$) in the annual and semi-annual cycles of surface temperature in the tropics were well correlated to periodic changes in the Schuman Resonance (Extremely Low Frequency [ELF] electromagnetic resonance in the earth-ionospheric waveguide associated primarily with lightning currents; Schuman, 1952). Based on the apparent sensitivity of lightning activity to small changes in θ_w , Williams (1992, 1994) suggested that total lightning flash density (or equivalently, the amplitude of the Schuman Resonance) could be used as a type of “global thermometer”.

Implicit in the hypothesis that the *magnitude* of the CAPE controls the intensity and/or amount of convection, is the assumption that magnitude of the CAPE can *vary*. This viewpoint is not universally held with regard to the variation of CAPE in the tropical atmosphere. Primary examples include the quasi (statistical)-equilibrium (QE) hypothesis originally advanced by Arakawa and Schubert (1974), and supported by studies such as Xu and Emanuel (1989) and Emanuel et al. (1994). This theory suggests that the amount of CAPE (calculated *reversibly*; i.e., condensed water is lifted with the parcel) in the tropics is approximately invariant. The QE hypothesis invokes observation and theory to suggest that cumulus convection consumes available potential energy on time-scales much shorter than that associated with the large-scale supply of potential energy (Arakawa and Schubert, 1974). Hence, the destruction of CAPE by

convection matches the instantaneous supply of CAPE (here we must emphasize the word “*available*” potential energy) provided by the large-scale environment. This hypothesis does *not* imply that there is *no* CAPE in the environment. Indeed, Renno and Ingersol (1996) suggest that CAPE on the order of 1000 J kg^{-1} must always be present in the environment in order for the atmospheric “heat-engine” to compensate the energy loss in frictional dissipation. It is simply the energy *made available* to cumulus convection by the large-scale circulation that is balanced by the occurrence of convection.

In support of the QE hypothesis, Xu and Emanuel (1989) presented results from sounding analyses which suggest that the *mean* tropical atmosphere in the western Pacific is approximately moist neutral for *reversibly* lifted sub-cloud layer parcels. The words “mean” and “reversible” are important to note for several reasons. First, the atmosphere does not truly operate within a framework of *reversible* thermodynamics. Reversible thermodynamics only provide a simplified theoretical framework to examine atmospheric thermodynamic processes. The presence of irreversible processes such as frictional dissipation, entrainment, or rain falling and evaporating into an updraft, all violate the “closed” nature of a “reversible” process. Second, and more importantly, rain falls to the surface of the earth, the condensate is not left suspended indefinitely in the upward moving parcel. Third, Xu and Emanuel emphasize a “mean” buoyancy of $\sim 1^\circ \text{ C}$ for the 2300 *soundings* examined in their study together with a “small” standard deviation in the mean buoyancy profile ($\sim 1^\circ \text{ C}$) to support their hypothesis of *approximate* moist neutrality. Xu and Emanuel interpret this result as indicating a “*nearly*” moist neutral condition. However, the word “*nearly*” is a rather ill-defined term when considering the sensitivities of CAPE to boundary layer θ_w , as discussed above.

It seems reasonable to assume that QE theory is valid over long time-scales (e.g., annually; consistent with Xu and Emanuel). However, observations suggest that this is *not* the case in the

tropics when shorter time-scales are considered (i.e., hourly, daily, weekly). Indeed, many previous observational studies of CAPE, convection, *and cloud-electrification* suggest that CAPE is quite variable in the tropics, and that lightning production and the amount/intensity of convection do vary as a function of CAPE (e.g., Rutledge et al. 1992; Williams et al. 1992; Williams, 1992; Williams and Renno, 1993; Jayaratne 1993; Mapes and Houze, 1993; and Williams, 1994). A similar sensitivity between CAPE, tropical convective development and electrical intensity was demonstrated in the modeling studies of Randell et al. (1994). The conclusions of these observational and modeling studies strongly suggest that *variations in the amount of CAPE exert a strong control on local convective development and subsequent electrification.*

While we have considered sensitivities between convection and the *amount* of CAPE present, it is also important to consider changes in the *vertical distribution* of the CAPE. Lucas et al., (1994) illustrated this point by examining the vertical profiles of thermal buoyancy for tropical oceanic soundings (taken near Darwin, Australia), and for a mid-latitude continental sounding (Fig. 1.6). The importance of the “shape of the CAPE”, as pointed out by Lucas et al. (1994), is that profiles of thermal buoyancy over the warm tropical ocean, while often containing a significant amount of CAPE, contain relatively smaller thermal buoyancy’s distributed over a greater column depth when compared to their mid-latitude counterparts. This enables entrainment and water loading in tropical oceanic convection to more easily counteract the positive thermal buoyancy resulting in a reduced updraft strength (e.g., LeMone and Zipser, 1980; Jorgensen and Lemone, 1989; Lucas et al. 1994; Cifelli, 1995), the amount of precipitation ice mass aloft, and hence the electrical intensity (Williams and Lhermitte, 1983; Rutledge et al., 1992; Williams et al., 1992; Zipser and Lutz, 1994). This suggestion is consistent with satellite observations of lightning (e.g., Turman and Edgar, 1982, Orville and

Henderson, 1986) which indicate that approximately an order of magnitude less lightning occurs over the oceans relative to that over land (Fig. 1.7). *One objective of this study will be to assess the relative sensitivity of convective development and electrical intensity to small changes in both the magnitude and the vertical distribution of CAPE over the tropical western Pacific Ocean.*

1.2. Large-scale dynamical inputs as related to cloud electrical intensity

Large-scale dynamical forcing of deep convection in the tropics is primarily associated with the propagation of wave disturbances that induce and/or enhance upward motion, mass/moisture convergence (linked to rainfall through the water budget), wind shear and surface sensible and latent heat fluxes (and by doing so, increase or decrease the CAPE). Over the western Pacific propagating wave disturbances often accompanied by deep convection, heavy rainfall, and well defined deviations in surface/upper-air wind, moisture and temperature fields, have been observed to occur with characteristic periods of approximately 1-5 days, 10-20 days, and 30-60 days (e.g., Madden and Julian, 1971, Reed and Recker, 1971; Knutzen and Weickman, 1987; Lau et al., 1991; Numaguti et al., 1995; Zhang, 1996; Lin and Johnson, 1996a). Similarly, deep convection over the tropical eastern Atlantic Ocean has been observed to propagate in phase with 3-5 day easterly waves that move off the African continent (e.g., Thompson, 1979).

We might assume *a priori*, that just as Outgoing Longwave Radiation (OLR) decreases in association with deep convection embedded in the disturbance (e.g., Knutzen and Weickman, 1987; Lau et al. 1991), the amount of electrical activity should increase. *Within* a given regime (e.g., tropical continental or oceanic), increased electrical activity (higher flash rates) is likely to

occur due to any imposed thermodynamic forcing that increases the CAPE which, in turn, must be balanced by increases in convection (if no storage of CAPE occurs over several day timescales). Dynamical forcing such as wind shear is supposed to enhance the level of organization and the vertical structure of the convection (e.g., Rotunno et al., 1988, Trier and Parsons, 1995, Trier et al., 1996). Hence, we expect to see increases in electrical activity with larger, more organized regions of convection.

The word “*within*”, when referenced to “regime” and the thermodynamic forcing of the wave is emphasized because the opposite effect (e.g., a relative reduction in electrical activity and convective intensity) can occur when the arrival of a disturbance causes a marked transition in the air mass characteristics. Specifically, the periodic occurrence of the Australian monsoon over the tropical continental region of northern Australia is associated with a factor of 2-10 *decrease* in lightning frequency but a marked *increase* in convection and rainfall (e.g., Hendon and Liebman, 1990; Williams et al. 1992). Hence, *within* specified regimes (e.g., airmass types; cf., Trewartha and Horn, 1980), to the extent that lightning frequency behaves temporally and spatially like the distribution of deep convection in a wave disturbance *it may be possible to use lightning observations to infer the most intense convective phase of a propagating disturbance.*

1.3 Diurnal cycles

Over most continental locations it has been demonstrated that the diurnal cycle of lightning and convective rainfall typically peak in the late afternoon approximately coincident with peaks in the solar radiation and boundary layer heating [and hence CAPE] (e.g., Lopez and Holle, 1986; Reap and MacGorman, 1989; Williams and Heckman, 1992; Zipser, 1994; Keenan et al., 1994; Jayaratne and Ramachandran, 1996; Finke and Hauf, 1996). The effect is even more

pronounced if the diurnal cycle of solar heating in the region being considered drives a solenoidal circulation such as a sea-breeze (Maier et al., 1984; Watson et al., 1991; Lopez et al., 1991) or mountain-valley type circulation (e.g., Reap, 1986). Over remote regions of the tropical ocean, the diurnal cycle of lightning is a relative unknown, *but this study will present new results from TOGA-COARE⁴ that suggest the presence of a very pronounced nocturnal maximum in cloud-to-ground lightning activity.*

Importantly, while few observations of the diurnal cycle of lightning over the tropical oceans exist in the literature, several studies *have* examined the diurnal cycle of rainfall and/or infrared (IR) brightness temperature ("cold-cloud area"; often used as a proxy for the presence of deep convection) over the tropical oceans (e.g., Holle, 1968; Gray and Jacobson, 1976; Janowiak et al. 1994, 1991; Mapes and Houze, 1993; Chen and Houze, 1996; Rickenbach, 1996; and others). The majority of these studies have suggested that the diurnal cycle of rainfall is characterized by a weak nocturnal or early morning maximum over the tropical ocean. The precise mechanism by which the nocturnal maximum in rainfall (and presumably deep convection) occurs is uncertain. Gray and Jacobson (1976) suggested that radiational cooling in cloud-free regions and the subsidence generated to offset the cooling, provide an enhanced low-level convergent flow into cloudy regions. This enhanced convergence drives stronger ascent in the cloudy region and an associated increase in surface rainfall. Chen and Houze (1996) explain a nocturnal maximum in cold cloudiness over the western Pacific as a result of the temporal evolution of large (the amplitude of the diurnal cycle was dependent on the size of the cold cloud area) mesoscale convective systems (MCS's) which are initiated as isolated convection in the late afternoon but increase in size and degree of organization throughout the night. The finding of a size dependence in the diurnal cycle of cold cloudiness by Chen and Houze (1996) seems

⁴ TOGA COARE (Tropical Ocean Global Atmosphere Coupled Ocean Atmosphere Response Experiment; Webster and Lukas, 1992)

consistent with both the “radiational cooling-enhanced convergence” hypothesis set forward by Gray and Jacobson (1976), and also with the hypothesized “gregarious” nature of deep convection associated with large tropical cloud clusters (Mapes, 1993). However, more recent radar studies (e.g., Rickenbach, 1996) of the diurnal cycle of precipitation over the western Pacific warm-pool do not support the conclusions of Chen and Houze (1996).

As an alternative view, Rickenbach (1996) demonstrated that the diurnal cycle in rainfall can be explained as the sum of various harmonics in rainfall associated with different types of convective organization that occur throughout the day (i.e., with no particular preference for *when* the particular type of organization occurred, in contrast to Chen and Houze, 1996). When the rainfall amounts were stratified by the type of convective organization, Rickenbach noted that the maximum variance in rainfall explained by the diurnal cycle was associated with the occurrence of large, non-linear MCS's. In the event that these systems developed at night, Rickenbach speculated that a preferential enhancement of stratiform precipitation would occur as radiational flux-divergences and associated cooling at cloud top further destabilized the stratiform cloud, strengthening the overturning circulation, and enhancing the rainfall. Rickenbach also pointed out that the kinematic and microphysical evolution of MCS stratiform regions is tightly coupled to the kinematics and microphysics of the parent convection. Hence any preference for relatively deeper convective activity at night in MCS's would tend to enhance the observed nocturnal maximum of rainfall observed in large MCS's. This suggestion and prior discussion suggests that *any* mechanism which enhances the vigor of nocturnal convection and rainfall over the tropical ocean (e.g., Gray and Jacobson, 1976; Chen and Houze, 1996) should also reveal itself in the diurnal signal of lightning frequency. *This study will present evidence that deep convection is nocturnally enhanced based on observations of the diurnal cycle of cloud-to-ground lightning, CAPE, and convective rainfall over the western Pacific warm-pool.*

1.4 Convective processing and electrical intensity

Latent heat release in heavily raining cumulonimbus "hot towers" (Riehl and Malkus, 1958) is the primary means by which excess radiative energy accumulated in the tropics is "*processed*" by deep convective clouds for redistribution poleward by the upward branch of the Hadley circulation. In this frame of reference the word "hot" refers to the tremendous amount of *latent heat* energy released in the middle and upper troposphere of deep tropical convective clouds, and "towers" refers to the great depth of the clouds (typically 15-20 km). Alternatively, Rutledge et al. (1992) and Williams et al. (1992) used the term "hot tower" to describe deep, *electrically* "hot" tropical continental cumulonimbi that produce significant amounts of lightning and rainfall over the maritime continent of northern Australia.

Studies such as Chauzy et al. (1985), Rutledge et al. (1992), and Williams et al. (1992) have related differences in tropical cloud electrical properties (electrification/lightning; Fig. 1.1) to observed *dynamical and microphysical* differences between lightning and non-lightning producing clouds. These "*processing*" differences are characteristically *internal* to the clouds and are related to the updraft strength (*dynamics*; Fig. 1.1) and the amount and type of hydrometeor mass present in the mixed phase region (0°C to -40°C) of the clouds (*microphysics*; Fig. 1.1). Due to the coupling between cloud-electrification processes, cloud microphysics and cloud dynamics, it is reasonable to suggest that cloud-electricity observations should provide important insights into the internal processes governing the lifecycle of electrified cumulonimbus clouds. Conversely, an ensemble of deep cumulonimbi (CB's) *not* producing copious amounts of lightning (e.g., typical oceanic or monsoon convection) can also provide important clues as to the dominant internal structure of clouds in the ensemble. To fully appreciate the coupling between

electrification, dynamics and microphysics implied by Fig. 1.1, an understanding of current hypothesized cloud-electrification processes is beneficial.

1.4.1 A brief review of cloud electrification hypotheses

To a first order, observations of electrified cumulonimbus clouds suggest that thunderstorms can be represented electrostatically as dipolar or tripolar distributions of charge (e.g. Williams, 1989). In the tripole model of a thunderstorm (Fig. 1.8) there is a concentrated layer of negative charge situated between the -10°C and -25°C temperature levels, with an upper layer of positive charge at temperatures colder than -25°C and a lower region of positive charge (often weaker in magnitude) situated between the 0°C and -10°C levels. Each of these layers is associated with a charge density on the order of $1\text{--}5\text{ nC m}^{-3}$. Balloon-borne observations (e.g., Marshall and Rust, 1991; Marshall et al., 1995a; Stolzenburg, 1996) suggest the frequent presence of more than three layers of charge. However, when the observations are averaged (Stolzenburg, 1996), the overall charge structure still resembles that of a thunderstorm tripole. Therefore, the gross electrical structures of cumulonimbi will be discussed within the framework of a tripole model, recognizing that significant deviations from the model do occur.

Theories offered to explain the development of thunderstorm tripoles (*electrification*; Fig. 1.1) can be partitioned into two distinct groups: precipitation and convectively based theories. The latter theory, proposed by Grenet (1947) and Vonnegut (1954), invokes a dynamical explanation for charge transfer in thunderstorms. In a convective charging process, convective updrafts lift positive space charge resident in the sub-cloud layer into the cloud and the charge is subsequently attached to cloud-particles while moving toward the cloud-top. The resulting positive electric field perturbation near the top of the cloud causes a flow of negative charge

toward the cloud, with the negative charge subsequently becoming attached to cloud and precipitation particles on the cloud periphery. The negative charge then descends in the downdraft and is folded into the center of the cloud. In this fashion a dipolar charge structure is realized in the cloud. However, as pointed out in Williams (1989), the convective theory does not completely account for the observed charge structure in thunderstorms (e.g., the observed “layering” of charge; the apparent temperature dependence of the tripolar structure etc.), and does not supply sufficient electrical current over time-scales required to maintain observed lightning flash rates..

Precipitation based theories are subdivided into two categories; inductive charge transfer (electric field dependent), and non-inductive processes (discussed briefly in Sec. 1.1). Inductive mechanisms (e.g., Elster and Geitel, 1913; Aufdermaur and Johnson, 1972; Illingworth and Latham, 1977, Gaskell, 1981; Caranti and Illingworth, 1985; Mason, 1988; Brooks and Saunders, 1994) rely on field-driven charge transfer in collisions between ice particles (graupel/hail) and cloud-droplets that have been polarized in an external electric field. The polarizing electric field vector typically points downward forcing positive charge to accumulate on the underside of the particles. Therefore, in grazing collisions with cloud droplets, a net positive charge is transferred to the droplet and negative charge to the ice particle. This charge transfer, followed by sedimentation of the large ice particles, strengthens the electric field in the cloud providing a positive feedback (exponential) to the charging process. However, inductive mechanisms cannot account for the initial charge buildup in thunderstorms because the initial “fair-weather” electric fields ($O[100 \text{ V m}^{-1}]$) are too small to drive the appropriate charge transfer (e.g., Kuettner, 1981; Brooks and Saunders, 1994). Laboratory studies of Aufdermaur and Johnson (1972) and Mason (1988) suggest that the role of inductive mechanisms *may* be important at electric field strengths characteristic of lightning-producing clouds ($O[100 \text{ kV m}^{-1}]$),

but this is highly uncertain due to the unknown nature of graupel-cloudwater collision interactions (e.g., Brooks and Saunders, 1994). Relegation of the inductive mechanism to a secondary process brings us to the non-inductive charging (NIC) mechanism; the mechanism most consistent with observations of the charge structure in electrified clouds.

The NIC mechanism does not require a pre-existing electric field for charge transfer to take place. Laboratory tests of the NIC mechanism confirm that significant charge transfer ($O[10-100 \text{ fC}]$) occurs during collisions between riming graupel “targets” ($\sim 3-5 \text{ mm}$ in diameter) and smaller ice crystals ($\sim 100 \mu\text{m}$ in diameter) in the presence of supercooled liquid water, at temperatures between -5°C and -35°C (e.g., Reynolds et al., 1957, Takahashi, 1978; Jayaratne et al., 1983; Baker et al, 1987; Keith and Saunders, 1990; Caranti et al., 1991; Saunders et al., 1991; Brooks and Saunders, 1994). Moreover, both the sign and magnitude of the charge transfer are functions of the temperature and liquid water content of the cloud (e.g., Fig. 1.9), the relative collision velocity of the particles, and the diameters of the small ice crystals. At typical thunderstorm cloud-water contents of 1 g m^{-3} and temperatures of -15°C (Fig. 1.9) the graupel charge negatively and ice crystals charge positively. At slightly warmer temperatures (-5°C to -10°C) in the same liquid water content, the sign of charge acquired by the graupel reverses to positive (hence descending graupel would traverse a “level of charge reversal” [LCR]). At points where descending graupel charge negatively (positively), smaller ascending ice particles charge positively (negatively). The charge transfer as a function of temperature and liquid water content, coupled with the relative movement of the graupel and ice particles results in a central layer of negative charge between the -10° to -25°C levels that is sandwiched between upper and lower layers of positive charge consistent with the development of a thunderstorm tripole (e.g., Fig. 1.4).

The physical mechanism responsible for charge transfer at the particle interface is still not well understood. Prominent hypotheses advanced to explain the charge transfer include charge transfer due to rime contact potentials (Caranti and Illingworth, 1980; Saunders et al., 1991), charge separated during rime fracturing (e.g., Caranti et al., 1991; Saunders et al., 1993; Avila and Caranti, 1994); removal of charged dislocations in the rimer surface (e.g., Takahashi, 1978; Saunders et al., 1991); and transfer of charge (and mass) in a chemical potential gradient associated with the presence of quasi liquid-like layers of differing thickness at the collision interface (Baker et al., 1987; Baker and Dash, 1994). To some extent, these mechanisms all depend on the relative surface state (e.g., ice surface growing or sublimating in a field of water vapor; e.g., Williams et al., 1991) of the ice particles and are virtually impossible to observe in-situ. This makes it difficult to verify any or all of the laboratory based hypotheses for charge transfer on the *particle scale* in real clouds.

Regardless of the precise nature of the charge transfer on the molecular level, *thunderstorm electrification requires precipitation sized and smaller cloud ice particles to be collocated in sufficient concentration, and in a mixed phase environment in order for significant charge transfer to progress on the cloud-scale.* Scaling arguments using continuous collection equations (e.g., Takahashi, 1978) suggest that concentrations of millimeter sized graupel and cloud ice should be on the order of 1 L^{-1} and $50\text{-}100 \text{ L}^{-1}$ respectively, to ensure that a sufficient number of particle collisions and charge transfer takes place over time-scales relevant to thunderstorm electrification.

1.4.2 Cloud processing and electrification summarized

The discussion in Sec. 1.4.1 establishes an important link between cloud kinematics (e.g., draft structure), microphysics (i.e., the type and amount of condensate present) and electrification processes (e.g. the NIC mechanism). To a first order, observations indicate that the presence of large quantities of precipitation size ice are necessary before a cloud can become significantly electrified (e.g., Williams and Lhermitte, 1983; Dye et al., 1989; Goodman et al., 1988; Jameson et al., 1996; Ramachandran et al., 1996). In order for convection to initiate an active ice process in which precipitation sized ice particles are produced in a mixed phase environment (*microphysics*, Fig. 1.1), a significantly strong updraft must first exist (*cloud-scale dynamics*, Fig. 1.1). When these requirements are met, the cumulonimbus cloud will electrify resulting in the production of lightning. *In this scenario, the presence of lightning then becomes a proxy for inferring the presence of other processes (e.g., the riming growth of large ice particles, strong updrafts at mid-levels of the cloud, elevated latent heating peaks, etc.) that are internal to the cloud.*

On the other hand, deep cumulonimbus clouds in the tropics (e.g., over the ocean) are often not associated with lightning production, even though heavy rainfall and substantial convective heating takes place. This is because the rainfall and convective heating processes (in contrast to the electrification process) do not require the presence of precipitation sized ice. In the case of tropical oceanic convection, a combination of reduced updraft speed (*dynamics*; Fig. 1.1) due to the vertical profile of thermal buoyancy, water loading, and entrainment (e.g., LeMone and Zipser, 1980; Jorgensen and LeMone, 1989; Williams and Renno, 1993; Lucas et al. 1994; Williams, 1995, DeMott and Rutledge, 1996), combined with lower cloud condensation nuclei concentrations, increased depth of the cloud below the freezing level, and higher moisture contents in the lower tropical atmosphere, combine to enable warm-rain microphysics (*microphysics*, Fig. 1.1) to dominate the rainfall production of the cloud (Young, 1993). To

some extent the formation of rain in the lower regions of a tropical cloud “short circuits” the generation of large quantities of precipitation ice above the freezing level by depleting the weaker, water loaded updrafts of necessary cloud and rain water contents. Further, the development of copious numbers of small ice particles (crystals, snow and graupel; e.g. Houze and Churchill, 1984; Gamache, 1990) are favored in the weaker updraft at temperature levels between -3°C and -8°C due to secondary ice processes such as rime-splintering (e.g., Hallett and Mossup, 1974). The presence of large concentrations of small ice near the -5°C level in the cloud essentially “overseeds” (e.g., Gamache, 1990) this region of the cloud, and through the Bergeron process, any small amounts of liquid water present in the updraft are quickly removed. Hence, the possibility of generating a sufficient concentration of large graupel above the -10°C level (necessary to electrify the cloud through NIC processes) is significantly reduced in the glaciated cloud.

Using cloud and lightning observations collected in tropical northern Australia, Zipser and Lutz (1994) speculated that in order for tropical oceanic clouds to produce lightning, the cloud updraft must be capable of lofting significant concentrations of millimeter sized liquid and frozen raindrops above the height of the -10°C level in order to provide sufficient precipitation ice mass for an efficient electrification process to take place. This hypothesis is consistent with more recent polarimetric-radar observations of continental cumulonimbi over sub-tropical Florida that suggest cloud-electrical processes rapidly intensify as upward moving raindrops begin to freeze near the -10°C level (e.g., Ramachandran et al., 1996; Jameson et al. 1996; French et al., 1996). Based on calculations of the terminal fall speed of liquid and frozen raindrops, Zipser and Lutz (1994) estimated that “threshold” updraft speed of $\sim 6\text{ m s}^{-1}$ would need to be exceeded at or near the -5°C level before lofting of frozen drops and subsequent strong electrification of the cloud could occur. In an oceanic convective cloud a 6 m s^{-1} updraft strength would fall into the upper

10% of all aircraft-observed updraft magnitudes in tropical convection (LeMone and Zipser, 1980; Jorgensen and LeMone, 1989; Lucas et al., 1994). The hypothesized “updraft threshold” of Zipser and Lutz and the microphysical relationship to electrification of the cloud are consistent with the “particle balance level” scaling arguments summarized in Sec. 1.1 (e.g., Rutledge et al., 1992; Williams et al., 1992). However, the updraft “threshold” hypothesis and the role of frozen drops in relation to electrification of deep convection over isolated regions of the tropical oceans has yet to be tested rigorously by observational or modeling studies.

This study will present new cloud-scale observations of both the electrical and kinematic properties of tropical oceanic convection using observations from TOGA COARE. The observations will be considered in light of current hypotheses advanced to explain cloud-electrification and will also be contrasted to similar observations made of deep continental convection (tropical and mid-latitude). Additionally, modeling studies of the dynamical, microphysical and electrical properties associated with a spectrum of deep tropical clouds will be undertaken to further examine the nature of the electrification process in tropical convection.

1.5 Cloud-scale “Outputs” and the Relationship to Observed Electrical Properties

The previous discussion has demonstrated that cloud-scale kinematics are inseparably linked to cloud microphysical and electrical properties. The manifestation of these interactions on the cloud-scale comprise observable outputs associated with feedbacks to larger-scales. One visible manifestation of cloud-scale processes is the production of lightning. When integrated over the surface of the earth, electrified convection (e.g., thunderstorms) is an important supplier of electric current to the global electrical circuit (e.g., Wilson, 1920; Whippel and Scrase, 1937)

and a processing step in the generation and redistribution of chemical species such as NO_x in the earth's atmosphere (e.g., Tuck, 1976). The global-electrical circuit and production of chemical species by lightning are not within the scope of this study. However, *meteorological "outputs" such as diabatic heating/moistening of the atmosphere, the production of rainfall, and the relationship of these variables to cloud electrical properties will be examined in detail.*

1.5.1 Convective Heating Profiles as Related to Cloud Electrification

Cloud kinematics and microphysics are directly related to the vertical distribution of the diabatic heating and moistening produced by the cloud-scale circulation through condensational and eddy transport terms in the budgets for heat and moisture (e.g., Yanai et al., 1973; Houze, 1982; Johnson, 1984; Yanai and Johnson, 1995; Lin and Johnson, 1996b). Vertical profiles of total heating (Q_1 , the "apparent" heat source; Fig. 1.10) associated with deep convective cloud systems have been composited using sounding network data collected in several locations of the tropics (e.g., Houze, 1982; Johnson, 1984; Frank and McBride, 1989; Lin and Johnson, 1996b). The observed total heating profiles are a convolution of both convective and mesoscale heating components (e.g., Houze, 1982; Johnson, 1984). When partitioned by rainfall fraction, as in Fig. 1.10a, the more variable convective heating profile (Fig. 1.10a) is characterized by a lower level heating peak with maximum heating rates situated between 500-700 mb (e.g., Johnson, 1984; Houze, 1989). The mesoscale heating profile is associated with an elevated heating peak situated near 350 mb and cooling below the 500 mb level. Vertical profiles of the apparent moisture sink associated with cloud systems (Q_2 ; Fig. 1.10 b) are also composed of convective and mesoscale components. The convective profile (Fig. 1.10b) exhibits a low level drying peak (~800 mb) and the mesoscale profile exhibits an upper level drying peak (~400 mb) and low level moistening

peak (~800 mb). The peaks in the mesoscale profiles are associated with condensation in the mesoscale updraft and evaporation in the mesoscale downdraft.

Accurate representation of diabatic heating profiles is essential for regional, synoptic and general circulation models (e.g., Puri, 1987) because mass divergence in dynamical models is coupled to the slope of the heating profile through the First Law of Thermodynamics. For example, modeling studies of tropical wave disturbances such as Lau and Peng (1987) and Sui and Lau (1989) suggest that the propagation speed of intra-seasonal oscillations such as the MJO and associated shorter period westward propagating Rossby waves are modulated by the vertical structure of the diabatic heating associated with deep convection (a “mobile-wave CISK” mechanism). In particular, the results of Sui and Lau (1989) suggest that slower moving waves will occur with stronger low level diabatic heating peaks (i.e., convective peaks centered near 700 mb), whereas faster moving waves will be associated with a more intense upper-level heating peak (centered at or above 500 mb).

Given that electrical intensity is related to the vertical development of convection, it is reasonable to expect a physical relationship between lightning frequency and the shape of the heating and moistening profiles. If this is the case, lightning observations (especially spaceborne observations that provide global coverage) might be a powerful tool to use in the diagnosis of *convective* heating profiles (which are more variable than the mesoscale component; e.g., Houze, 1989). For example, in contrast to the more “typical” non lightning-producing convection observed over the ocean, we might expect to find an elevated convective heating peak in regions of deep, highly electrified convection associated with a more robust ice-process situated in the mid-levels of the cloud. Hence, we ask the following question: *To what degree can lightning flash rates be used to formulate quantitative statements about the kinematic and microphysical structure of the cloud, including convective heating and moistening profiles?*

1.5.2 Lightning-Rainfall Relationships

In addition to lightning, an additional obvious output of deep convection is rainfall. Physically, it would be reasonable to expect lightning and rainfall to be somewhat correlated in deep convective rainfall regimes (e.g., warm seasons over continental locations; tropical continental locations during monsoon break periods etc.) as both rainfall and lightning production are functions of convective microphysics and kinematics. Many previous studies (Table 1.1) have attempted to quantify the relationship between rainflux and either cloud-to-ground lightning (CG) or total lightning flash density by formulating a ratio of rain-mass to the number of lightning flashes (here defined the “rain-yield”; kg/flash) observed over a given time period for individual or ensemble thunderstorm case studies (e.g., Parczewski, 1958; Israel and Ries, 1964; Battan, 1965; Kinzer, 1974; Maier et al. 1978; Piepgrass and Krider, 1982; Reap and MacGorman, 1989; Buechler et al. 1990; Goodman and Buechler, 1990; Nielsen et al. 1990, Lopez et al., 1991; Williams et al., 1992; Richard and Lojou, 1996; Cheze and Sauvageot, 1997). While the rain-yield values computed in each study vary by an order of magnitude or more, and are a function of the rainfall regime, the majority of the studies find high correlations between *convective* rainfall and lightning (for both total and CG flash densities; Table 1.1).

Recently, Zipser (1994) compared thunderdays to rainfall (expressed as a ratio of rainfall to thunderdays, the “rain-to-thunderday ratio”; RTR) over many different areas of the globe, including the tropics. To some extent, Zipser’s results (Table 1.1) demonstrate a negative correlation between thunderdays and rainfall, especially in the tropics. For example, thunderday

Table 1.1: Previous studies of rainfall-lightning relationships

Study	Result
Parczewski, 1958 (via Israel, 1973)	In 8 out of 11 cases examined, increases in lightning frequency were accompanied by increases in precipitation amount.
Israel and Ries, 1964 (via Israel, 1973)	A weak relationship between lightning frequency and rainfall amount (measured by dam outflow) was found.
Battan (1965)	Observations of 29 thunderstorms in southern Arizona found that rainfall (gauge est.) and CG lightning (visual counts) were well correlated. A mean rain-yield of $3-6 \times 10^7$ kg/fl was computed.
Kinzer, 1974	Rainfall (radar est.) and CG lightning were positively correlated. The ensemble mean rain-yield was computed to be 1.6×10^7 kg/fl
Maier et al., 1978	An ensemble of cases were studied in Florida. Rain-yields of 10^7-10^8 kg/fl were diagnosed.
Piepglass and Krider, 1982	An ensemble of cases exhibited a high correlation between rainfall (gauge est.) and total lightning. Two cases were used for calculating a rain-yield value of $\sim 2 \times 10^7$ kg/fl.
Reap and MacGorman, 1989	A good correspondence was found between radar echo intensity and CG lightning frequency for ensemble of cases examined over Oklahoma.
Buechler et al., 1990	Peak rainflux (radar est.) and CG flashes were well correlated, total rain volume and CG flashes were weakly correlated. The rain-yield varied from 10^7-10^8 kg/fl.
Goodman and Buechler, 1990	Total and CG lightning were well correlated to convective rain volumes (radar est.). Rain-yields varied from 10^7-10^8 kg/fl.
Nielsen et al., 1990	CG lightning and rainflux (radar est.) were well correlated ($r=0.85$). Rain-yields varied from 10^7-10^8 kg/fl. A likely regime dependence of rain-yield was noted.
Lopez et al., 1991	Hourly mean rain-yields near Cape Canaveral, Florida varied as a function of convective regime from 10^8 kg/fl in offshore flow to 10^{10} kg/fl in onshore flow. (radar est.)
Williams et al., 1992	Tropical Maritime Continent. Daily rain-yields (radar est.) varied from 10^8 kg/fl to 10^{10} kg/fl depending on whether Darwin was in a break or monsoon rainfall regime respectively.
Zipser, 1994	Rain-to-thunderday (RTR) ratios were used to demonstrate the regime dependence of RTR over many global locations. Low RTR's were found over continents in mid-latitudes or during continental rainfall regimes; high RTR's were found for oceanic/monsoon environments.
Richard and Lojou, 1996	Investigated rainfall (radar est.)-lightning relationships for flash flood forecasting in France. 60 storms were studied. Good correlations were found between rain-volume and total lightning frequency.
Cheze and Sauvageot, 1997	Significant correlations were found on a case-by-case basis between radar-estimated convective rainfall and lightning. Correlations between CG-lightning and rainfall were higher than correlations between total flash rate and rainfall.

and rainfall statistics compiled by Zipser suggest that the heavy rains associated with tropical monsoon and oceanic convection were generally associated with a much lower number of

thunderdays, or higher RTR's (due to a lower number of thunderdays but heavy convective rainfall). For areas and/or rainfall regimes (e.g., the tropical "break" period in northern Australia; Rutledge et al. 1992) associated with a higher number of thunderdays and/or less precipitation, the RTR was low.

Combining Zipser's thunderday observations and other observations from the tropics such as Williams et al. (1992), the following summary of Table 1 is offered:

- 1) Cloud-to-ground and total lightning flash rates are reasonably well correlated to convective rainfall.
- 2) Rain-yields vary by a factor of 10 or more for any given location, and a factor of 10^3 between rainfall regimes (10^7 kg/fl in the continental U.S. to 10^{10} kg/fl during monsoon rainfall over the tropical maritime continent of northern Australia; e.g., Nielsen et al., 1990; Lopez et al., 1991; Williams et al., 1992; Zipser, 1994).
- 3) A portion of the variation observed in the rain-yields can be attributed to differences in lightning/rainfall measurement technique (e.g., Goodman and Buechler, 1990). For example, Battan (1965) and Piepgrass and Krider (1982) used raingauge networks to estimate rain volume, whereas many of the other studies used radar reflectivity estimates of rainfall which can suffer from a multitude of errors.
- 4) Rain-yields are affected by natural convective scale variability (i.e., no two thunderstorms are exactly alike; e.g., Cheze and Sauvegeot, 1997).
- 5) On a case-by-case basis, rain-yields are affected by the amount of stratiform precipitation that falls in a mesoscale convective system (MCS) (e.g., Goodman and Buechler, 1990).

Considering (2) and the relative dominance of warm rain process in tropical oceanic convection (which is essentially decoupled from the electrical process), it would seem that any

attempt to relate lightning and rainfall over the tropical oceans might be futile. However, it is important to note that the correlation between *tropical oceanic lighting and rainfall has not been thoroughly tested within an oceanic or monsoon rainfall regime*. Previous studies such as Zipser (1994) and Williams et al. (1992) have not tested the rainfall-lightning relationship *within* given rainfall regimes. Further, it has been suggested in the literature that: 1) lightning can be used as a proxy for the intensity of deep convection over the ocean (Molinari et al. 1995); 2) the deepest 1-5% of oceanic convection over the tropical western Pacific Ocean produces greater than 40% of the total rainfall observed (Janowiak et al., 1994; Liu et al., 1995); and 3) the frequency of lightning *over the ocean* may be related to the amount of CAPE present (e.g., Rutledge et al. 1992; Williams et al. 1992). Therefore, to the extent that lightning varies coherently with large scale forcing (e.g., “CAPE” and “dynamics” in Fig. 1.1) and is associated with the deepest raining convective systems in oceanic or monsoon rainfall environments, lightning *may* be positively correlated to rainfall *within* tropical oceanic or monsoon rainfall regimes. That is, any supposed correlation between lightning and rainfall production suggested by Table 1.1 must be interpreted in the context of the rainfall regime (as a function of the dominant microphysical precipitation process) being examined.

The question we pose is; *how and where can the apparent correlation between total or CG lightning flash density and rainfall be exploited in a quantitative sense, using rain-yields that appear to be highly variable from storm to storm and location to location?* The answer to this question has particular relevance to ongoing and planned field programs in the tropics such as TRMM that will attempt to diagnose both tropical rainfall and the vertical structure of latent heating associated with deep tropical/sub-tropical convection using spaceborne observations (Simpson et al., 1988). As previously noted the TRMM satellite will carry a Lightning Imaging Sensor (LIS). The LIS will enable detection of total lightning flash rate on cloud-scales with a

better than 90% detection efficiency (personal communication, Dr. H. Christian, NASA). Thus a physically based method or approach by which lightning and convective rainfall could be objectively related (e.g., through a variable such as the rain-yield) could prove valuable to programs such as TRMM.

1.6 Summary of the objectives of the research

This research is concerned with meteorological interpretation of cloud-scale electricity observations in the tropics set in the framework of multi-scale interactions between deep convection and the environment (Fig. 1.1). Specific questions addressed by the research include:

- 1) What are the sensitivities of convective intensity and electrical development to small changes in both the magnitude and the vertical distribution of CAPE over the tropical western Pacific Ocean? What processes control variations in the CAPE?
- 2) Can lightning observations be used to infer and/or track the presence of large scale dynamical processes over the tropical ocean, such as wave disturbances?
- 3) What is the diurnal cycle of lightning over the tropical western Pacific warm-pool and how does it relate to other meteorological measures of convection such as rainfall, CAPE, and cold-cloud area?
- 4) What processes are associated with the electrification of tropical oceanic convective clouds over the warm-pool and how do these processes differ from those observed in continental convection? Microphysically, are frozen drops explicit or implicit participators in the electrification of tropical clouds? Are observations of vertical velocity, convective intensity and electrical development associated with tropical

oceanic clouds consistent with the “thresholds” in vertical velocity suggested by Zipser (1994)?

5) To what degree can electrical activity be used to make quantitative statements about the kinematic and microphysical structure of clouds, including convective heating and moistening profiles?

6). To what extent are lightning and rainfall correlated by rainfall regime?

These issues are addressed through analysis of data collected in TOGA COARE, the DUNDEE, the U.S., and French Guyana. To address some of the more intricate cloud-scale and microphysical questions (where observational data were lacking) modeling studies were also undertaken using the Ferrier and Houze 1-D, time-dependent cloud-model with a parameterized electrification scheme. Use of the model with specified sounding profiles permitted the testing of convective scale sensitivities to large scale thermodynamic forcing (e.g., CAPE and the distribution of thermal buoyancy).

Flow of the dissertation follows the order of discussion in this Chapter and Fig. 1.1. Chapter 2 contains a description of the data used in the analysis and brief descriptions of the methodology employed. Large-scale lightning, thermodynamic and dynamical observations are presented in Chapter 3 followed by cloud-scale kinematic and electrical analysis in Chapter 4. Chapter 5 presents a description, synthesis, and discussion of 1-D modeling results (including model predicted convective heating rates). Chapter 6 presents analysis of the rainfall-lightning relationship for many regions of the globe, including the tropics. Chapter 7 presents concluding remarks and recommendations for future research.

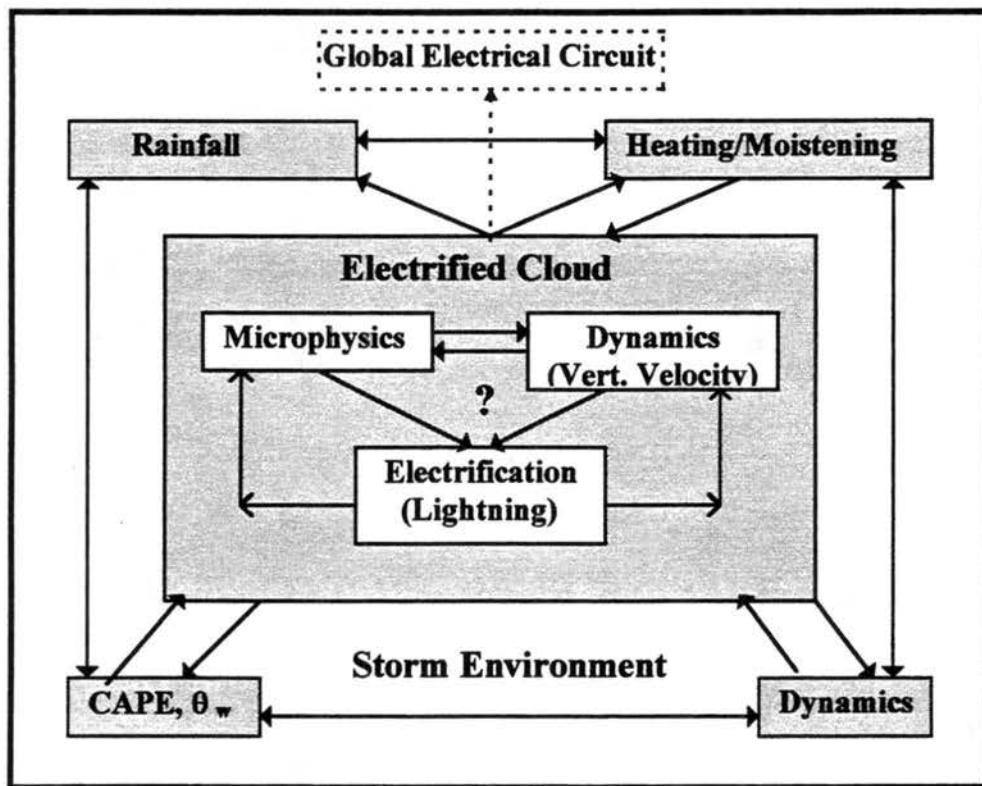


Figure 1.1 A process diagram illustrating the multiple scales affected by, or associated with, deep convective clouds. Arrows represent feedbacks between the different processes. Primary large-scale inputs are situated at the bottom of the diagram and primary outputs from the cloud scale are situated at the top of the diagram.

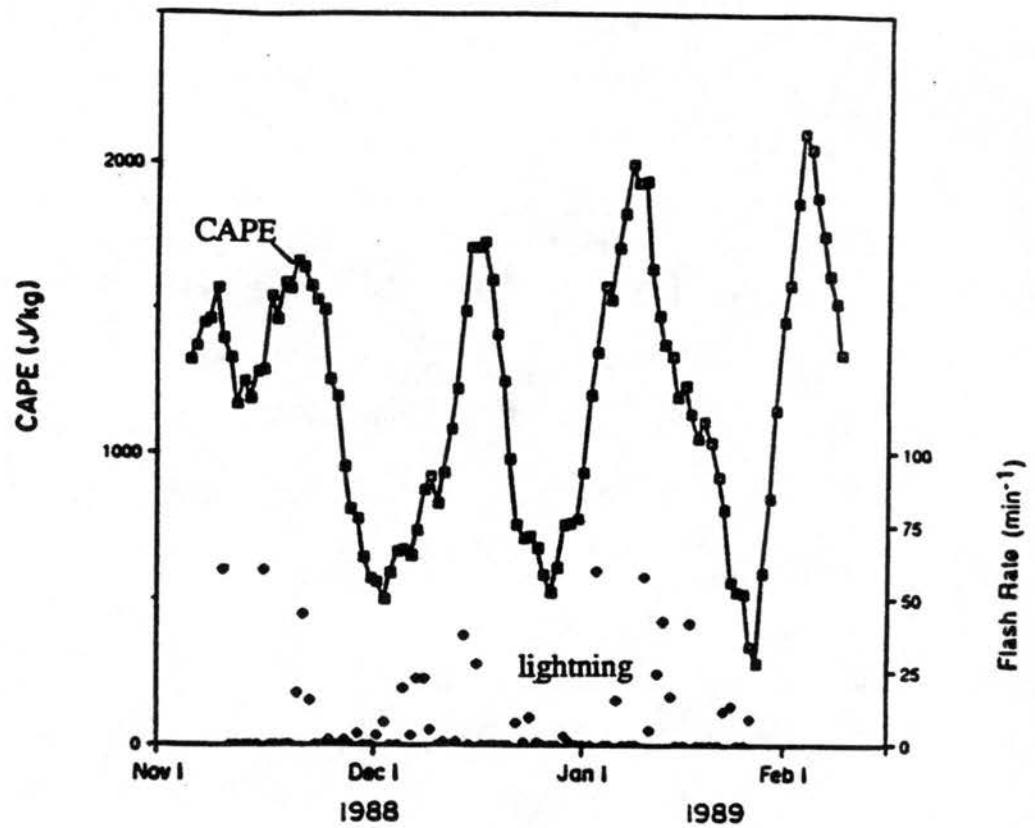


Figure 1.2 Time-series of 24 hour mean CAPE (line) vs. peak total lightning flash rate (dots) observed during the 1988-89 wet season of the DUNDEE (Darwin Australia). Figure adapted from Rutledge et al., 1992.

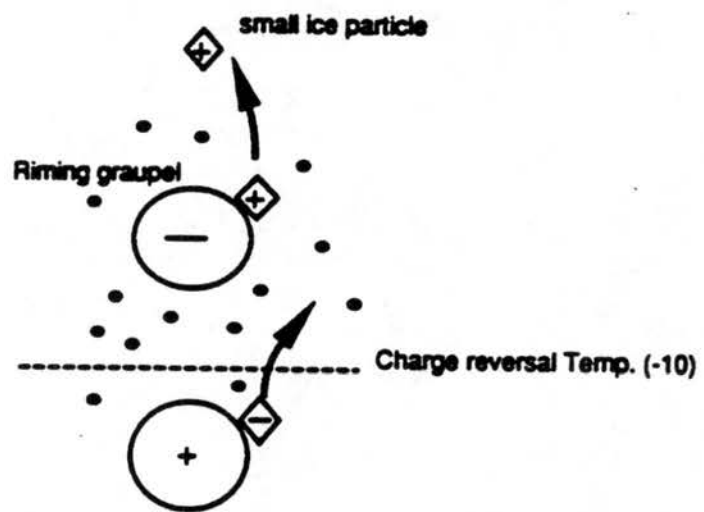


Figure 1.4 Illustration of the NIC process involving small ice particles and riming graupel in the presence of supercooled water droplets (small dots). Below the level of charge reversal (-10°C), the sign of charge transferred to the graupel (ice) reverses to positive (negative).

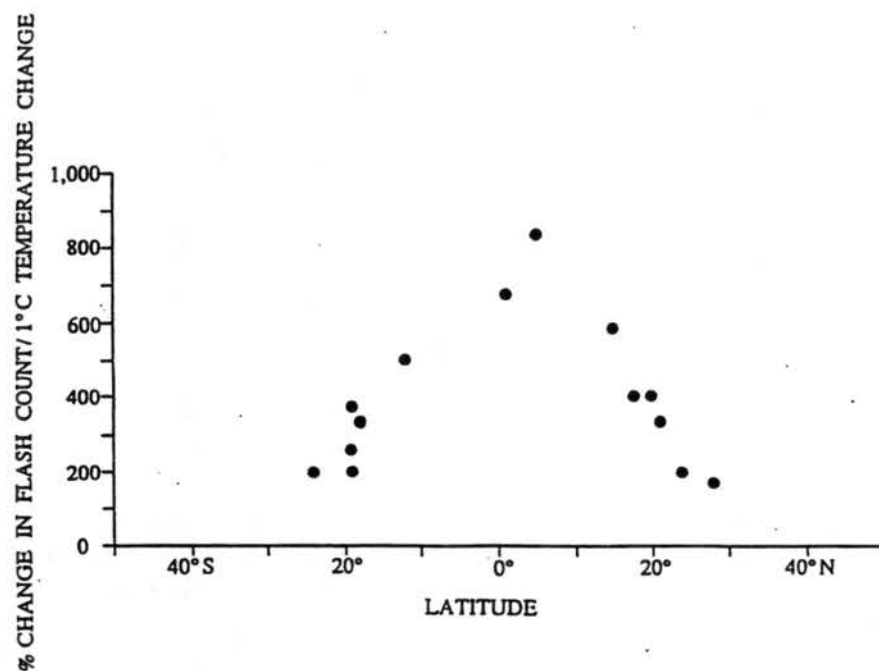


Figure 1.5 Percent change in total flash count per 1°C change in surface wet-bulb temperature with latitude (adapted from Williams et al., 1994).

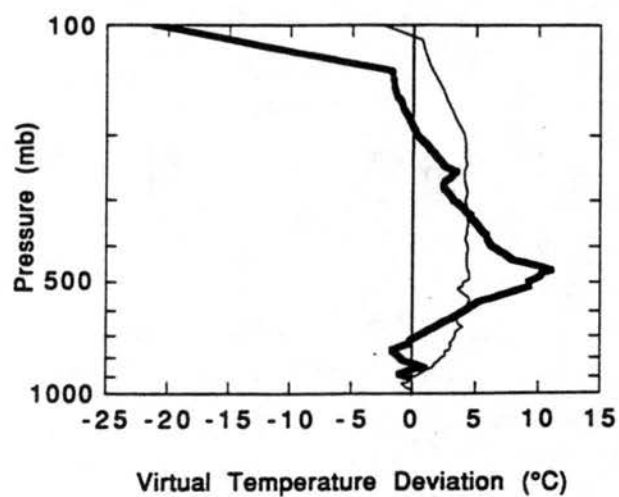


Figure 1.6 Example of different thermal buoyancy profiles vs. pressure (virtual temperature deviation) for parcels lifted in tropical oceanic (light solid line) and mid-latitude continental (bold solid line) environments (adapted from Lucas et al., 1994a).

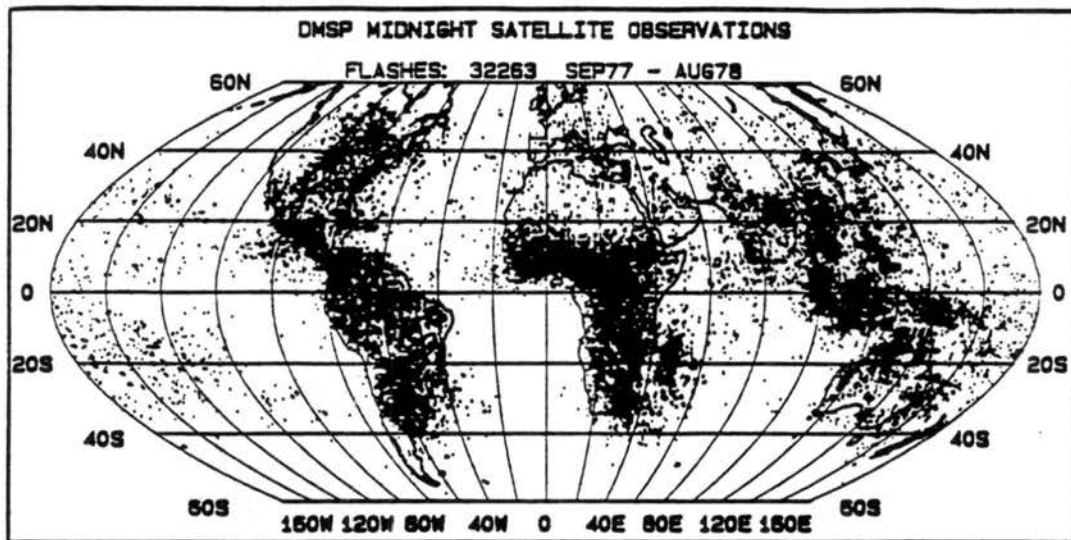


Figure 1.7 September 1977 to August 1978 DMSP midnight satellite observations of total lightning from 60°N to 60°S (adapted from Orville and Henderson, 1986).

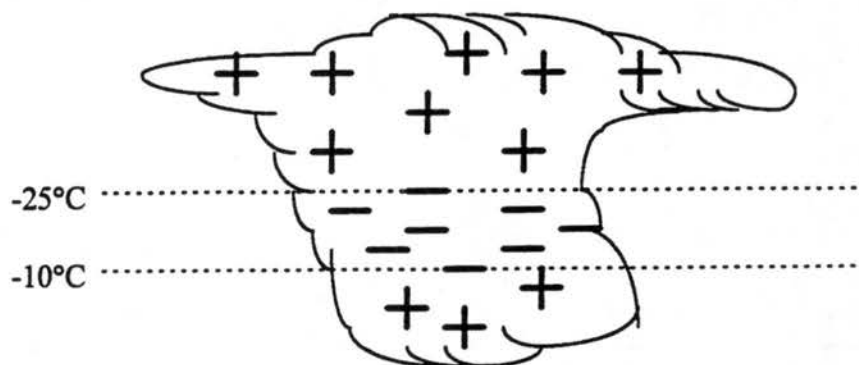


Figure 1.8 The classic tripole model of thunderstorm. Positive charge (+) is found at temperatures above -10°C , negative charge (-) between -10°C and -25°C . Diffuse positive charge is found at temperatures colder than -25°C .

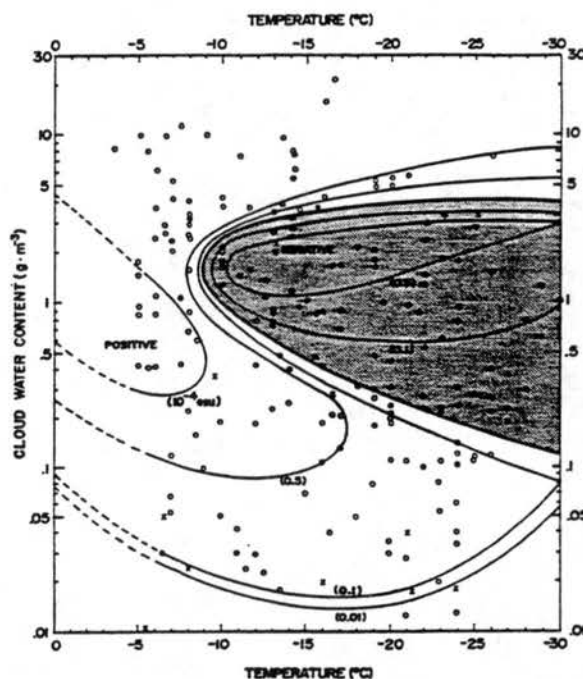
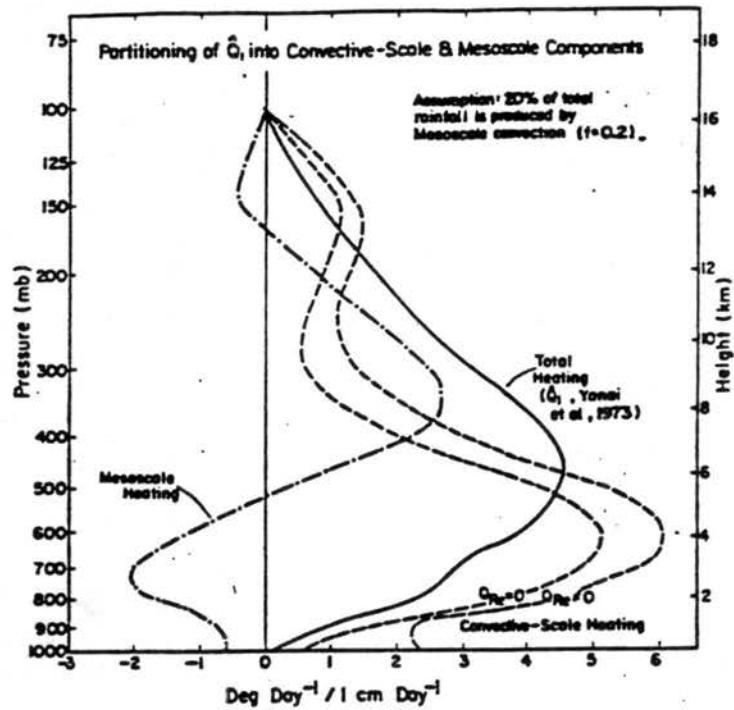


Figure 1.9 Non-inductive charging results from the laboratory experiments of Takahashi (1978). Cloud-water content (g m^{-3}) is indicated on the ordinate, temperature ($^{\circ}\text{C}$) on the abscissa. Data points represent charge transferred (e.s.u.) to a riming ice target in collisions with an ice-particle. Shaded areas represent regions of temperature-liquid water content space where negative charge is transferred to the target. Contours are in units of 10^4 esu.

a)



b)

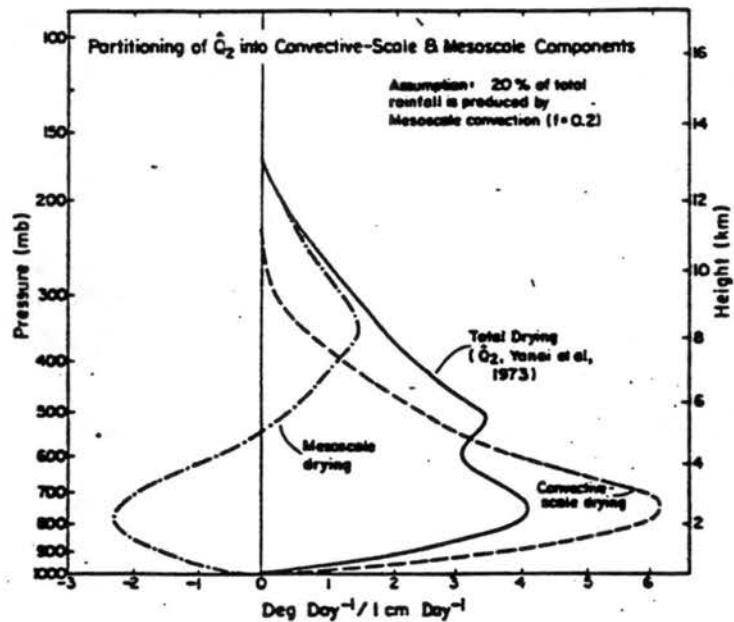


Figure 1.10 Example of convective and mesoscale a) heating and b) drying profiles (adapted from Johnson, 1984). Total heating and drying profiles are indicated by a solid line, convective heating/drying profiles by a dashed line, and mesoscale heating/drying profiles by a dash-dot line.

CHAPTER 2

DATA ANALYSIS AND METHODOLOGY

This chapter describes *observational* platforms and data analysis methods utilized in this study (modeling methods are discussed in Chapter 5). Data analyzed in this study were collected over in a variety of locations (Fig. 2.1) across the globe including the western Pacific warm-pool (TOGA COARE; CEPEX¹); the maritime continent region of northern Australia (ITEX²; DUNDEE); French Guyana; and the continental U.S.

2.1 TOGA COARE: Data and observational platforms

During the period from 1 November 1992 through 28 February 1993 observations of electrified oceanic convection and associated cloud-to-ground lightning were obtained over remote regions (Figs. 2.1, 2.2a-b) of the tropical western Pacific Ocean (equatorial warm-pool) during TOGA COARE (Webster and Lukas, 1992). These observations were obtained using several different instrument platforms including a three-station advanced lightning detection and location network (Advanced Lightning Direction Finder network; ALDF; Orville et al. 1994), a variable gain flat plate antenna (operated by M. Brook, New Mexico Institute of Mining and Technology, and described in Orville et al. 1994), shipboard and airborne Doppler radars

¹ CEPEX; Central Equatorial Pacific Experiment (CEPEX Operations Summary, UCAR, 1993)

² ITEX; Island Thunderstorm EXperiment, (Keenan, 1989).

(Rutledge et al. 1993; Jorgensen et al. 1995), and shipboard and airborne electric field mills (Petersen et al. 1993; Blakeslee and Christian, 1994). Other primary meteorological instrument platforms deployed during COARE included network sounding data (Johnson et al. 1993; Parsons et al., 1994), and a variety of oceanographic buoys and research vessels instrumented with meteorological sensors (e.g., wind, temperature, humidity and precipitation).

2.1.1 The COARE lightning detection network

During the months November-December 1992, three high gain ALDF antennas manufactured by LLP (Lightning Location and Protection Inc., Tucson, Arizona) were installed at Kavieng, Papua New Guinea (PNG), Rabaul PNG, and Kapingamarangi Atoll (2.58°S, 150.81°E; 4.22°S, 152.18°E; and 1.07°N, 154.81°E respectively; Fig. 2.2a). The three-station network became fully operational in late December 1992, and data were collected through virtually all of 1993. The data collected by the ALDF network for each CG return stroke included the time of occurrence (to the nearest microsecond, and time stamped at each antenna using a Global Positioning Satellite [GPS] receiver), azimuth, polarity, and the electric and magnetic field strengths.

Since the COARE ALDF network needed to accurately detect and locate CG return strokes that occurred within the Intensive Flux Array (IFA; Fig. 2.2a-b), the required maximum effective range of the network was on the order of 900 km. This range far exceeds the 600 km maximum previously recommended for use in ALDF networks (Brook et al. 1989). Therefore, the network antenna gains were increased by a factor of 1.5 (Orville et al. 1994) and lightning flash waveform acceptance criteria were adjusted to increase the sensitivity (and range) of the network.

In contrast to lightning location networks operated in the U.S. and other areas of the world, no central position analyzer (e.g., Orville et al. 1983) was used to locate the CG lightning in real time due to the remote nature of the ALDF sites. Thus, the data were collected and stored at each ALDF site, and processed after the data were later returned to the U.S. Due to the limited number of sites available for antenna installation, there were towers, fences and buildings located near the direction finders. Therefore, site errors in the measured azimuth bearings were potentially quite large. This made magnetic direction finding methods (e.g., DF; triangulation of at least *two* azimuth bearings) less accurate for ranges exceeding 200-300 km from the network. Fortunately, each ALDF antenna in the network was connected to an absolute time source (GPS clock) which enabled time of arrival methods (TOA; e.g., Hughes and Gallenberger, 1974) and “hybrid” TOA-DF methods (requiring three ALDF’s to detect a given stroke) to be developed to accurately determine *flash* locations and site errors at each antenna location (herein, *flash* refers to a set of cloud-to-ground return *strokes* separated by less than 500 ms in time of occurrence and less than 10 km in distance; e.g., Uman, 1987; Watson and Holle, 1994). The location accuracy of the TOA-DF “hybrid” method was approximately 10 km at a range of 900 km. Specific algorithms developed to calculate CG flash locations from raw data are described in Appendix A.

The lightning data were examined over many different spatial and temporal scales. Accurate “hybrid” (TOA-DF method) CG locations were used for examining cloud-scale electrical development, while less accurate (location errors of 20 km or more at ranges of 600 km) DF solutions were used over longer ranges for large-scale studies where cloud-scale accuracy was not required. However, in order to use the “hybrid” location method, *all three* ALDF antenna’s must have detected a given CG return stroke. This requirement effectively reduced the relative detection efficiency of the network to approximately 60% of the total number

of flashes detected by any *two* ALDF's in the network. Hence, using two-station DF solutions in the "long-range" mode reduces the accuracy of the locations but more accurately represents the total number of flash detections.

In the case of "large-scale" studies of lightning and other atmospheric variables such as CAPE, the ALDF network data functioned as a type of "constrained" CG flash counter for the areas and times specified. Of primary interest to this study were CG flashes detected over isolated regions of the tropical ocean since those data comprised a truly unique set of observations. Hence, a large "ocean-sector" was selected (area $O[10^6 \text{ km}^2]$; Fig. 2.5) over which development of the flash location algorithms were concentrated. Flashes located in the ocean-sector were then processed to determine the diurnal cycle of oceanic lightning (ensuring a large sample), and later compared to the diurnal cycle of variables such as cold-cloud area (proxy for the presence of deep convection), CAPE, and rainfall. When examined in time series form with CAPE, 24 hour CG flash counts were smoothed (e.g., tophat filter) to eliminate some of the noise present in the data.

In addition to the presentation of unique observations of tropical oceanic lightning flash density, a new approach was taken in this study with regard to the presentation of spatial and temporal variations in CG lightning. This involved plotting the CG data in time-longitude coordinates over a section of the COARE domain, similar to the conventional presentation of OLR data in Hovmöller diagrams (e.g., Fig. 2.3). The lightning-Hovmöller diagrams were created to investigate the potential for using CG lightning data in the detection of deep convective phases associated with propagating wave disturbances. To quantify and/or validate any visually inferred spectral characteristics in the "lightning Hovmöller diagram", spectral analyses of the CG lightning data were also performed.

2.1.2 Shipborne electric field data

An inverted rotating shutter, electric field mill was deployed on the R/V Vickers (Fig. 2.2a) from approximately mid-December, 1992 through the end of March, 1993 (the field mill was operational during the Vickers crossing of the equatorial Pacific Ocean during CEPEX). The field-mill possessed a 10-90% response time of 0.1s, and a dynamic range of $\pm 20 \text{ kV m}^{-1}$. Thus the field mill could make coincident measurements of both the slowly varying DC component of the surface electric field in fair weather conditions or under a thunderstorm (vertical component), and any high-frequency transients associated with lightning (within a 20 km range). Therefore, when deep convection was near the ship, field mill data could be used concomitantly with radar and CG lightning data to study the electrical characteristics of the convection. Field mill-detected lightning flashes also served as a type of "ground-truthing" for the ALDF network when CG lightning occurred near the R/V Vickers.

From mid-December, 1992 to late January, 1993 the field mill was connected to a non-recording volt-meter, which required the presence of a scientist to monitor and transcribe values of the measured mill voltage during convective events. Upon return from the second cruise of the COARE intensive observation period (IOP), the field mill was connected to a variable-gain, multi-speed analogue strip chart recorder and also to a personal computer (so that digital data could be recorded). The strip chart recorder and PC-computer provided a relatively continuous data collection from 26 January-March 20, 1993.

Unfortunately, it was not possible to calibrate the electric field mill during the COARE IOP. In the post-collection phase it was decided that values of the recorded field mill voltages would be converted to values of electric field (V/m) by assuming the fair weather electric field over the ocean was "well-behaved", possessing a value of 120 V/m based on previous

measurements (the R/V Carnegie cruise; Torreson et al., 1946). Previous laboratory calibrations of the field mill were linear, and therefore a simple linear calibration of measured voltage to electric field value was developed based on the output voltage of the field-mill under fair weather conditions. However, the form factor due to the presence of the ship's conductivity and associated enhancement of the electric field is unknown (and may not be a linear function of field strength). It is estimated that the true magnitude of the electric field (but not the trends or the sign of the electric field) may vary from the measured field by as much as a factor of two in some of the more vigorous thunderstorm cases (based on visual inspection of the data).

2.1.3 Sounding analysis

Sounding data collected during TOGA COARE were collected over a Large Scale Array (LSA; Fig. 2.2b; Webster and Lukas, 1992) containing an Outer Sounding Array (OSA) and an Intensive Flux Array (IFA). The data collection sites were comprised of ISS sites (Manus, Kapingamarangi, Kavieng, Nauru, research vessels R/V Kexue 1, R/V Shiyan 3, R/V Xiangyanghong 5 [PRC #5], and the R/V Moana Wave), non ISS sites such as the R/V Vickers (late January-February 1993), and other first order sounding stations in the vicinity of the COARE LSA. Soundings were launched within the LSA approximately four times per day (00, 06, 12, 18 UTC). Further description of the radiosonde network and sonde type can be found in Parsons et al. (1994).

In this study, high quality (calibrated and quality controlled) sounding data collected four times daily aboard the R/V Vickers (Jan. 28 - February 27, 1993) were used to compute variables such as the CAPE (using a parcel with the mean characteristics of the first 50 mb of the sounding and Eq. 1.1), vertical profiles of thermal buoyancy, mixed-layer θ_w , and the mean

mixed layer mixing ratio (MLMR), all highly dependent on the accuracy of measured values of temperature and moisture in the boundary layer. Errors discovered (and still present) in the ISS near surface moisture measurements render those data of questionable value for calculations of CAPE (recall the sensitivity of CAPE to relative humidity discussed in Sec. 1.1). Therefore, the ISS data were not used for calculating CAPE (at least not as the primary platform). Gridded sounding data (including the ISS sites, gridding process described in Lin and Johnson, 1996a), were used to examine moisture convergence (gridded values supplied by P. Ciselski, CSU), large-scale vertical motion in the IFA (point values supplied by X. Lin; CSU) and wind-shear relative to lightning flash counts in the IFA.

When using the R/V Vickers sounding data for computing storm environment variables such as CAPE, θ_w , or MLMR, care was taken to eliminate soundings contaminated by the presence of rainfall over or in the immediate vicinity of the R/V Vickers. This was accomplished by visually inspecting MIT radar logs and optical raingauge summaries to assess the presence or absence of rainfall. The individual soundings were then composited into four soundings collected for each six hour period over the diurnal cycle. Parameters such as CAPE, wind-shear, MLMR, PBL-temperature, and PBL- θ_w , were then computed for each of the 6-hour composites, providing diurnal cycles (four points) of different derived sounding parameters for comparison to the diurnal cycle of CG lightning.

Following computation of the diurnal cycles, the CAPE and CG flash counts were filtered (three point top-hat filter) and compared in time-series form with weekly mean 1000 mb moisture convergence, rainfall and upward motion (ω). Further, the filtered time-series of lightning data and CAPE were used to compute a mean CAPE and CG flash density that were used to partition and composite the soundings into an "active" regime (a regime characterized by CG lightning flash count much larger than the mean) and a "weakly-electrified" regime (regime characterized

by the presence of rainfall, but less lightning activity than the mean). This enabled a gross comparison to be made of mean thermal buoyancy profiles partitioned by electrical intensity. The composited soundings also served as the initial thermodynamic input for 1-D cloud-modeling experiments (discussed in Sec. 1.6 and Chap. 5).

2.1.4 Surface meteorological data

The COARE surface data analysis includes time-series of buoy and shipboard observations of surface temperature, wet-bulb temperature, relative humidity, wind speed and direction, and sea surface temperature (SST) at .05 m, .5 m and 4 m below the surface. In addition to the raw surface meteorological observations, derived fields such as surface sensible and latent heat fluxes calculated from IMET buoy data were examined (Fig. 2.2a; data supplied by Dr. R. Weller, WHOI). Time-series of the meteorological data were then compared to that of CAPE and CG lightning in order to investigate the phasing relationship between the surface variables (e.g., Zhang, 1995) and the occurrence of electrified convection.

2.1.5 Radar platforms and analysis

During TOGA COARE the MIT³ and TOGA C-band Doppler radars were installed on the R/V Vickers and PRC #5 (Fig. 2.2a) respectively. An inertial navigation unit (INU) mounted at the base of the antenna was used to provide rapid updates (200 Hz) of pitch, roll and heading information to the radar antenna controller, thus enabling the azimuth and elevation of the antenna to maintain desired values (Rutledge et al., 1993). Global Positioning System (GPS)

³ The MIT radar was also used during CEPEX.

receivers provided the latitude and longitude coordinates of the radar. The operational characteristics of the individual radars are summarized in Table 2.1 and further discussed in Rutledge et al., (1993). The shipboard radars used during COARE were tasked with mapping surface rainfall and characterizing the lifecycle and morphology of convection over the warm-pool, and as a result, the scanning strategy reflected these requirements. Each 10 minute scanning cycle of the radar was comprised of a long range (low Pulse Repetition Frequency; PRF) rainfall surveillance scan (3 unique elevation angles), a high PRF full volume scan (17 unique elevation angles), and approximately 1 minute of time for completing specialized scans such as RHI's (Range-Height-Indicator; vertical cross-sections taken at user specified azimuth angles). This scanning strategy provided rainfall rate information (using an empirically derived reflectivity-rainfall [Z-R] relationship; e.g., Short et al., 1995), and three-dimensional information on storm structure (using volume scans of radar reflectivity and Doppler velocity).

Table 2.1 Characteristics of the MIT, TOGA and NOAA P-3 Tail-Radars (Cifelli, 1995)

Characteristic	MIT	TOGA	NOAA P-3
Wavelength (cm)	5.37	5.35	3.22
Peak Power (kW)	155	250	60
Pulse Length (μ sec)	1.0	.48, 1.82	0.5
Beamwidth ($^{\circ}$)	1.6	1.55	1.35 Horz., 1.90 Vert.
Minimum Detectable Signal (dBm)	-115	-105	-111
Pulse Repetition Frequency (s^{-1})	250-1500	250-1500	1600
Antenna Gain (dB)	40.5	40.0	40.0
Number of Gates	581	581	512

The COARE radar data (reflectivity and radial velocity) were used in this study to examine the *cloud-scale* vertical and horizontal structure of lightning and non-lightning producing convection over the warm-pool. Raw radar reflectivity (dBZ) data were first viewed and edited using Research Data Support System software (RDSS, developed at the National Center for Atmospheric Research, NCAR; Mohr, 1986) and subsequently gridded in three dimensions (with a range dependent grid spacing to maximize resolution) using the "REORDER" software package also developed at NCAR. The gridded reflectivity data were further processed to indicate the amount of storm volume (mass) present at specified reflectivity thresholds (e.g., 0, 10, 20, 30, 40, 50 dBZ) for each height bin above the -10°C level (i.e., the mixed-phase region of the cloud). In this fashion a reflectivity (mass) "threshold" in the vertical plane associated with lightning producing storms could be determined (hypothesized to be 30-35 dBZ at -10°C based on the author's observations during COARE). When this analysis was completed over the lifecycle of a given case, the maximum height of occurrence associated with each reflectivity threshold could be plotted in time-height form with observed electricity variables such as electric field strength or CG lightning flash count. Hence, coincident examination of both the electrical and convective lifecycles of tropical oceanic convection were possible.

Additionally, composite analyses of the MIT radar data presented in DeMott (1996; vertical structure) and Rickenbach (1996; rainfall and horizontal convective morphology) were compared to CG-lightning flash density over daily and monthly time-scales for January-February 1993 (months where lightning data were collected during COARE). Vertical-structure (dBZ profiles) statistics discussed in DeMott (1996) were used to extend the "case-by-case" cloud-scale examination to broader temporal and spatial domains (e.g., daily/monthly time-scales over the radar coverage area, $\sim 7 \times 10^4 \text{ km}^2$). For example, daily mean convective 30 dBZ heights

computed by DeMott (used as an electrification discriminator; Larson and Stansbury, 1974) were compared to area-integrated daily CG-lightning flash counts.

Similarly, a time-series of daily mean rainfall over the MIT coverage area computed by Rickenbach (1996) was combined with a time-series of CG-flash count over the approximate area of radar coverage. This enabled a comparison to be made between CG-lightning frequency and oceanic rainfall amount over daily time-scales. The relationship between the rainfall and CG-lightning was expanded to a larger temporal scale by computing a mean monthly rain-yield (area normalized ratio of rainmass to CG flash count; see Sec. 1.5) for the coverage area of the radar. The resulting tropical oceanic rain-yield value could then be compared to rain-yield values calculated for other tropical and mid-latitude locales (e.g., Table 1.1; Chapter 6).

In addition to the reflectivity characteristics of the convection, the TOGA COARE radar platforms also sampled radial velocity. When two Doppler radars sample the same cloud-volume (dual-Doppler), the two independent radial velocity measurements can be combined with the anelastic continuity equation, and appropriate boundary conditions to synthesize the 3-D wind field (Ray et al., 1980; Doviak and Zrnic, 1993). Indeed, on 9 February 1993 the first *successful* dual-Doppler data collection conducted at sea took place between the MIT and TOGA radars. On this date, an oceanic squall-line (weakly electrified) formed and passed through the lobes of the dual-Doppler coverage area, and was sampled for 2.5 hours at 10 minute temporal resolution.

A dual-Doppler synthesis was performed on the 9 February case, and u , v and w components of the wind were retrieved (e.g., Cifelli, 1996). The vertical motion (w) was utilized to diagnose the vertical velocity found at the top of individual 30 dBZ reflectivity cores (convective cells), at or above the height of freezing level, for each of the 12 sample times processed. This provided a method to address the hypothesis of Zipser and Lutz (1994) that an updraft "threshold" of $\sim 6 \text{ m s}^{-1}$ is required (relative to the terminal fall speed of a frozen drop)

prior to a tropical oceanic cloud becoming significantly electrified (e.g., Fig. 2.4). Moving to the system scale, the system vertical motion field associated with the weakly electrified convection of the 9 February case was partitioned and averaged into convective/stratiform, updraft/downdraft components (Cifelli, 1996; DeMott, 1996) for comparison to strongly electrified convection sampled by P-3 aircraft the same day (e.g., Smull et al., 1994; Petersen et al., 1995).

In contrast to the dual-Doppler analysis of shipboard radar data, airborne X-band Doppler radars flown on two NOAA P-3's (Table 2.1) and the NCAR Electra aircraft, provided a more precise method (quad-Doppler; "3 or 2" solution technique; Jorgensen et al., 1996) of determining the 3-D wind field. The airborne radars were operated in the FAST mode (Fore-Aft Scanning Technique; Jorgensen et al., 1996) which essentially provides four separate samples (four beams) of the same storm volume when two aircraft are sampling the same system. Hence, the "quad-Doppler" technique can be utilized to achieve a higher degree of accuracy in determination of the 3-D wind field (e.g., Jorgensen et al., 1996). In comparing dual and quad-Doppler techniques, Jorgensen et al., (1996) note that the vertical velocity in the upper most regions of the quad-Doppler synthesized wind fields can be up to 50% larger than that diagnosed from dual-Doppler techniques (a result of the zero vertical velocity assumption typically made for the top boundary in conventional dual-Doppler syntheses).

For comparison to the shipboard case, this study utilizes wind-fields sampled in three cases (9 February 1993, 21 February 1993, and 22 February 1993) by the NOAA P-3's during TOGA COARE. The computed wind-field data were supplied by Dr. D. Jorgensen, Dr. B. Smull and Dr. S. Lewis of NOAA. The reflectivity and wind-field data (especially the vertical velocity field) for each case were compared to lightning flash data collected by the ALDF network to assess the relative degree of electrical activity. As in the shipboard case, the vertical motion data associated with the tops of *individual* 30 dBZ convective reflectivity cores (lightning and non-

lightning producing) for each aircraft case were examined to investigate the relationship between 30 dBZ height, storm electrification and updraft intensity. Note however, that the dual/quad-Doppler synthesis process is associated with a high degree of filtering (i.e., the data are gridded and smoothed first) and as such, the diagnosed vertical velocities are likely to be more representative of an *average* over a given grid box.

In addition to the examination of individual convective cores and associated vertical velocities, the airborne wind-field data were partitioned (Cifelli, 1995) into stratiform and convective components based on the horizontal and vertical profiles of radar reflectivity (algorithm discussed in DeMott, 1996). The partitioned wind-fields observed in a highly electrified system sampled by the P-3 aircraft on 9 February 1993 (situated approximately 400 km southeast of the system sampled earlier in the day by the shipboard radars) were then composited and directly compared to the weakly electrified shipboard case.

2.1.6 Satellite data analysis

The Japanese Geosynchronous Meteorological Satellite (GMS) provided IR brightness temperature (TBB) data over the COARE domain (Fig. 2.2b). The data were analyzed from several different perspectives: 1) using the TBB as a proxy for cloud-top temperature, the TBB's were compared to the locations of CG flashes and inferences were drawn regarding the location and organization of convection embedded in the cold-cloud shield; 2) the areas of cold-cloudiness were averaged by hour of the day for several different brightness temperature thresholds (e.g., 245, 235, 225, 215, 205, 190 K) over the "ocean sector" (Fig. 2.5) to construct a diurnal cycle of cold-cloud area at different thresholds. These diurnal cycles were then compared to diurnal cycles of cloud-to-ground lightning and rainfall; 3) time-series of cold-

cloudiness and CG flash counts were constructed for the area covered by the MIT radar for the 60 day period 1 Jan. 93 to 28 Feb. 93, to compare trends in the two variables; and 4) GPI (Goes Precipitation Index; Arkin, 1979) IR-based mean monthly estimates of rainfall over 2.5° grid squares in the COARE domain (oceanic and mixed island-oceanic regimes; COARE areas 1, 2 and 3 in Fig. 2.6a) for January and February, 1993 were combined with coincident mean monthly flash density to produce rain-yields that could be compared to the MIT radar derived rain-yield estimate, and rain-yield estimates computed other regions of the globe (Fig. 2.6a-b). Some error is introduced into the rain-yield analysis over the ocean due to differences in the way the rainfall was diagnosed (e.g., satellite vs. radar estimates). However, based on the quality of the data (*all* of the rainfall data have been quality-controlled), and the temporal/spatial scales of concern in this study, we estimate that the absolute difference in diagnosed rainfall amounts due to differences in measurement technique is a factor of two or less.

2.2 DUNDEE/ITEX data and analysis

The DUNDEE (Rutledge et al., 1992) and ITEX (Keenan et al., 1989) experiments were conducted in the vicinity of Darwin, Australia (Figs. 2.1, 2.7a-b) during the years 1988-1990. Both field campaigns were conducted to learn more about the characteristics of deep convection occurring over the tropical Maritime Continent region. DUNDEE (Fig. 2.7a) focused on the dynamical, microphysical and *electrical* processes associated with deep convection while ITEX emphasized the study of convective initiation and the dynamical and microphysical processes associated with deep convection (often highly electrified) occurring over the Tiwi Islands (75 km northwest of Darwin in Fig. 2.7b). Both field programs utilized C-band Doppler radars, special sounding networks, mesoscale rain-gauge networks, and automated weather station networks

(Fig. 2.7a-b). Additionally, a four-station ALDF network was deployed for approximately 75 days during the 1989-1990 wet season of the DUNDEE (Fig. 2.7a)

In contrast to the COARE ALDF network, the DF network operational during the DUNDEE was not connected to a GPS time source. Hence time-of-arrival methods for locating CG lightning flashes could not be applied. Instead, the network data were processed by Lightning Location and Protection Inc. (LLP; Tucson, Arizona) using magnetic direction finding techniques (statistically optimized) with known antenna site errors (site errors were not measured or known for the COARE ALDF antenna's during deployment). The detection efficiency is assumed to be $\geq 70\%$ within a 300 km radius of the center of the network and errors in position are ≤ 10 km for non-baseline flashes (e.g., Petersen, 1992). Information provided for each CG flash by the DUNDEE network included the time of occurrence (to the nearest millisecond), location, polarity and peak current of the first return stroke, and the flash multiplicity (i.e., the number of return strokes comprising a single CG flash).

For comparison to COARE analyses, the DUNDEE lightning data were combined with raingauge data over the islands and the Australian continent to examine the relationship between lightning and rainfall for the island and continental regimes. Diurnal cycles were computed and partitioned into monsoon and break period rainfall regimes for both continental and island regimes. For the Tiwi Islands, the diurnal cycle of lighting was partitioned into "active" and "inactive" portions of the day to illustrate the effect of sea breeze forcing and convective organization on the associated lightning patterns observed over the islands. Once partitioned by rainfall regime (monsoon/break; island/continent), mean area-normalized rainfall and CG lightning counts were computed for areas of 10^4 - 10^5 km² for each of the respective rainfall regimes. The area mean rainfall and flash density were then formulated as ratios (rainmass to flash density) to produce rain-yield estimates for the different regimes (~45 break period days, 10

monsoon days). The rain-yield values calculated for the tropical island and continent were then compared to values for other regions of the globe (Fig. 2.6a-b; discussed in Chapter 6).

In addition to the CG-lightning, rainfall and CAPE analyses, a single, but representative island convective system (e.g., “Hector”; Simpson et al., 1993; Keenan et al., 1994) was also analyzed using C-band Doppler radar data collected by the NOAA/TOGA radar (radar characteristics similar to that shown in Table 2.1 for the TOGA radar) situated in Darwin. Due to the 1.5° beam width of the TOGA radar, the elevation angles used for scanning, and the radial distance to the islands, the analysis of the “Hector” case focuses primarily on the co-evolving convective CG lightning structure in the horizontal plane.

Sounding data collected at four locations over the Tiwi Islands during ITEX (data courtesy of Dr. Tom Keenan, BMRC) were composited for the closest launch times prior to peaks in the diurnal cycle of CG lightning (computed from the DUNDEE DF network data). A pre-storm “environment” CAPE and thermal buoyancy profile were then computed for comparison to other sounding profiles composited for different geographical locales (Fig. 2.1). Following this comparison, the island or “Hector” sounding composite was used as the initial condition or ambient “environment” for several simulations of “Hector” in cloud-modeling experiments (Chapter 5). Sounding data collected during the DUNDEE in Darwin, and lightning flash counts within range of the DUNDEE network were also combined to form a time series of CAPE and CG lightning. A plot of CG flash density in time-longitude format, similar to the COARE analysis (and complementing the results of Rutledge et al., 1992; Fig. 1.2) was then created to test the relationship between *CG lightning flash density* (as opposed to the total flash counts used in Rutledge et al., 1992) and CAPE in a regime associated with large CG flash-densities.

2.3 French Guyana lightning and rainfall analysis

To examine large-scale tendencies between rainfall and lightning flash density for other regions of the tropics, monthly rainfall totals (using both gauge and satellite estimates) and *total* lightning flash counts (in-cloud and cloud-to-ground components) detected by the SAFIR VHF lightning detection system (Richard, 1990) located in Kourou, French Guyana (5°S, 52°W; Fig. 2.1) were analyzed for the year of 1989 (lightning data courtesy of Dr. E. Williams, MIT). Lightning locations provided by the SAFIR system for 1989 extend to a maximum (though not constant) range of approximately 200 km (personal communication, E. Williams, MIT) providing a maximum areal coverage of $O(10^5 \text{ km}^2)$ centered on Kourou.

Mean monthly rainfall data for two 2.5° grid squares centered on Kourou, were computed by the Global Precipitation Climatology Center (GPCC) as part of a merged precipitation data set compiled by the Global Precipitation Climatology Project (GPCP; Janowiak and Arkin, 1991). The GPCC rainfall data are comprised of both gauge and satellite (e.g., GPI) estimates of monthly rainfall. Unfortunately, the GPCC area-wide estimate (2.5° x 2.5° grid square) of mean monthly rainfall for 1989 over the region centered on Kourou, is heavily weighted toward the data collected by a *single* raingauge located in Cayenne, located approximately 50 km south of Kourou. Single point measurements of rainfall are inconsistent with the both the areal coverage provided by the SAFIR network, and previous large-scale comparisons of rainfall to lightning data (e.g., the rain-yield calculations shown in Table 1.1 and discussed herein). Therefore, GPCP satellite based (GPI) estimates of rainfall (e.g., Janowiak and Arkin, 1991) were used for comparison to the lightning data collected in French Guyana. The mean monthly rainfall and lightning data were then plotted in time-series form, partitioned by wet or dry season (similar to the partitioning of the DUNDEE monsoon and break period rainfall/lightning data),

and then diurnal cycles of lightning were computed for each rainfall regime (hourly rainfall data were not available so only the diurnal cycles of lightning could be computed).

2.4 United States

In order to place the tropical observations and modeling results into a useful frame of reference; lightning, radar and rainfall data from the mid-latitude continental U.S. (Fig. 2.1) were analyzed to illustrate distinct differences in the dynamical, microphysical (including the rainfall process), and electrical processes associated with mid-latitude continental and tropical oceanic convection. This was accomplished three ways: 1) from the perspective of cloud-scale processes (Fig. 1.1), radar and lightning data from the arid southwestern U.S. (central Arizona) were analyzed and contrasted with similar analyses from the TOGA COARE; 2) mid-latitude continental sounding data were composited from rawinsonde observations collected during PRE-STORM (Preliminary Regional Experiment for Stormscale Operational Research Meteorology; Cunnings, 1986) for use in both observational and modeling comparisons; and 3) rainfall and lightning data were combined to produce rain-yields for many areas of the continental U.S.

2.4.1 Data used in the analysis of electrified convection in central Arizona

An analysis of electrified convection that occurred over central Arizona during the summer monsoon season of 1994 provided an observational contrast to the convection sampled over the tropical ocean. Cloud-to-ground lightning information collected by the U.S. National Lightning Detection Network (NLDN; e.g., Orville et al., 1983; Cummins et al., 1995) were combined with NEXRAD Doppler radar data (c.f., Doviak and Zrnica, 1993 for a general description of the

NEXRAD radar) collected in Phoenix, AZ to facilitate study of the convective and electrical evolution of desert monsoon convection. The Phoenix NEXRAD reflectivity and radial velocity fields were processed (noise removal, velocity dealiasing), gridded, and then analyzed in a fashion similar to the TOGA COARE data. Time-height-reflectivity profiles were created with time-series of CG flash counts for several different cases. Additionally, the evolution of the differential radial velocity associated with six downburst cases (characteristic of the desert convection examined) were compared to CG flash rates and reflectivity structures. The analysis of the arid continental electrified convection provided an interesting contrast to the tropical oceanic cases examined, and served to illustrate important dynamical and microphysical differences associated with the electrical evolution of the two distinct types of cloud (arid continental vs. tropical oceanic cumulonimbi).

2.4.2 PRE-STORM sounding data

To complement the observational comparison made between electrified mid-latitude continental convection (Sec. 2.4.1) and tropical convection, the distinct convective environments associated with the two locales (tropical vs. mid-latitude) were also tested in 1-D cloud-modeling simulations (Chapter 5). In order to accomplish this comparison representative, 10 quality controlled rawinsonde observations launched in the immediate vicinity of convective systems observed during PRE-STORM were composited into a single sounding (data supplied by J. Meitin; NOAA-NSSL). The composited PRE-STORM sounding is intended to represent the gross structure of a pre-convective storm “environment” for a convective system situated in the middle of the continental U.S. Consistent with the “multi-scale” approach of the dissertation, the composited mid-latitude sounding provides yet another “data-point” for quantitative comparisons

of thermal buoyancy with respect to climate regime (i.e., tropical oceanic, continental, island, mid-latitude continental), and also provides the basis for a test of the cloud-scale response to the characteristic buoyancy differences in cloud-model simulations.

2.4.3 Data and methods used for calculation of rain-yields in the continental U.S.

Rain-yields were computed for the summer seasons (June-August) of 1994 and 1995 for a large portion of the United States (Fig. 2.6b). NLDN CG lightning data were used with quality controlled raingauge data of the highest resolution possible (typically gauge densities in all networks were $O[10^{-3} - 10^{-4} \text{ km}^{-2}]$) provided by the National Climatic Data Center (NCDC) cooperative raingauge network. Once the rainfall and lightning data were tabulated, monthly and seasonal area mean rain-yields (kg/fl) were calculated. To maintain consistency with the underlying assumption concerning the “convective” nature of the rain-yield (Sec. 1.5.2), data points associated with significant amounts of precipitation judged to be “non-convective” (e.g., frontal/stratiform in character with no initial convective phase; a problem at latitudes $> 40^\circ \text{ N}$), or dominated by single heavy precipitation events such as hurricanes (which tended to cause a step function increase in the rain-yield) were considered outliers and removed from the general sample, *but not from further analysis*. Because rainfall and CG lightning were continuously collected over the continental U.S., mean summer season rain-yields could be calculated and partitioned into rainfall regimes based on the seasonal mean rain-mass and CG flash counts associated with areas defined to fall within a given rainfall regime (“regimes” were defined as regions with similar mean rain-yields and geographical/meteorological similarities; Fig. 2.6b).

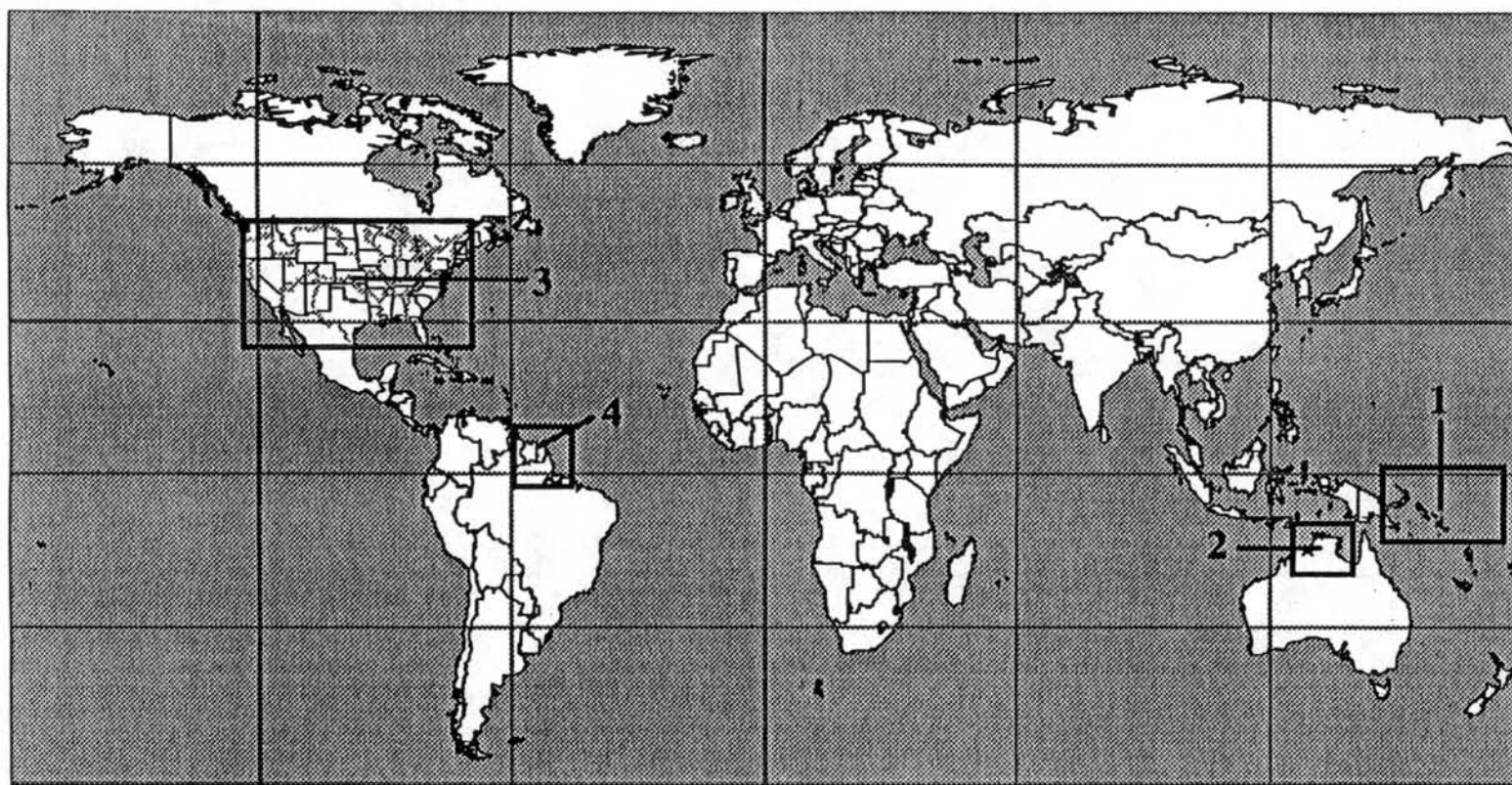


Figure 2.1 Map of locations examined in various aspects of this study.

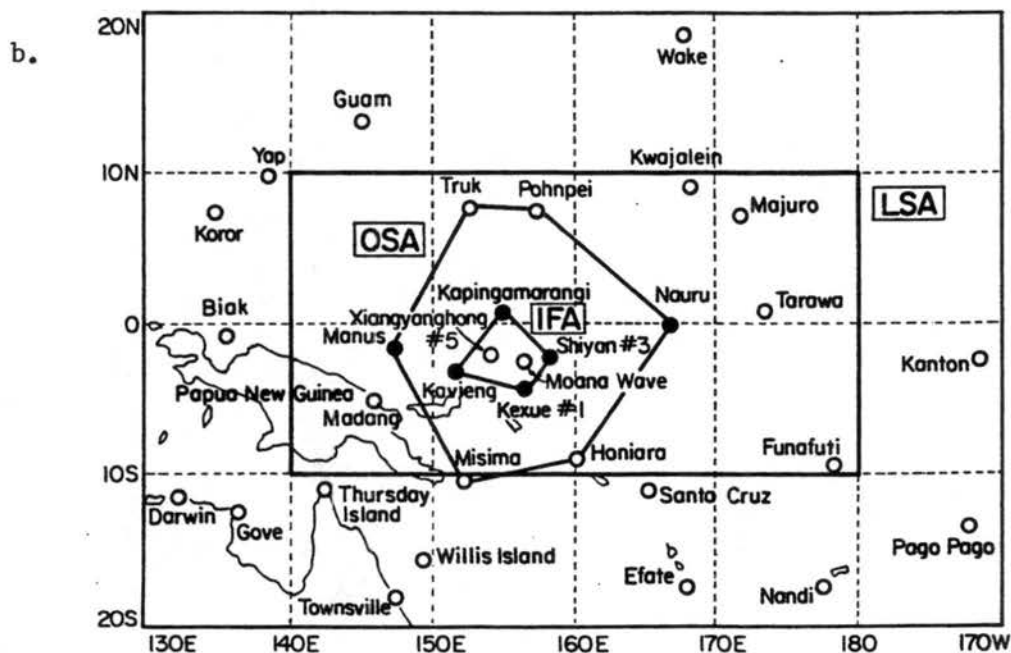
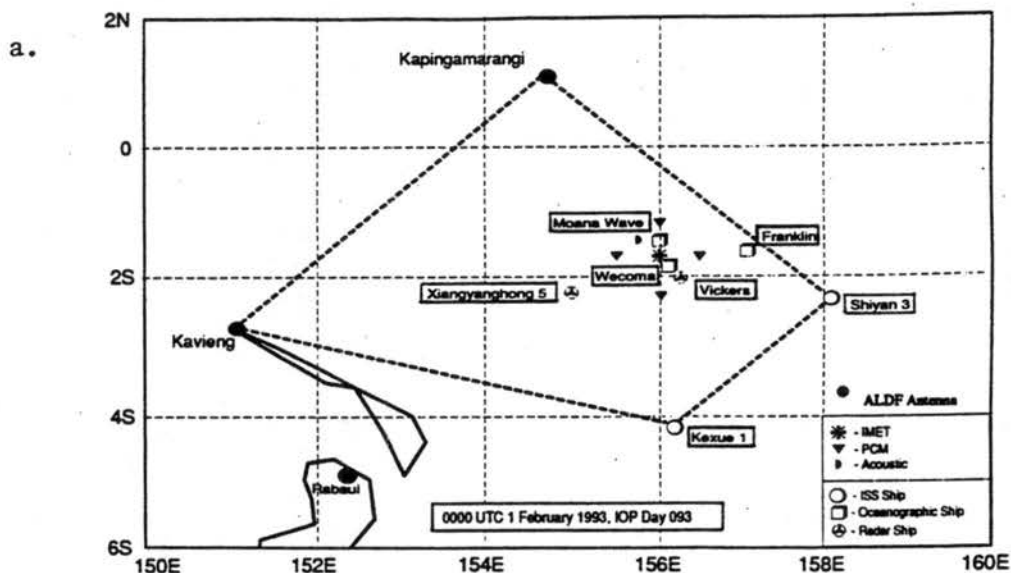


Figure 2.2 a) The COARE Intensive Flux Array (enclosed in dashed line) on 0000 UTC 1 February 1993. Dark filled circles indicate ALDF antenna locations; b) The TOGA COARE large-scale (LSA) and outer (OSA) sounding array locations.

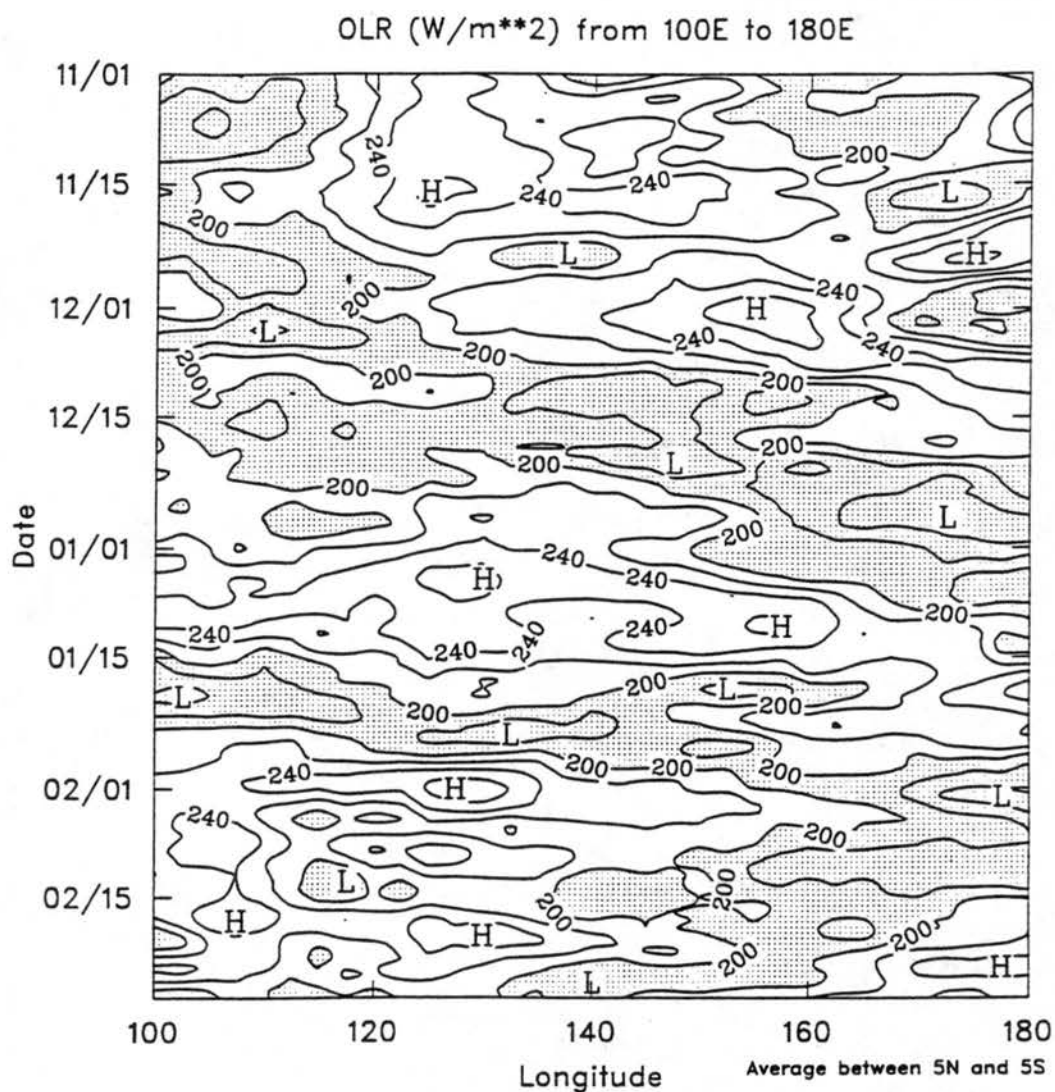


Figure 2.3 Time-longitude diagram of filtered mean OLR (W m^{-2}) for latitudes 5°N to 5°S during the COARE IOP (courtesy Mr. X. Lin, CSU). Values $\leq 200 \text{ W m}^{-2}$ are shaded.

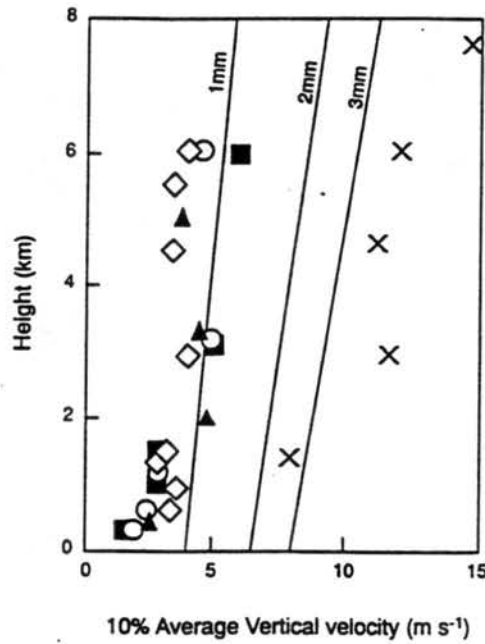


Figure 2.4 Average vertical velocity in the upper 10% of updraft cores sampled over tropical oceans (triangles, circles, diamonds and squares) and over land (x's), vs. terminal fallspeeds for raindrops of diameter 1, 2 and 3 mm respectively (solid lines), plotted as a function of height (adapted from Zipser and Lutz, 1994).

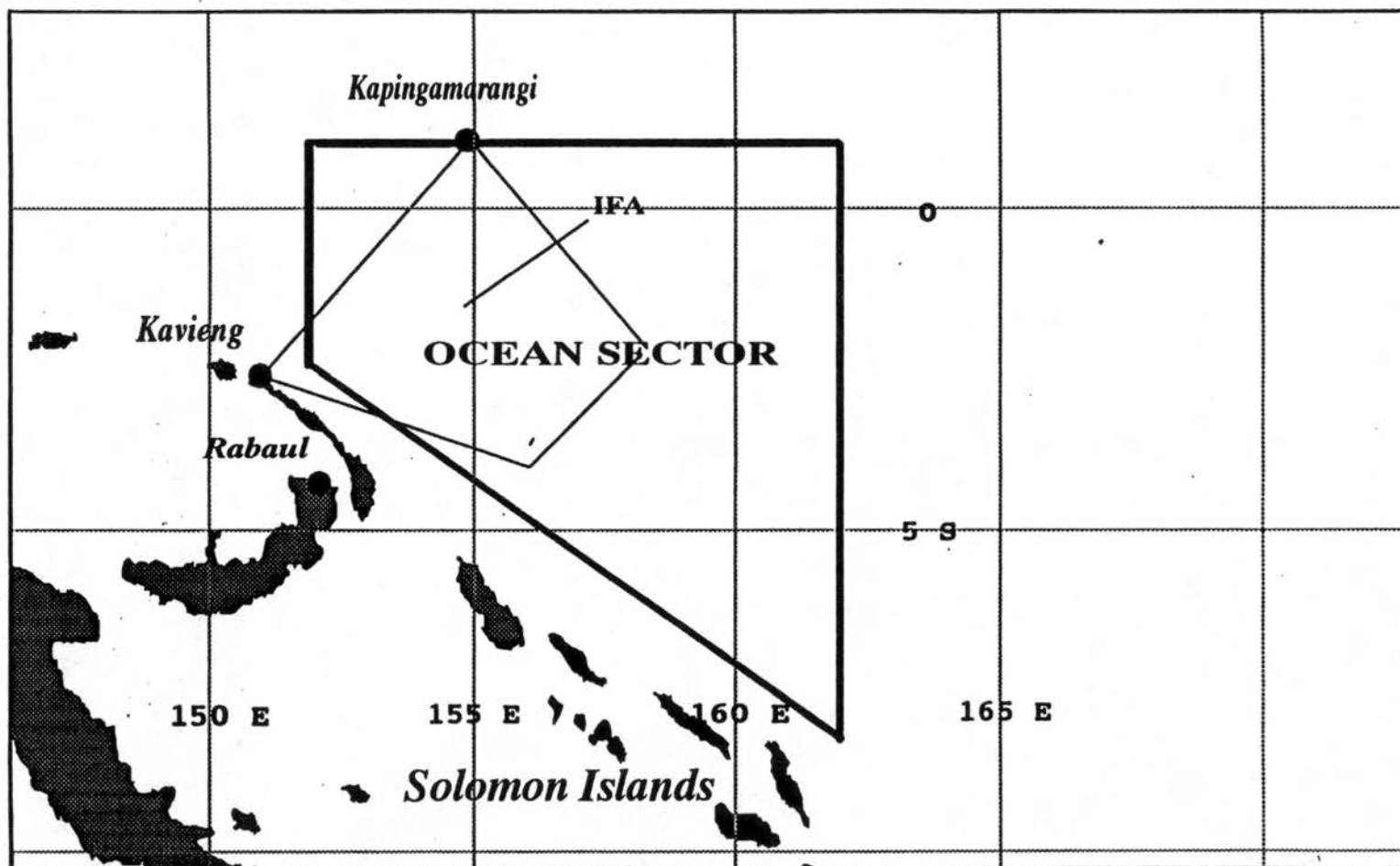


Figure 2.5 Map of the "Ocean Sector" used in diurnal cycle computations of cold-cloudiness and CG lightning relative to the COARE IFA and the ALDF antennas (dark circles).

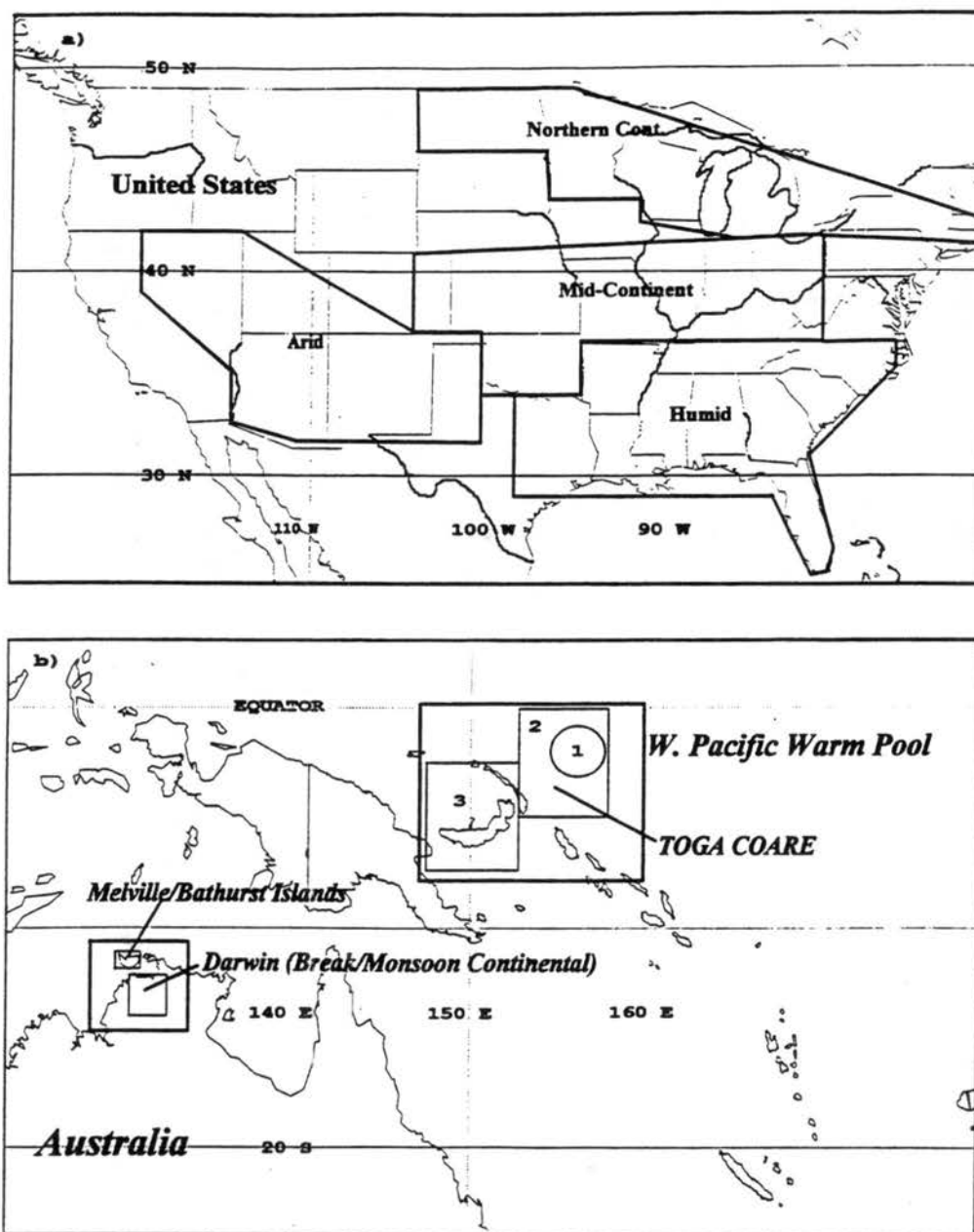


Figure 2.6 a) Map of partitioned rainfall regimes for the continental U.S. based on rain-yield computations; b) as in (a) but for regions of the tropical western Pacific. COARE areas 1-2, and 3 represent oceanic and mixed island oceanic regimes respectively.

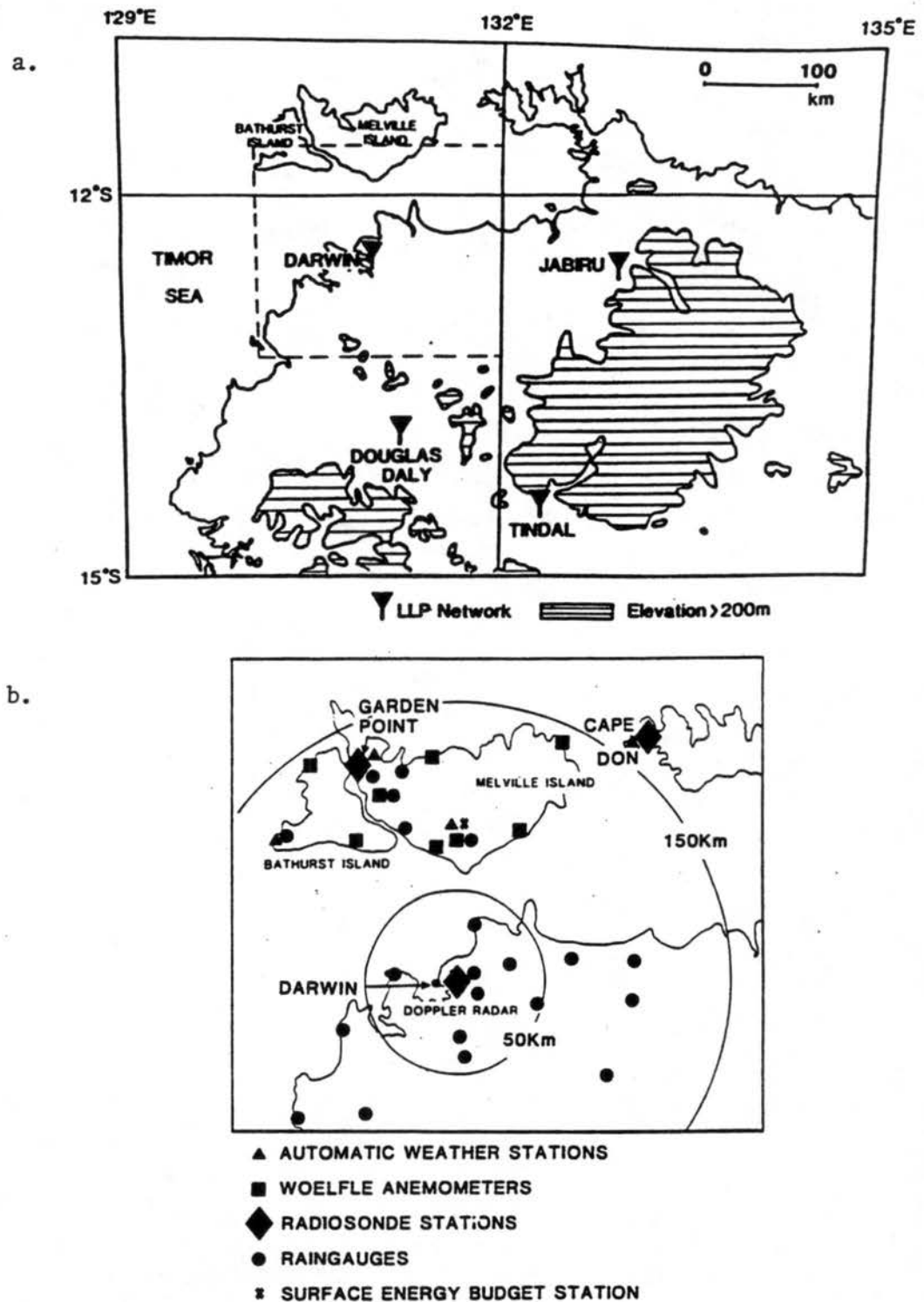


Figure 2.7 a) Map of the DUNDEE operations area. Location of four DF antennas (LLP network) and elevated terrain (hatched) are indicated. b) Map of ITEX operations area enclosed in the dashed box of (a), and centered on Darwin. Locations of TOGA radar, raingauges and other meteorological instrumentation are indicated.

CHAPTER 3

LIGHTNING OBSERVATIONS AND LARGE-SCALE ATMOSPHERIC PROCESSES

Goodman and Christian (1993) suggested that observations of integrated global lightning activity might provide a “signature of changing patterns of tropical moisture and energetics”. In this Chapter we address a subset of the “global-scale” relationship postulated by Goodman and Christian (1993) by examining the relationship between tropical atmospheric processes and the occurrence of lightning over the western Pacific Ocean warm-pool (WP). We focus on cloud-to-ground lightning observations collected during TOGA COARE as related to large-scale atmospheric variables such as temperature, moisture, CAPE, vertical profiles of thermal buoyancy, wet-bulb potential temperature (θ_w), wind shear, wave-disturbances, cold-cloud area coverage, and diabatic heating profiles.

3.1 Thermodynamic forcing of electrified convection over the warm-pool

Previous observations concerning the relationship between CAPE and total lightning flash rate in a tropical continental regime have been summarized in Rutledge et al. (1992) and Williams et al. (1992) and are shown in Fig. 1.1. The collection of observations displayed in Fig. 1.1 suggest a positive correlation exists (albeit weak) between the peak *total* lightning flash rate (in-cloud and CG components) and CAPE for the tropical continental climate near Darwin,

Australia (DWN). However, the lightning observations collected during TOGA COARE are comprised of only the *CG component* of the total lightning flash rate, and it is well established that the *in-cloud* (IC) flash rate typically dominates the total lightning production of electrified cumulonimbus clouds (e.g., Uman, 1987; Goodman et al., 1988; Rutledge et al., 1992; Carey and Rutledge, 1996). This poses an interesting question regarding the “observability” or “robustness” of a CAPE-lightning relationship if only the CG component of the flash rate is compared to the CAPE.

3.1.1 The variation of CAPE and CG lightning frequency over a tropical continent

We address the “observability” question posed in the preceding section by first examining the relationship between CG lightning frequency and CAPE for the DWN area. This is an appropriate convective regime for assessing the potential of observing a relationship between CG lightning and CAPE given the prolific nature of the lightning over this region of the tropical Maritime Continent (e.g., Williams et al., 1992). If a correlation between the CG flash frequency and CAPE can be identified in this environment, then both a motivation and “baseline” are established for examining the relationship between CAPE and CG lightning over the tropical ocean. Hence, mean daily values of CAPE and CG flash count observed during the wet-season of 1989-1990 were computed from the DUNDEE data-set for the vicinity of DWN. Values of CAPE were calculated using 0000 UTC (0930 L) sounding data taken at Darwin and daily CG lightning flash counts were tabulated in 0.25° longitude bins over a latitude-longitude box situated near DWN (similar to the domain shown in Fig. 2.7a). The binned CG data were filtered (1-step Leise) and then plotted in time-longitude coordinates for the time-period 27 December 1989 to 30 January 1990 (Fig. 3.1). This time-period was selected because the CAPE

and CG lightning data sets were continuous during this period. The time-series of CAPE (smoothed) for the same time-period was then overlaid on the CG data shown in Fig. 3.1.

When examined in this fashion (Fig. 3.1), the results suggest that a strong relationship exists between CAPE and CG flash rate over the tropical continent. The relationship between CAPE and CG lightning shown in Fig. 3.1 is similar to that found by Rutledge et al. (1992; Fig. 1.2), however Fig. 3.1 contains significantly more information than Fig. 1.2. In a temporal sense, the maximum (minimum) in CAPE appears to slightly lead (lag) the maximum (minimum) in CG flash density, and the peaks in CG activity exhibit an 8-10 day periodicity. During the 35 day sample, one pronounced monsoon period occurred over the DWN region from approximately 8-16 January 1990 (days 49-57 in Fig. 2.1). Consistent with the arrival of the monsoon trough over DWN, moist westerly winds dominated the lower troposphere, the CAPE decreased to values of 200-500 J/kg, and the CG lightning activity was reduced by a factor of 10^{-10^2} . If the spatial distribution of the lightning is considered in Fig. 3.1, it becomes obvious that topographical influences exert a strong control over the spatial distribution of CG lightning (and by proxy, convective intensity). For example, if the zonal trends in CG flash count are examined relative to Fig. 2.7a, it is clear that the CG flash count decreases (increases) as one progresses westward (eastward) toward the ocean (higher terrain) located west (east) of $\sim 130^\circ\text{E}$.

Given that the CAPE, total and CG lightning flash counts vary together in a relatively consistent fashion over the tropical continent, a reasonable hypothesis regarding the relationship between CAPE and CG lightning over the warm-pool would be that *the electrical intensity of warm pool convection, as inferred by the number of CG's produced in a given time period, should be correlated to the CAPE*. To investigate this hypothesis, sounding data for non-raining periods collected aboard the R/V Vickers during the time period 1/30/93 - 2/26/93 were used to compute CAPE at six hour intervals. Recall from Sec. 2.1.3 that the CAPE calculations were

limited primarily to data collected aboard the R/V Vickers since the boundary layer moisture measurements were of high quality (Dr. Jens Meywerk, personal communication). For the purposes of analysis, the temporal trend in the CAPE computed from the R/V Vickers sounding data is assumed to be representative of an area extending 300 km in radius from the nominal position of the R/V Vickers (Fig. 2.2a)- the extreme limit of MIT radar coverage. The diameter of this area is on the extreme end of the mesoscale but smaller than the length scale $O[1000 \text{ km}]$ of a wave disturbance (e.g., the 4-5 day oscillations discussed in Numaguti et al., 1995). It should also be noted that this area is similar to that examined in Fig. 3.1. In order to compare values of CAPE to CG lightning flash count, CG flash counts were also computed over the same 300 km radius area.

3.1.2 CAPE, θ_w , and moisture variation during TOGA COARE

Sounding data collected aboard the R/V Vickers were used to calculate CAPE (eq. 1.1) by pseudo-adiabatically lifting a parcel with the thermodynamic properties characteristic of the lowest 50 mb (500m, the approximate depth of the mixed-layer) of each sounding from the LFC to the EL. Ice processes have been neglected in this calculation. For several soundings with missing data at upper levels but good data in the boundary layer, we calculated values of CAPE from a regression equation (based on approximately 100 soundings) developed to correlate the mean mixed-layer (50 mb deep) mixing ratio (w_m ; g/kg) to the CAPE (Fig. 3.2):

$$\text{CAPE} = 748(w_m) - 1.1538 \times 10^4 \quad (3.1)$$

The correlation between the two variables was high ($r=.95$), and therefore it is reasonable to calculate missing CAPE values in this fashion. The resulting calculations yielded an overall mean CAPE of approximately 1800 J kg^{-1} for the entire 27 day time period with a standard

deviation of approximately 600 J kg^{-1} . During convectively active periods (associated with more lightning) the CAPE often exceeded 2500 J kg^{-1} . During one six day period of little or no convective activity in the vicinity of the Vickers (early February, 1993), the mean CAPE was lower than average with a mean of approximately 1000 J kg^{-1} . These results are comparable to previous estimates of CAPE made for other tropical oceanic environments (Zipser and LeMone, 1980; Williams and Renno, 1993).

From eq. (3.1) we calculated a zero-CAPE intercept corresponding to a mean mixed layer mixing ratio (MLMR) of 15.4 g kg^{-1} . Interestingly, if one uses this value of MLMR and the mean mixed layer potential temperature of $\sim 301 \text{ K}$ (calculated by compositing all non-raining soundings), a wet-bulb potential temperature (θ_w ; Bolton, 1980) of approximately 22.6°C is obtained at the zero-CAPE intercept. This θ_w is very similar to the zero-CAPE θ_w intercepts obtained by Williams and Renno (1993) for soundings taken over many other tropical locations. When the correlation coefficient between the MLMR and θ_w was calculated, a strong positive correlation was found to exist ($r=0.98$), suggesting that moisture variations in the oceanic boundary layer had a pronounced effect on the value of θ_w . Indeed, when the computed θ_w was compared to both the PBL mean temperature and the MLMR (Figs. 3.3a-b), variations in MLMR were found to exert the strongest control on the mean mixed layer θ_w . Further, when the mixed layer θ_w was compared to CAPE (Fig. 3.4), a strong correlation was found to exist between θ_w and CAPE ($r=.95$). This result is internally consistent with the relationship between MLMR and CAPE (Figs. 3.2-3.3b). It is interesting to note that the change in CAPE with respect to θ_w (slope, $1179 \text{ J/kg/}^\circ\text{C}$) over the warm-pool is very similar to the slope of $1150 \text{ J/kg/}^\circ\text{C}$ reported by Williams and Renno (1993) for Ponape Island ($7^\circ \text{ N } 158^\circ \text{ E}$), also located in the tropical western Pacific Ocean.

The observed sensitivity of CAPE, and θ_w to the MLMR in the tropical atmosphere is supported theoretically by considering an extension of the scaling analysis discussed in Sec. 1.1 between CAPE, PBL temperature and PBL humidity (e.g., Mapes and Houze 1993). Recall from Sec. 1.1 that an increase or decrease in CAPE of $\sim 1000 \text{ J kg}^{-1}$ is approximately equivalent to a 1°C (6%) increase or decrease in PBL temperature (relative humidity). We can extend this scaling analysis to include changes in mixed-layer mixing ratio using the following formulation:

$$\frac{\partial \text{CAPE}}{\partial w} = \frac{\partial \text{CAPE}}{\partial RH} \cdot \frac{\partial RH}{\partial w} \quad (3.2)$$

$$\frac{\partial RH}{\partial w} = \frac{\partial \left(\frac{w}{w_s} \cdot 100 \right)}{\partial w} = \frac{100}{w_s} \quad (3.3)$$

with w = MLMR, RH = relative humidity, and w_s = saturation mixing ratio. Assuming that the relationship in eq. (3.2) holds for a mixed atmospheric boundary layer, and that the change in CAPE with relative humidity is given by the Mapes and Houze relationship (1000 J kg^{-1} per 6% change in RH), we need only calculate the term on the right hand side of eq. (3.3). If w_s is fixed at the mean PBL temperature of $\sim 26^\circ\text{C}$ (taken from the mean COARE sounding, 50 mb deep layer), this implies a mean value of PBL w_s of $\sim 22 \text{ g kg}^{-1}$. Hence, eq. (3.3) has a value of approximately $5\%/(g \text{ kg}^{-1})$ and this implies for eq. 3.2 that,

$$\frac{\partial \text{CAPE}}{\partial w} = \frac{830 \text{ J kg}^{-1}}{g \text{ kg}^{-1}} \quad (3.4)$$

Note that eq. (3.1) suggests a slope for the CAPE- θ_w relationship of $\sim 750 \text{ J kg}^{-1}/(\text{g kg}^{-1})$ which is similar to that diagnosed in eq. (3.4). Therefore, observations such as that shown in Fig. 3.2 which relate the sensitivity of CAPE to mixing ratio are supported by the simple scaling analysis of eq. (3.2). Finally, if the observed or scaled (e.g., eq. 3.4) variation of mixing ratio with CAPE is considered relative to the variation of CAPE with θ_w , the resulting variation in PBL θ_w with MLMR is $O[1^\circ\text{C}/1 \text{ g kg}^{-1}]$.

3.1.3 The variation of CG lightning and CAPE over the warm-pool

To examine the relative trends in CAPE and CG lightning over the sample period, time series of smoothed 24 hour average CAPE's and smoothed 24 hour CG flash totals (flashes that occurred in a 300 km radius circle centered on the R/V Vickers), were created (Fig. 3.5; note that other aspects of this figure such as the moisture convergence will be discussed in the following section). Similar to the results shown for DWN (Fig. 3.1), Fig. 3.5 indicates a reasonable correlation ($r=0.77$) between the CAPE and the 24 hour CG flash total over synoptic-scale time frames (i.e., time periods of several days). That is, cyclic increases or decreases in CAPE over the warm pool modulate to some extent, the occurrence or intensity of lightning producing convection. It is important to note that the actual fraction of clouds producing lightning in a given cloud-ensemble is a relative unknown. Radar echo statistics discussed in DeMott (1996) suggest that approximately 1-10% of the raining cloud population sampled by radar at any given time (and over the same period/area the lightning was observed), exhibited a vertical structure consistent with that of lightning-producing convection (30 dBZ contour above 6.5 km; discussed in Chapter 4). However, it would be a mistake to interpret the relatively small fraction (e.g., 1-10%) of potential lightning-producing clouds as being negligible. This is because rainfall

statistics reported in DeMott (1996) suggest that approximately 20-30% of the total convective rainfall sampled in the vicinity of the R/V Vickers was produced by the same 1-10% of the echo distribution that exhibited “lightning-producing” characteristics (consistent with studies such as Janowiak et al., 1994; Liu et al., 1995). Hence, the small fraction of lightning producing convection in a cloud ensemble is non-negligible from a rainfall and/or a convective heating standpoint.

If variations in the lightning activity are representative of variations in convective cloud intensity (over some fraction of the ensemble), or perhaps variations in total convective cloud area coverage, the significant variation in both CG lightning and the CAPE suggests a relative lack of “equilibrium” (or a lag) between the production of the CAPE and its consumption by convection. Indeed, some storage of CAPE occurred in the 5-8 February time-frame when values of CAPE were rapidly increasing (e.g., $\frac{\partial \text{CAPE}}{\partial t} > 0$) but convective activity (as indicated by radar and lightning activity) remained minimal. During this time period the atmosphere was “capped” by the presence of a mid-level intrusion of warm, dry air (e.g., Yoneyama and Fujitani, 1995).

If we consider the relationship between θ_w and CAPE shown in Fig. 3.4, we can infer from Fig. 3.5 that only a small change in the mean mixed layer θ_w (recall the slope of 1179 J/kg/°C θ_w) or MLMR (recalling eq. 3.4) is needed to produce a marked change in the electrical intensity of convection over the warm-pool. Indeed, the sensitivity of electrical intensity to θ_w is verified in Fig. 3.6 which compares daily mean values of θ_w to the numbers of CG flashes that occurred near the R/V Vickers (300 km radius). The mean θ_w (~24.2°C) for the points plotted in Fig. 3.6 is indicated by the bold arrow, along with the “best-fit” exponential curve. If we use 1800 J/kg as the mean CAPE, and 2300 J/kg as an “active” CAPE (active electrically based on Fig. 3.5),

we find from Fig. 3.4 that this is equivalent to a 0.5°C increase in θ_w . In Fig. 3.6, a 0.5°C increase in the PBL mean θ_w from the “mean” to “active” storm environment corresponds to a factor of 5 increase in CG lightning activity. While the sample represents only 30 days of data collection over the warm-pool, these results are strikingly similar to sensitivities previously noted between θ_w and lightning frequency at other locations in the tropics (e.g., Williams 1992, 1994; Jayaratne 1993).

3.1.4 The influence of moisture convergence and surface latent heat fluxes

In Sec. 3.1.2 it was demonstrated that variations in boundary-layer moisture were highly correlated to variations in CAPE. In turn, variations in CAPE produce variations in lightning activity (e.g., Figs. 3.1, 3.5-3.6). These observations suggest that the mechanisms responsible for altering boundary-layer moisture values over the warm-pool are also implicitly related to changes in the observed lightning flash density. Examples of processes that can alter the PBL specific humidity include surface moisture or latent heat (LH) fluxes, moisture flux convergence, convective downdrafts and rainfall, and turbulent entrainment of dry air through the top of the mixed-layer. Since we are considering changes in PBL moisture content that *allow* deep convection to occur, drying/moistening tendencies associated with convection and/or rainfall are not explicitly considered since they are the “result” we are seeking.

Enhanced turbulent entrainment and/or mixing of relatively dry air through the top of the boundary layer is an important process to consider in the moisture budget of the boundary layer (e.g., Raymond, 1995) and may partially explain the reduction in CAPE (also convection and lightning) observed during the week of 3-9 February 1993 (Fig. 3.5). During this time period weather in the IFA was under the influence of a dry subsiding airmass associated with a

pronounced region of low specific humidity ("dry-tongue") that advected southward across the equator (e.g., Yoneyama and Fujitani, 1995). This "dry tongue" event was associated with pronounced inversions of both temperature and humidity above the boundary layer that peaked near the 700-800 mb levels (e.g., Yoneyama and Fujitani, 1995; Mapes and Zuidema, 1996). The extreme moisture and temperature inversion provided both a strong "cap" over the boundary layer, thus suppressing cumulus convection (e.g., DeMott, 1996), and a coincident drying of PBL air (e.g., Numaguti et al., 1995) through turbulent mixing (including transports by shallow cumulus) across the top of the mixed layer (mixing ratios were typically depressed 2 g kg^{-1} below the mean).

Analyses by Yoneyama and Fujitani (1995) suggest that the dry tongue gradually advected south and east of the IFA, allowing the boundary layer θ_w (and also the CAPE) to steadily recover as surface LH fluxes replenished the boundary layer specific humidity (e.g., Lin and Johnson, 1996 a,b). Even though surface LH fluxes replenished the PBL moisture, the boundary layer remained somewhat "capped" under the influence of a temperature inversion until approximately the 9 February. The persistence of the "cap" and relatively high mixed-layer moisture values characteristic of the end of the dry tongue period explains the high θ_w 's and low CG flash counts associated with the two "outlier" data points indicated in Fig. 3.6. Hence, θ_w , (i.e., substantial CIN was present at low levels).

In addition to surface LH fluxes, the analysis of Yoneyama and Fujitani (1995) also suggests that enhanced low level moisture convergence was present along the trailing western edge of the dry tongue as it moved southeast. In Fig. 3.5, the weekly mean 1000 mb moisture convergence is also indicated in addition to the lightning and CAPE. The moisture convergence data points plotted in Fig. 3.5 were computed for a $1^\circ \times 1^\circ$ grid box centered on the nominal

position of the R/V Vickers (2°S 156°E) using COARE gridded sounding data (supplied by P. Cieselski, CSU). Note that trends in the moisture convergence vary nearly in unison with those of CAPE and CG lightning. Available observations suggest that surface LH fluxes were responsible for the initial recovery of the PBL specific humidity, while increases in moisture convergence occurred during a period of increased deep convection, CAPE and CG lightning (e.g., Fig. 3.5). This seems consistent with studies such as Waliser and Graham (1993) and Zhang (1996), who suggested that LH fluxes tend to slightly lead increases in high cloudiness (a proxy for deep convection) over the western Pacific Ocean, while low-level mass and moisture convergence tend to increase during peaks in convection.

If we now compare a time-series (Fig. 3.7) of surface meteorological data (temperature, SST, and relative humidity [RH]) collected at the location of the IMET buoy (Fig. 2.3 a) to the occurrence of CG lightning within 120 km of the Vickers for the entire 60 day period of COARE ALDF coverage (1 January - 28 February, 1993), the phase of the surface RH and temperature extrema relative to local enhancements in the CG lightning flash density can be examined. Note in Fig. 3.7 that the SST and air temperature were approximately 180° out of phase with the fluctuations in RH, which in-turn varied approximately in-phase with the CG lightning activity in the vicinity of the IMET. Consistent with the studies of Graham and Barnett (1987), Waliser and Graham (1993), and Zhang (1996), peaks in both the SST and surface air temperature tended to lead (or follow) peaks in RH associated with the occurrence of convection (also lightning), with peak SST's and air temperature occurring during convectively quiescent periods. Note that with the exception of the first part of January 1993 (days 367-385), SST variations were typically very small, being less than 0.5°C at a depth of 0.45 m. Measurement of the SST at a depth of 0.05 m during the month of February (considered a "disturbed" period by Lin and Johnson, 1996a) at the location of the R/V Vickers (not shown) revealed a similar trend to that

observed at the IMET location (diurnal variations of $O[0.5^{\circ}\text{C}]$). Further, no correlation between the SST and the amount of CG lightning was observed within the area of study.

The approximate phase coherence of the relative humidity and lightning (and the shifted phase of both the temperature and SST relative to the humidity) are consistent with the results of Zhang (1996) who noted a similar phasing between high cloudiness, specific humidity, temperature and SST over the western Pacific Ocean. Over the warm-pool, both LH fluxes (e.g., Lin and Johnson, 1996 a,b) and moisture convergence (Fig. 3.5) will work in concert to enhance the intensity and/or amount of deep convection. The analysis shown in Fig. 3.5 suggests that local increases in moisture convergence (vs. LH fluxes; Lin and Johnson, 1996 a,b) are particularly well correlated to *coincident* increases in CAPE and CG lightning. Therefore, lightning flash density may be a useful remote sensing tool for 1) identifying the location and relative magnitude of moisture convergence maxima; and 2) identifying regional maxima in convective rainfall (discussed further in Chap. 6), since column integrated moisture convergence is directly proportional to surface rainfall through the "P-E" term in water budget equations (neglecting storage; e.g., Piexoto and Oort, 1993).

If lightning flash density is related to enhancements in moisture convergence over the ocean, then regional peaks in the mean flash density should correspond to regional maxima in moisture convergence. To test this hypothesis, two separate aspects of the problem were investigated. First, weekly mean flash counts and locations (latitude, longitude) were computed for the ocean sector (Fig. 2.5), for the months of January-February, 1993. Next, the mean CG flash counts were examined relative to 1) the moisture convergence value at the location of the mean CG flash position; and 2) the distance between the weekly mean CG flash positions and the location of 1000 mb moisture convergence maxima. Several examples of the weekly gridded mean moisture convergence and the weekly mean CG flash position are shown in Figs. 3.8 a-c.

These particular examples correspond to weeks associated with relatively active “electrical” conditions in and around the COARE ocean sector.

Figures 3.8 a-c suggest that major increases in moisture convergence and CG lightning were encountered to the south and east of the COARE IFA. Similar trends were also noted in the mass convergence field (not shown). The locations of the relative maxima of the two variables are consistent with the more general location of the South Pacific Convergence Zone (SPCZ). Indeed, when the entire 60 days of CG lightning data were examined for the *ocean sector* (not shown), the majority of the CG lightning activity was situated over oceanic regions located east and southeast of the center of the IFA. The area situated just to the southeast of the IFA also corresponds to the location of the heaviest mean daily rainfall total diagnosed from sounding moisture budgets for the COARE IOP (Lin and Johnson, 1996b). Interestingly, the moisture convergence patterns at 900 mb (Figs. 3.9 a-c) were most similar to those observed at 1000 mb (Figs. 3.8 a-c) when the CG flash densities were collocated with moisture convergence maxima (i.e., near strongly forced convection), suggesting that the moisture convergence may have occurred through a deeper layer of the lower troposphere in areas of lightning producing convection.

When the values of 1000 mb moisture convergence were compared to the weekly locations of maximum CG flash density, a non-linear tendency (albeit weak) for increased values of moisture convergence to occur with increases in flash count was noted Fig. 3.10a). Further, when the distance between the locations of weekly mean moisture convergence and CG flash density maxima was plotted as a function of the weekly mean flash count (Fig. 3.10b), there was a tendency for larger CG flash counts to occur closer to the location of maximum moisture convergence (and heaviest rainfall; e.g., Lin and Johnson, 1996b). The results of Figs. 3.9-3.10 are consistent with the observed variation of CAPE, CG flash density and moisture convergence

shown in Fig. 3.5. Moreover, the correlations between moisture convergence, CAPE and CG lightning suggested by Fig. 3.5, and Figs. 3.8-10, support the more general finding that CAPE and θ_w (and hence lightning) are largely controlled by the value of the MLMR over the warm-pool (e.g., Figs. 3.2-3.6).

3.1.5 Vertical profiles of thermal buoyancy as a function of observed electrical intensity

We have demonstrated that the CG lightning flash density and the *magnitude* of the CAPE are correlated for both tropical continental (Fig. 3.1) and tropical oceanic climate *regimes* (Fig. 3.5). Note that the CAPE values computed for the tropical continent and the warm-pool are approximately the same, and even similar to mean CAPE values observed over the continental U.S. (e.g., Weissman and Klemp, 1986). Yet, lightning flash densities observed over *continental* locations are typically an order of magnitude larger than those observed over the tropical oceans (e.g., Orville and Henderson, 1986; Rutledge et al., 1992; Williams et al., 1992; Zipser, 1994). This point is further illustrated in Fig. 3.11, where CG flash densities over the tropical oceanic domain of COARE are directly compared to flash densities computed for the tropical continental/island regions of DWN, the sub-tropical Gulf Stream, Grand Bahama and Andros Islands, and the mid-latitude continental U.S. Clearly the CG flash density over the warm-pool is substantially lower than that observed for either the tropical or mid-latitude continental regimes (including islands and the Gulf Stream in the sub-tropics).

Care must be taken when attempting to meteorologically interpret different CG flash densities for areas located at different latitudes. Studies such as Williams (1992, 1994) and Mackerras and Darveniza (1994) have suggested that the *annual total* (IC and CG components) lightning flash densities over *continental* locations increases with latitude. At first this may seem

inconsistent with the results of Fig. 3.11, which suggest that higher *CG* flash densities occur over the *continental* U.S. However, it should be noted that 1) flash densities indicated in Fig. 3.11 for the mid-latitude continent/sub-tropical islands were computed for the summer convective season (June-August, 1994, 1995), when a relative *maximum* in electrical activity occurred; 2) the intra-cloud (IC) to CG flash ratio observed in *continental* regimes has been observed to increase slightly with decreasing latitude (e.g., by a factor of two from the mid-latitudes to near-equatorial regions; Mackerras and Darveniza, 1994); and 3) regardless of the latitude or IC to CG ratio, the *total* flash rate over the tropical ocean has been observed to be a factor of 10 lower than that observed over the continents (e.g., Orville and Spencer, 1979), consistent with Fig. 3.11. Hence the decrease in *CG flash density* over the tropical oceans, relative to that of the continents in Fig. 3.11 is consistent with the overall decrease in total flash density observed over the ocean. The ~50% decrease in CG flash density noted between the tropical continental (tropical island) and the mid-latitude continental (sub-tropical island) CG flash densities may be due to a combination of sampling bias (e.g., sampling the mid-latitudes at strong peak in activity; Williams, 1994; Mackerras and Darveniza, 1994), and slight latitudinal gradients in IC-CG ratio. The island regimes (e.g., Bahamas and Melville/Bathurst Island) seem to represent “middle-ground” with regard to electrical activity when compared to continental and oceanic regimes sampled in the same latitudinal belt and over the same period of time. Hence we conclude that differences in *CG* flash density observed between continental, island and oceanic regimes, largely reflect differences in *total* electrical activity *between* those regimes.

Recall that differences in the magnitude of the CAPE as a function of regime (e.g., active tropical oceanic vs. mid-latitude continental) could not explain the differences observed in CG flash density. However, differences in electrical intensity as a function of cloud-kinematics and the large-scale environment, can also be interpreted within the framework of different regime-

dependent vertical profiles of thermal buoyancy (e.g., the “shape of the CAPE”; Williams 1995). For example, the *total* buoyancy of a cloud-parcel is directly related to the vertical velocity (w) through the vertical momentum equation (representing frictional eddy diffusion and convergence effects as “F”);

$$\frac{dw}{dt} = -\frac{1}{\rho} \frac{\partial p'}{\partial z} + B + F \quad (3.5)$$

where

$$B = \frac{T'}{T_e} - \frac{p'}{p_e} + 0.61q'_v - q_h \quad (3.6)$$

In eq. (3.6) B = total buoyancy, T' , p' , and q'_v represent perturbation temperature; pressure and water vapor mixing ratio respectively, q_h =hydrometeor mixing ratio, and p_e , T_e are the environmental pressure and temperature, respectively. A comparison of the perturbation temperature and pressure terms suggests that the pressure term can be neglected if either the velocity scale of motion is much less than the speed of sound (e.g., Emanuel, 1994), or the pressure of the environment is much less than that typically encountered at the tropical tropopause (~70-80 mb). For the sake of discussion, we will neglect the buoyancy perturbation pressure term and discuss the perturbation temperature and vapor terms relative to the “hydrometeor-loading” term. Hence, the thermal buoyancy (B_t) can be represented by the temperature and water vapor perturbation terms in eq. (3.6), which when combined, are equivalent to the virtual temperature excess represented in the integrand of eq. (1.1).

If we examine the partial derivative of eq. (3.5) with respect to height (z), and explicitly consider the contribution to the acceleration by only the thermal buoyancy force, the following relationship results:

$$\frac{\partial}{\partial z} \left\{ \frac{dw}{dt} \right\} = \frac{\partial B_t}{\partial z} - \frac{\partial q_h}{\partial z} \quad (3.7)$$

Equation (3.7) states that in order for parcel acceleration to increase with height, the right hand side of eq. (3.7) must be greater than zero. If the hydrometeor loading derivative dominates the right hand side of (3.7), the parcel will eventually decelerate with respect to height, leading to a reduction in updraft speed and the eventual formation of a downdraft. It is also important to note that entrainment of environment air should be included in the T' and q_v' terms through mixing in the thermodynamic energy equation and dynamic entrainment forced by mass continuity. Clearly, it is advantageous to have an increasing thermal buoyancy profile with height to enable the updraft to counteract hydrometeor loading and entrainment.

In tropical oceanic clouds, the distribution of thermal buoyancy profile with respect to height is often characterized as having small thermal buoyancy's distributed over a greater depth of the troposphere relative to that observed over continental locations (e.g., Fig. 1.6; Lucas et al., 1994; Williams 1995). Hence, reductions in the amount and slope of B_t with respect to height (e.g., Fig. 1.6), larger sub-cloud layer mixing ratios (a "causative" agent for coalescence), and a deeper layer of cloud below the freezing level coupled with efficient warm rain processes (e.g., Jorgensen and LeMone, 1989), all combine to suppress the formation of a robust precipitation ice phase in tropical oceanic clouds (or monsoon clouds over a tropical continent; e.g., Williams et al., 1992). However a robust precipitation ice phase is a microphysical condition considered necessary in order for significant electrification of the cloud to occur (Williams, 1995).

We examined the importance of the "shape of the CAPE" over the tropical ocean, by partitioning R/V Vickers sounding data into electrically "active" and "inactive" periods, compositing the soundings from each period, and subsequently computing CAPE and thermal buoyancy profiles for each period. Sounding profiles classified as electrically "active" ("inactive") were comprised of non rain-contaminated soundings taken on days that the CG flash density was greater (less) than the mean CG flash density for the month of February (7

flashes/ 10^4 km²/day; the mean for January was 3.5 flashes/ 10^4 km²/day). Note that the term “inactive” does not imply that *no* lightning occurred, and it certainly doesn’t imply a lack of deep convection and rainfall (as heavy rainfall was often observed in the complete absence of electrical activity during COARE). The vertical profiles of B_t were then compared to pre-convective soundings from other locations (e.g., ITEX soundings taken in the late morning over the islands of Melville and Bathurst near DWN, PRE-STORM mid-latitude continental soundings taken in the vicinity of approaching convective systems). The resulting vertical profiles of B_t are shown in Fig. 3.12.

The CAPE for the COARE “active” (“inactive”) period was 2400 J kg⁻¹ (~1700 J kg⁻¹) and the values of CAPE for the DWN-Island and PRE-STORM soundings were 2600 J kg⁻¹ and 2500 J kg⁻¹ respectively. It is important to note that the magnitude of the CAPE for the COARE “inactive” sounding is still substantial, albeit 700 J kg⁻¹ less than the “active” CAPE. Indeed, the COARE “inactive” CAPE of 1700 J kg⁻¹ falls in the range cited by Weissman and Klemp (1986; 1500-2500 J kg⁻¹) as being “typical” for midlatitude thunderstorms (including severe thunderstorms). The CAPE values computed for the DWN-Island and PRESTORM profiles are only slightly larger (100-200 J kg⁻¹) than the COARE “active” period. However, there are substantial differences in CG flash density between the COARE “active”, tropical continental/island, and mid-latitude convective regimes.

For example, even during the COARE “active” periods, the *absolute peak* in flash density over the 300 km radius circle surrounding the R/V Vickers was 63 flashes/ 10^4 km²/day (similar to the *mean* CG flash density for Melville and Bathurst Islands; Fig. 3.11). The mean “active” flash density was actually 16 flashes/ 10^4 km²/day for the January-February, 1993 time-frame, and 20 flashes/ 10^4 km²/day if only the month of February is considered (which is the month the Vickers sounding data was collected). Note that the mean flash densities (i.e., we are not

considering the most active times electrically) for virtually all of the other locations in Fig. 3.11 were a factor of 3-10 larger than the most *active* periods observed over the warm-pool. Even if we allow for a factor of two increase in the flash density due to a reduced detection efficiency in the COARE ALDF network (i.e., assume the detection efficiency is only 50% of the detection efficiencies of all the other networks- an extreme assumption), the flash density in the COARE “active” period is still not as large as the mean flash density observed at other continental locations.

Given the minor differences between the CAPE of the COARE “active” regime and that of the DWN-Island and/or PRE-STORM regimes (equating to a 3 ms^{-1} difference in the maximum possible updraft speed), we pose the question: How is the CAPE related to the observed differences in CG (or total) flash density *between the different convective regimes*? It is certainly possible that moderate differences in CAPE (or θ_w) could exert some controlling influence on the relative electrical intensities (e.g., Fig. 3.5). However, it doesn’t appear that such a slight difference in magnitude (e.g., 100 J kg^{-1} at CAPE’s $\approx 2500 \text{ J kg}^{-1}$) could *entirely* account for the observed differences in CG flash density. We therefore turn to the B_t profiles shown in Fig. 3.12 to explain the difference in electrical activity between the storms as a function of regime. Significant differences that exist between each of the respective profiles in Fig. 3.12 can be summarized as follows:

- 1) Relative to the PRE-STORM B_t profile, positive B_t ’s in the tropical soundings all span greater depths of the troposphere ($\sim 800 \text{ mb}$ in the tropics vs. $\sim 600 \text{ mb}$ for the PRE-STORM profile).
- 2) The first primary inflection point in the PRE-STORM profile (first point above the LFC where the B_t increases at a much slower rate with height or becomes constant) is located between the -10°C and -15°C isotherms (510-460 mb). The first primary

inflection point in *all* of the tropical soundings occurs near the 580-560 mb level, at or below the height of the freezing level (+5°C to 0°C). Hence the positive vertical slope of B_t (with respect to pressure) for the PRE-STORM soundings extends to a much lower temperature than that of the tropical soundings.

- 3) The *peak* B_t in the PRE-STORM profile is a factor of 2, 1.6 and 1.4 larger than that of the COARE “inactive”, COARE “active”, and ITEX/DWN peak B_t ’s respectively. The mean vertical *slope* of the PRE-STORM profile below the first primary inflection point is larger than those of the tropical “inactive”, “active” and ITEX/DWN profiles by a factor of 2.2, 1.7, and 1.4 respectively.
- 4) The level of free convection in the PRE-STORM sounding is approximately 1.5 km above ground level and the LFC in the tropical profiles is approximately 500-700 m above ground (sea) level (near the top of the mixed-layer).
- 5) Consistent with trends in the height of the LFC, the PRE-STORM sounding contains more CIN ($\sim 15 \text{ J kg}^{-1}$) than the tropical soundings ($3\text{-}7 \text{ J kg}^{-1}$).

The differences in vertical structure noted between the buoyancy profiles in Fig. 3.12 are consistent with the idea that the vertical distribution of B_t exerts a very large control on the electrical intensity of the clouds (as reflected by differences in CG-flash density *between the regimes*). For example, if we first consider the observed differences (points 1-5 above) between tropical oceanic “active” and mid-latitude continental convective regimes (the “extremes” of the sample), we see that even though the CAPE’s are similar, the distribution of buoyancy in the mid-latitude case (larger peak and vertical slope of buoyancy at colder temperatures in the cloud), favors a more electrically active cloud. This is because the more “robust” mid-latitude B_t profile is: a) better able to counteract entrainment and water loading effects with larger thermal buoyancy’s; b) more conducive to “explosive” growth of the cloud (especially at higher CINS’s,

typical of the convective environment found in the Central Plains of the U.S.; e.g., Emanuel, 1994); c) more favorable to the development of stronger updrafts at lower elevations in the cloud; d) following (c) able to provide a larger mass of condensate to the mixed phase region of the cloud; and e) more capable of producing a robust precipitation ice phase leading to rapid electrification.

Note in the tropical B_t profiles that there is a tendency for the slope of B_t to become more positive as more electrically active convective regimes are encountered. For example, the COARE “inactive” profile has shallower slope of B_t than the “active” profile, and the “active” profile has a B_t slope approaching that of the DWN-Island profile. In turn, the slope of the ITEX/DWN B_t profile at mid-low levels even closer to the mid-latitude profile. That is, the observations (Fig. 3.12) suggest that a gradual progression in the vertical slope of the mid to low level thermal buoyancy profile occurs as one moves from a less electrically active convective regime, to a highly electrified convective regime, as inferred from CG flash densities. Hence, over the tropical ocean and continent, the *magnitude* of the CAPE appears to be related to the amount of CG lightning observed *within* the regime being considered (e.g., Figs. 3.1, 3.5), but the *vertical structure* of the B_t profile plays a large role in determining the relative flash densities observed *between* climate regimes. This finding is consistent with the hypothesis of Lucas et al., (1994a) regarding differences in the “shape of the CAPE” vs. the magnitude of the CAPE between convective regimes. The degree to which electrified convective clouds in each specific regime can be characterized using tropospheric soundings partitioned by thermal buoyancy profile (and perturbations to those profiles) will be further addressed in modeling studies summarized in Chapter 5.

3.1.6 Diurnal cycles of CG lightning, CAPE and rainfall over the tropical western Pacific

One additional test of the relationship between CAPE, flash rate and the presence of deep, vertically developed convection is to examine the average diurnal cycle of all three variables. To examine the diurnal cycle of CAPE, we composited CAPE's calculated from non rain-contaminated soundings at six hour intervals (00, 06, 12 and 18 UTC) over approximately 30 separate days in the time period (22 January - 26 February 1993) for the R/V Vickers and three other research vessels in the IFA (the R/V's Moana Wave, PRC #5, and PRC #3; Fig. 1). The CAPE data for each vessel were averaged separately and then normalized to the peak CAPE of the composite in order to eliminate biases in magnitude associated with systematic errors in the different ship sounding profiles; *here we wish only to examine the trend in the CAPE.*

Implicit in the CAPE diurnal cycle calculation is the assumption that errors introduced into the composite by sounding data collected aboard the R/V's Moana Wave, PRC #5 and PRC #3 are independent of the diurnal cycle (i.e., the trend, but perhaps not the magnitude, of the true diurnal cycle of CAPE would be preserved). As a check on this assumption, the diurnal cycle of CAPE was also computed from the R/V Vickers data exclusively, with little difference noted in the results between the composite and Vickers CAPE diurnal cycles. Hence, the diurnal cycle of CAPE computed using the four-ship composite, will be the focus of this analysis (Fig. 3.13). To determine the diurnal cycle of cloud-to-ground lightning flashes over the ocean, we calculated the frequency of all CG flashes that occurred over an ocean-sector (shown in Fig. 2.5) for 24 one-hour time bins, which provided a much larger lightning sample. Flashes were counted from a sample of 28 separate days in January and February 1993 (approximately 25,800 flashes).

To examine the diurnal cycle of deep tropical oceanic convection we used areas of low satellite IR brightness temperature (cold-cloud areas) as a proxy for the occurrence of deep convective clouds. Cold-cloud areas were computed over the ocean-sector (Fig. 2.5) for several

different temperature thresholds. First, areas of cloudiness associated with brightness temperatures colder than 263 K, 255 K, 245 K, 235 K, 225 K, 215 K, 208 K, 205 K and 195 K were summed over each one hour period for the time period 30 January - 28 February, 1993. The sums of the areas were then normalized by the total area of the ocean sector used, and the mean normalized area in each temperature bin for the entire time period was calculated. Next, the mean normalized cold cloud areas for each one hour period of the day (same temperature thresholds as above) were calculated and subsequently plotted as a function of the number of standard deviations (Z-score) from the mean normalized area calculated for the entire 30 day time period. The results indicated that weak diurnal cycles in cold-cloudiness were present only for convection with brightness temperatures colder than approximately 225 K (tops near 11.5 km). The diurnal signals were more pronounced for brightness temperatures less than 205 K (i.e., tops ≥ 13.5 km). We have plotted the diurnal cycle of cloud-top brightness temperature <205 K in Fig. 3.13 for each one hour bin as the number of standard deviations from the mean normalized area. To further emphasize the relationship between the diurnal cycle of CG lightning, cold-cloud area and CAPE relative to that of precipitation, we have also added the diurnal cycle of convective and total rainfall to Fig. 3.13 (deviation from the mean over a 90 day time period; e.g., Rickenbach, 1996) for the approximate coverage area ($70,000 \text{ km}^2$) of the MIT radar. Note that the coverage area of the radar-rainfall estimate is smaller than the ocean sector area by a factor of 10, but this is compensated to some extent by the temporal sample of rainfall (90 days) which is larger than that of the CAPE and CG lightning samples by a factor of three.

In Fig. 3.13, the CAPE exhibits a broad, nocturnal to early morning maximum, and a relative minimum at 1200 L. Both the CG lightning and rainfall appear well correlated though the amplitude variation [Max-Min/Mean] of the lightning is much larger than that of the rainfall. Peaks in the lightning and rainfall occur between 0200-0400 L, with broad minima similar to that

of the CAPE centered near 1200L. The peak in the diurnal cycle of CG lightning appears to precede the absolute peak in the diurnal cycle of the CAPE by approximately 3 hours. Note however, that the difference between the CAPE at 2300 L and 0500 L is very small (on the order of 100 J/kg), and that the soundings taken aboard the ships were collected at 6 hour intervals. It is therefore possible that the actual peak in CAPE occurred between 2300 L and 0500 L.

The variation in CAPE over the diurnal cycle is forced by a combination of tropospheric temperature and moisture variations. An example of the diurnal variation in vertical temperature and moisture profiles is displayed in Fig. 3.14. In Fig. 3.14 the temperature and moisture variations are plotted as differences between composited daytime (1100-1700 LST) and nighttime (2300-0500 LST) sounding profiles (R/V Vickers sounding data). The nocturnal (daytime) profiles of moisture and temperature are more moist (dry) and slightly cooler (warmer) in the lower troposphere, with the difference in moisture content between the two profiles being much smaller in the upper troposphere. In general, the daytime profile exhibits a warmer temperature throughout the column as a result of the enhanced net radiational cooling in the column that occurs at night (c.f., Mapes and Zuidema, 1996). There is a slight warming and moistening of the nocturnal sounding profiles relative to the daytime profiles at pressures of 650-500 mb. This is most likely associated with a relative increase in the sampling of nocturnal mid-level cloudiness during nighttime soundings.

The relative difference in CAPE between the day and night hours ($\sim 500 \text{ J kg}^{-1}$) is consistent with the ($\sim 0.5 \text{ g kg}^{-1}$) 0.4°C increase in θ_w (MLMR), observed during the night hours (e.g., Figs. 3.2-4 and 3.14; eqs. 3.1-3.4). Note that the relative difference in mean *surface* temperature between the night and day (for the four composite sounding times) was only 0.2°C . It is interesting to note that if the soundings had been collected over a tropical continental location, one might expect the peak in the CAPE to occur in the afternoon when the *surface* heating by

solar radiation produces a corresponding maximum in surface wet-bulb temperature. Indeed, Fig. 3.15 reveals that a peak in *surface* wet-bulb temperature (measured by a sling-psychrometer aboard the R/V Vickers) did occur in the early afternoon hours (local time) over the tropical ocean. However, as suggested by Fig. 3.13, the CAPE does not peak coincident with the maximum *surface* wet-bulb temperature ($\sim 180^\circ$ out of phase), but does peak coincident with nocturnal maxima in the mean mixed-layer wet-bulb temperature and mixing ratio. Hence, PBL mixed-layer properties appear to dominate the diurnal cycle of CAPE, and perhaps the occurrence of deep convection as indicated by the peaks in CG lightning, rainfall and cold-cloudiness.

With regard to the diurnal cycle of cold-cloud area relative to that of lightning, CAPE, and rainfall, the diurnal cycle of brightness temperatures < 205 K shown in Fig. 3.13 exhibits a weak (Z-score only 0.3) but broad relative maximum in areal coverage near 0500 L, approximately two hours after the peak in the CG lightning and rainfall. The cold cloud area begins a slow decline in coverage after 0900 L reaching a relative minimum in coverage near 1700 L. It is interesting to note that as the CG lightning frequency and rainfall begin to increase near 1700 L, the cold cloud area reaches a minimum and begins to increase again, lagging the trend in CG lightning and rainfall by approximately 1-2 hours. The increase in cold cloud area coverage also seems to lag the increase in CAPE by several hours over the 24 hour period. The weak diurnal cycle observed in the cold cloud area is somewhat consistent with previous studies of the diurnal cycle of precipitation and cold cloud area coverage over the western Pacific ocean (e.g., Gray and Jacobson, 1976; Janowiak et al., 1991, 1995; Mapes and Houze, 1993; Nitta and Sekine, 1994; Chen and Houze, 1996), though it must be emphasized that we have only examined a 30 day period.

The diurnal cycles of CAPE, flash rate, cold cloud area and rainfall suggest that increases in CAPE are associated with increases in CG lightning and associated deep convection (consistent with the increase in cold cloud area coverage and both the convective and total rainfall). The slight difference in phase between the peaks in CG lightning, total and convective rainfall, and the areas of cold cloudiness (cold cloud area lagging lightning frequency and rainfall by two hours), coupled with the observation that the cold cloudiness rapidly increases as the CG lightning and rainfall begin to peak (Fig. 3.13), suggests that the most intensely electrified convection and the heaviest rainfall occur prior to the maximum coverage of the cold cloud area. This suggestion is further supported by the *partitioned* diurnal cycle harmonics of convective (peak ~2 a.m. LST) and stratiform (peak ~5-6 a.m. LST) rainfall *specifically associated with larger organized convective systems* (e.g., linear and non-linear MCS's) over the WP reported in Rickenbach (1996).

While the observations of CG lightning when combined with those of CAPE, cold-cloudiness and rainfall presented in Fig. 3.13, certainly suggest the presence of a nocturnal peak in the diurnal cycle of rainfall/deep convection, and even a relative ordering of specific processes (e.g., coincident peaks in deep convection, lightning, and convective rainfall, followed later by peaks in stratiform rainfall and then cold-cloud area), the observations do not point to any one particular large-scale atmospheric process being responsible for the diurnal cycle. Indeed, the observations in Fig. 3.13 are consistent with a variety of hypotheses advanced to explain the presence of a nocturnal peak in the diurnal cycle of deep convection over the ocean (discussed in Sec. 1.2.1). At this point it is worth restating that the *primary hypothesis* being addressed herein is: *If some process in the tropical atmosphere (yet to be specifically identified) produces a diurnal cycle of deep convection and rainfall, there should also be a signature of the process in observations of the diurnal cycle of lightning.* The lightning observations presented herein

strongly support this hypothesis. Further, the observations suggest that lightning flash counts (apparently quite sensitive to small changes in thermodynamic parameters; e.g. Fig. 3.6) may be extremely useful for identifying the presence and/or phase of the diurnal cycle of deep convection. Note for future reference, that the relative amplitudes of oceanic and continental diurnal cycles of lightning and rainfall will be further examined in Chapter 6 relative to lightning-rainfall relationships.

3.2 Observations of CG flash counts as compared to cold-cloud area, rainfall, large-scale upward motion and apparent heating rates

Based on scaling arguments presented in Sec. 1.1, it is expected that lightning producing cumulonimbi over the tropical ocean should be associated with relatively larger vertical velocities and hydrometeor masses in the mixed phase regions of the clouds (i.e. temperatures colder than 0°C) compared to non-lightning producing cumulonimbi. This suggests the possibility that lightning producing convection might exhibit a characteristic diabatic heating, moistening and/or vertical velocity profile in comparison to non-lightning producing convection. This becomes clear if we consider an approximate form of the equations describing the “apparent” heat source and moisture sink (neglecting horizontal transports and including deposition/sublimation in the condensation-evaporation terms):

$$Q_1 \approx L(C - E) + L_f q_m - \frac{\partial \omega' s'}{\partial p} \quad (3.9)$$

$$Q_2 \approx L(C - E) - \frac{\partial \omega' q'}{\partial p} \quad (3.10)$$

where Q_1 = apparent heat source; Q_2 = apparent moisture sink; p = pressure; C = condensation and deposition (ice); E = evaporation and sublimation (ice); ω' = perturbation pressure velocity;

s' = dry static energy; q' = water vapor mixing ratio; q_m = meltwater mixing ratio; L_f and L = latent heats of fusion and condensation/sublimation (ice) respectively. The first term on the right hand side of eq. 3.9 (3.10) represents the condensational heating (drying) term, while the third term on the right of eq. 3.9 (3.10) represents an eddy flux divergence of heat (moisture). The second term on the right hand side of (3.9) represents heat liberated during the freezing (melting) of water (ice). To the extent that electrified oceanic cumulonimbi are associated with more intense convective circulations, which in turn lead to larger eddy heat flux convergence, condensation, accretion, and evaporation rates (especially above the height of the freezing level), we might expect to see characteristic *regional* convective heating and vertical velocity profiles associated with lightning producing convection (yet to be determined) *if the fractional coverage of deep electrified convection is a large enough percentage of the area integrated convective cloudiness.*

To examine this hypothesis, smoothed daily time-series of CG lightning, rainfall (MIT radar estimate for vicinity of the Vickers; supplied by Dr. T. Rickenbach, NASA GSFC) and daily mean fractional cold-cloudiness (lightning, rainfall and cold-cloud fraction within 150 km of the R/V Vickers) were compared to time-series of pressure vertical velocity (ω ; Fig. 3.17), Q1 (Fig. 3.18a), and Q2 (Fig. 3.18b) as diagnosed and gridded for a $1^\circ \times 1^\circ$ grid square centered on the position of the R/V Vickers (heating/vertical velocity data described in Lin and Johnson, 1996b). First, consider the observed relationship between CG flash count, rainfall, and cold cloud area at two different temperature thresholds (245 K and 208 K; only two thresholds are plotted for clarity, a total of seven were examined in the range 255 K - 195 K), as shown in Fig. 3.16. Figure 3.16 suggests that CG lightning activity varied approximately in-phase with that of the rainfall and cold-cloudiness. However, Fig. 3.16 also suggests that no *particular* cold-cloud temperature can be associated with increases in the daily CG lightning count or the amount of

rainfall, and there is apparently little correlation between the *amount* of cold cloudiness and the amount of CG lightning detected by the COARE ALDF network. A similar result (not shown) was noted between CG flash counts and cold-cloud areas when the two parameters were examined over the much larger ocean-sector shown in Fig. 2.5. Hence, cold-cloud area coverage, at any brightness temperature, does not appear to be a reliable indicator for the *amount* of CG lightning activity or rainfall observed over the *daily* time-scales considered in Fig. 3.16 (c.f., Restivo, 1996).

Now consider the time-height profile of diagnosed daily mean vertical pressure velocity (ω) derived from the COARE gridded sounding data as compared to a coincident time-series of CG lightning flash count and rainfall (Fig. 3.17). Note that the CG lightning, rainfall, vertical velocity, and cold-cloud area (Fig. 3.16) vary approximately in-phase during time periods of significant ω (i.e., $\omega \geq 3\text{-}6 \text{ mb hr}^{-1}$). In this case, the covariation of all four variables suggest the *occurrence* of deep, intense convective events. However, the relationship between the CG flash count, rainfall, cold-cloud area and the *magnitude and/or structure* (note the distinction made from “occurrence”) of the vertical velocity as diagnosed from the gridded sounding data is ambiguous. Similar ambiguities with regard to the *structure and magnitude* of the heating/moistening profiles are also observed when compared to the CG flash counts (Figs. 3.18 a-b; computed by Lin and Johnson, 1996b). As in the case of the vertical velocity (Fig. 3.17), CG lightning activity, rainfall, cold-cloudiness and significant anomalies in heating/moistening appear to *occur* in-phase, but *no particular vertical profile* or structure of heating and/or moistening can be associated with enhancements in CG lightning activity in these observations.

Considering; 1) the characteristic differences in the temporal and spatial scales associated with the occurrence and/or observation of CG lightning (e.g., radar echo statistics suggest that only 1-10% of the cloud ensemble is capable of producing lightning [DeMott, 1996], hence

lightning-producing convection is atypical); 2) the localized nature of heavy convective rainfall in space and time; 3) the effect on cold-cloud area by divergent circulations and upper tropospheric winds which disperse small ice particles produced in convection over large distances in the atmosphere; and 4) the sampling limitations associated with diagnosing Q1 and Q2 over large areas from radiosonde data (e.g., 6-12 hour sounding data; e.g., Lin and Johnson, 1996b); it is not surprising that the observed CG lightning, rainfall, cold-cloud area, heating/moistening and vertical velocity profiles exhibit little correlation over *daily* time-scales with regard to areal coverage (cold-cloudiness) or vertical structure. Similar “ambiguities” with regard to cold-cloud temperature/coverage and vertical velocity were noted by Lin and Johnson (1996b) who suggested that very cold-cloud temperatures over the IFA were useful for indicating persistent upward motion in the uppermost levels of high clouds, but that vertical motions at 700 mb were often independent of cloud top temperature. In fact, Lin and Johnson (1996b) noted that approximately one third of the observed *downward* vertical velocities (same data being used in Fig. 3.16) at 700 mb were associated with very cold cloud, and suggested that this was indicative of the presence of a thick stratiform anvil and associated mesoscale subsidence.

The observations presented herein point out the difficulty in comparing distinctly different types of observations, and/or attempting to interpret the observations in terms of other processes. For example, Price and Rind (1994) used cold-cloud temperature as a proxy for total lightning flash frequency in a General Circulation Model (GCM) to estimate lightning flash density changes in a doubled CO₂ climate. Yet the relationship between CG flash count and cold-cloud coverage shown in Fig. 3.16 suggests that this type of parameterization, when applied over the tropical ocean, *may*¹ be invalid (and this is likely to be true for the mid-latitudes in certain circumstances; e.g., Goodman and MacGorman, 1986).

¹ Fig. 3.15 shows only the CG flash count as compared to the *total* flash count used in the parameterization of Price and Rind, 1994.

3.3 Observations of CG lightning as related to wind shear over the warm-pool

Up to this point we have only considered thermodynamic processes (e.g., CAPE) with regard to the intensity of deep tropical convection. However, in certain situations variations in the ambient wind vector with height (i.e., wind shear) can exert a large control over the lifecycle, intensity and precipitation efficiency of convection (e.g., Rotunno et al., 1988; Trier et al., 1996; Ferrier et al., 1996). For example, in the presence of sufficient CAPE, tropospheric wind shear (particularly at low-levels) can provide a suitable environment for the development of intense, organized, long-lived convection through the production of enhanced dynamic pressure perturbations associated with interactions that occur between the ambient environmental wind profile and storm-scale circulations such as the gust front (e.g., creation of shear induced vorticity balances and associated pressure perturbations on the storm scale affect the vertical orientation and intensity of up and downdrafts; Thorpe et al., 1982; Rotunno et al., 1988; Trier and Parsons, 1995; Trier et al., 1996). On the other hand, if the CAPE is too small (leading to weaker cumulus updrafts), and the tropospheric shear too large, a strong vertical draft structure cannot be sustained in the cloud (e.g., Rotunno et al. 1988). A classic example of the negative feedback that vertical wind shear can have on tropical convective organization is to consider the effect of deep tropospheric wind shear on tropical storm/hurricane development. In this case, strong vertical wind shear “vents” the convection, which effectively prohibits the development of a “warm-core” circulation in the cyclone, thus stifling the intensification and organization of the system (e.g., see summary in Houze, 1993). Hence, in addition to the “first order” thermodynamic effects on the electrification of tropical oceanic convection, we must also consider dynamical controls such as the vertical wind shear (hereafter referred to as “shear”).

To examine the effect of shear on the development of electrified oceanic convection, we consider the wind shear for two particular layers of the atmosphere. More specifically, we compare the magnitudes of the mean daily (24-hour mean; four times averaged per day) low-level (~1000-850 mb) and deep tropospheric (850-125 mb) shears in time-series form, to that of the CG lightning and CAPE time-series (e.g., Fig. 3.5). The daily mean shear magnitudes for the immediate vicinity of the R/V Vickers were computed from the COARE gridded sounding data set (Lin and Johnson, 1996a) and the R/V Vickers sounding data (111 soundings). Daily mean low-level wind shear values computed in the gridded data set were heavily weighted by the Vickers wind data, and therefore the two computations (gridded and Vickers data) yielded very similar values of low-level/deep tropospheric shear. It should be noted that for the shear calculation, soundings taken in or near convection from both the gridded and R/V Vickers sounding data sets were not removed (these were only a small proportion of the sample in the Vickers data, e.g. <5 %). Therefore, the presence of convectively induced motions in the wind-field are implicit in the shear calculations. We consider the magnitude of the shear and CG lightning activity relative to the value of CAPE, and therefore focus on the 30-day time-period shown in Fig. 3.5, for which most reliable values of CAPE were available.

Smoothed time-series of shear (low-level and deep tropospheric), CAPE and CG lightning are shown in Figs. 3.19 a-b. Both the low-level and deep tropospheric shear can be classified as being unidirectional, as zonal wind components dominated the wind profile, hence increases in the vector shear are due almost exclusively to increases in the zonal wind speed (e.g., Rickenbach, 1996). The low-level shear vector was oriented from west to east, and the deep tropospheric shear vector was oriented east to west (a "reverse" shear). In Fig. 3.19a, the low-level shear varies in phase with both the CG lightning and the CAPE, however the amplitude variation (peak to trough differences by a factor of 2-3) of the low-level shear is much (slightly)

smaller than that of the lightning (CAPE). Additionally, variations in the CAPE are better correlated to the variations in lightning activity (e.g., shear-CG, $r^2=.37$; CAPE-CG, $r^2=.62$). In comparison, the deep tropospheric shear (Fig. 3.19b), shows even less variation than the low-level shear as function of time, however the relative “peaks” in the deep shear do appear to occur approximately coincident with relative peaks in the lightning and CAPE.

Based on the three-dimensional modeling studies of tropical convection discussed in Trier and Parsons (1995), in moderate to light wind shear regimes (characteristic of the wind profile over the tropical ocean) we would expect to see only a slight increase in the area coverage, upscale growth, and rainfall production associated with deep convection, when a doubling of the wind shear occurs at low levels (e.g., an increase in velocity differential of 5 m s^{-1} , to 10 m s^{-1} over a depth of 1-1.5 km). Interestingly, the observations of Rickenbach (1996) suggest that periods associated with enhanced CG lightning in the IFA (Fig. 3.21), were also associated with a relatively greater frequency of *larger organized* convective systems (linear and non-linear MCS's; via Rickenbach's classification), and this is consistent with the variations of shear and CAPE shown in Fig. 3.19 and the modeling studies of Trier and Parsons. Further, the diurnal cycle of low-level wind shear (not shown) computed from R/V Vickers sounding data suggests that a relative increase in the wind shear occurred during the night (factor of 1.5, which is also consistent with the observed nocturnal peaks in lightning, rainfall and deep convection (Fig. 3.13). Interestingly, both the daily time series (Fig. 3.19) and diurnal cycle of low-level wind shear, suggest that small (e.g., a factor of 2) increases in the shear were associated with disproportionate increases in the CG lightning (factor of 5-10). This behavior is similar to the previously noted variations between CG flash count, θ_w , MLMR and CAPE (e.g., Figs. 3.6-7)..

3.4 Time-Longitude variations of CG lightning over the western Pacific Ocean

The life-cycle characteristics of deep convection in tropical disturbances are often studied using time-longitude diagrams of satellite measured OLR or IR brightness temperature (e.g., “Hovmöller diagrams”). This type of satellite analysis can geo-locate deep convective cloudiness over synoptic spatial temporal scales (e.g., Nakazawa, 1986; Knutzen and Weikman, 1987; Lau et al., 1989, 1991; Chen et al. 1995; Numaguti et al., 1995). Once located, OLR and brightness temperature anomalies can be used to infer the movement and spectral characteristics of given disturbances. Importantly, cold-cloud brightness temperatures do not always yield accurate location, intensity, or the heaviest rainfall associated with the deepest convection in a tropical disturbance. That is, OLR and/or brightness temperature anomalies do not provide an unambiguous “look” at the internal characteristics of the clouds being sampled. For example, low brightness temperatures associated with large, non-precipitating cirrus canopies generally comprise the largest areal fraction of tropical disturbances (e.g., Riehl and Malkus, 1958), but can be incorrectly classified as “deep convection” or “heavy rainfall” in algorithms that use IR brightness temperature or OLR as a proxy for the presence of deep convection and rainfall (e.g., Arkin and Meisner, 1987; Mapes and Houze, 1993; Nitta and Sekine, 1994; Sheu et al. 1995). This is an important point because recent studies suggest that approximately 40% of all tropical rainfall is produced by the deepest 3%-7% of observed convection (e.g., Janowiak et al., 1994; Liu et al., 1995; DeMott, 1996). This relatively small percentage of deep convection is also likely to be associated with the release of significant latent heat (e.g., the small percentage of undilute “hot-towers” described by Riehl and Malkus, 1958). Therefore, *any* observable parameter that can identify the deepest, most intense convection in a tropical disturbance is likely to be useful for identifying regions of heavy convective rainfall and significant latent heat release.

Given that tropical disturbances force changes in atmospheric stability (e.g., CAPE) and dynamical variables such as wind shear, it is reasonable to suggest that variations in lightning activity will also occur. Based on previous observations and the scaling arguments presented in Sec. 1.1, it is also likely that the lightning activity will be associated with the most intense (vertically developed) convection observed in the disturbance. To investigate the variation of lightning in both time and space relative to the movement of disturbances over the western Pacific Ocean, we generated a "CG lightning-Hovmoller diagram" in time-longitude space (similar to Fig. 3.1) for a large region of the TOGA COARE domain. This domain encompasses both island and isolated oceanic regions and is contained in a rectangular area centered on the equator between 5° N - 5° S latitude and 146° E - 164° E longitude. The lightning data were counted daily and binned into $0.5^{\circ} \times 0.5^{\circ}$ grid boxes for the time period 1 January 1993 - 28 February 1993. The resultant diagram of lightning variation with time and longitude was then compared to a subsection of OLR data taken from Fig. 2.3 for a collocated latitude-longitude box and coincident time-frame, and the variation of both variables (CG lightning and OLR) is shown in Fig. 3.20. Prior to making explicit comparisons between the CG lightning and OLR in Fig. 3.20, we must exercise a note of caution. The OLR data (as supplied by Mr. X. Lin, CSU) were substantially filtered prior to being used in Fig. 3.20 (the data were filtered using a 5-day running mean and averaged over $1^{\circ} \times 1^{\circ}$ grid boxes). The lightning data were also low-pass filtered but in a different fashion (e.g., a 1-step Leise filter was used). The smaller temporal and spatial coverage of the lightning data dictated the use of a different filter (i.e., a filter that retained higher wavelengths). Also, it is likely that the COARE ALDF network experienced a significant reduction in detection efficiency at longitudes located east of $\sim 160^{\circ}$ E. Therefore, even though we extend the region of lightning coverage to 164° E for the purpose of the analysis, flash counts will be biased toward significantly lower values east of $\sim 160^{\circ}$ E.

Having acknowledged the potential problems in analysis of the data shown in Fig. 3.20, we are now prepared to discuss several interesting observations revealed in Fig. 3.20. First, visual inspection the lightning data (but not necessarily the OLR data) suggest the occurrence of intense convection (and heavy rainfall) at 4-6 day intervals in the western portion of the domain over the islands located in the meridional belt of 150°E to 152°E (bins 9-13), and a frequency of ~6-10 days for the occurrence of lightning over the entire domain. From a rainfall standpoint, the temporal “banded” structure of the lightning data situated between bins 17 and 21 (155°E - 157°E), although representing a much larger area latitudinally, is consistent with coincident peaks in the rainfall measured at the location of the MIT radar (Fig. 3.16-18; c.f., Rickenbach, 1996). Secondly, if the daily trend in the maximum eastward extent of the CG lightning counts is followed, weak propagation of the lightning signal in both eastward and westward directions is observed at varying frequencies. It is interesting to note that during the two time periods of late December 1992 - January 1993, and early February 1993, the near equatorial regions of the COARE domain were influenced by the passage (and associated deep convection) of 30-60 day intra-seasonal oscillations (ISO; e.g. Madden and Julien, 1971). Both the lightning and OLR data in Fig 3.20 reveal the trailing edge of deep convection associated with the *first* ISO, which subsequently exited the area during the first week of January (days 1-5 in Fig. 3.20) coincident with the development of strong westerly winds at the surface (e.g., Lin and Johnson, 1996 a,b).

The passage of the *second* ISO is indicated in the OLR data of Fig. 3.20 by the eastward propagation of an envelope of low OLR (e.g., values $\leq 200 \text{ W m}^{-2}$) that enters the domain of Fig. 3.20 on approximately day 21 (21 January 1993). The OLR and lightning data indicate the presence of a collocated enhancement in convection situated in the western half of the domain on day 26, approximately. The CG lightning and OLR data then indicate a discrete, step-like propagation of the convection occurred as it moves toward the eastern side of the domain

between days 27-34. At day 34 (3 February 1993) the lightning data suggest a rather abrupt halt to the convection in the region, and this electrically inactive period extends to approximately day 38 (7 February 1993). Coincident with the abrupt drop in lightning activity, the OLR data also show a relative decrease in cloudiness, and radar reflectivity statistics (DeMott, 1996) suggest the presence of only weak, very shallow convection (30 dBZ heights of 3 km). The decrease in lightning and convection that occurred during this time-period was associated with the advection of dry air (a dry-intrusion) into the area at low to mid-levels of the troposphere (e.g., Yoniamama and Fujitani, 1995; Mapes and Zuidema, 1996). This dry-intrusion produced a marked reduction in CAPE (e.g., Fig. 3.5) and effectively suppressed the development of deep convection. After day 39, the dry intrusion advected out of the domain of Fig. 3.20 and was replaced by deep convection associated with a disturbance that appeared to propagate into the region from the east. This is shown clearly in the lightning and OLR data in Fig. 3.20.

In Fig. 3.20 it appears that the lightning and OLR data both indicate a similar location and movement for the deep convection. However, the filtered OLR data do not give as precise a location for the most intense convective activity as the lightning data do. In this sense, the lightning data are helpful in distinguishing where intense, electrically active convection is located under the broader coverage area of the cloud-shield. It is important to point out that the lack of lightning activity in regions under the cold-cloud shield does not imply the absence of deep convection. However, in the event that neighboring *regions* (as opposed to individual cells) of deep convection with similar cloud-top brightness temperatures, but dissimilar degrees of electrical activity, are observed coincident in time (e.g., Fig. 4.1, discussed in Chapter 4), it is likely that the clouds in those regions are associated with slightly different microphysics and/or kinematics. In this particular situation even the “null” case, with regard to electrical activity,

might be useful (e.g., if a significant fraction of one cloud-ensemble were associated with lightning) when used in a contrast mode for assessing the gross internal structure of the clouds.

To confirm the more “qualitative” visual indication of apparent 4-10 day cycles in the lightning data shown in Fig. 3.20, a spectral analysis was carried out for flash locations situated in the meridional band of 150E-160E. The spectral analysis (Fig. 3.21) suggests the presence of weak cycles in the 2-5 day (e.g., Lau et al., 1991; Numaguti et al, 1995) and 8 day period ranges (with the 2-day cycle being most prominent). The peak in the variance at 20 days is probably real but of questionable statistical significance from an analysis standpoint given that only 60 days of CG lightning data were available for analysis. Note that the 4-10 day cycles that were identified visually in Fig. 3.20 do appear (albeit weakly) in the spectral analysis.

3.5 Concluding Remarks

The observations presented herein suggest that variations in “large-scale” thermodynamic and dynamical quantities (e.g., CAPE, profiles of thermal buoyancy, wind shear, wave disturbances etc.) are reflected in variations of CG lightning activity over the western Pacific warm-pool. In particular, relatively *small variations* in thermodynamic quantities such as CAPE and mean mixed-layer θ_w , which are forced to a large extent by small variations in boundary layer specific humidity brought about by LH fluxes or relative increases in boundary layer moisture convergence (e.g., Figs. 3.2, 3.3b, 3.5, 3.9, 3.10), seem to produce *large variations* in electrical activity as reflected in CG lightning flash counts (e.g. Figs. 3.4-7). These sensitivities are manifested in both time-series form (e.g., Fig. 3.5) and over the diurnal cycle, where CG lightning exhibits a pronounced nocturnal maximum over the tropical ocean coincident with peaks in rainfall and CAPE (Fig. 3.13).

Changes in the thermodynamic and dynamical environment of the tropical atmosphere over the warm-pool forced by phenomena such as propagating wave-disturbances, cause an associated “signature” of the wave-disturbance in the lightning data. In turn, when plotted in time-longitude space, the CG lightning data are capable of unambiguously identifying the presence of “intense convection” associated with the disturbances (e.g. Figs. 3.20). However, questions remain as to what defines the term “intense convection” (a cloud-scale phenomenon), when referring to cloud and lightning observations analyzed over large domains such as that shown in Fig. 3.20. More specifically, the following questions need to be addressed in order to link the “large-scale” lightning and cloud observations to convective/cloud-scale processes:

- 1) What does lightning-producing convection over the tropical ocean “look” like on the cloud-scale (e.g., microphysically, dynamically), and how is the convective lifecycle related to the production of lightning?
- 2) How are the apparent sensitivities to small changes in thermodynamic variables such as CAPE and/or the profile of thermal buoyancy manifested in the cloud-scale electrification process?
- 3) Does lightning producing tropical convection possess a distinctly different diabatic heating profile relative to deep non-lightning producing tropical convection?
- 4) How are rainfall and lightning (processes which occur on the cloud-scale), related on the large-scale?
- 5) When lightning is present, are the average properties of a cloud-ensemble different when compared to the same properties in lightning-free cumulus population?

CG/DWN CAPE (11-15 S; 129.5-134E): 12/27/89-1/30/90

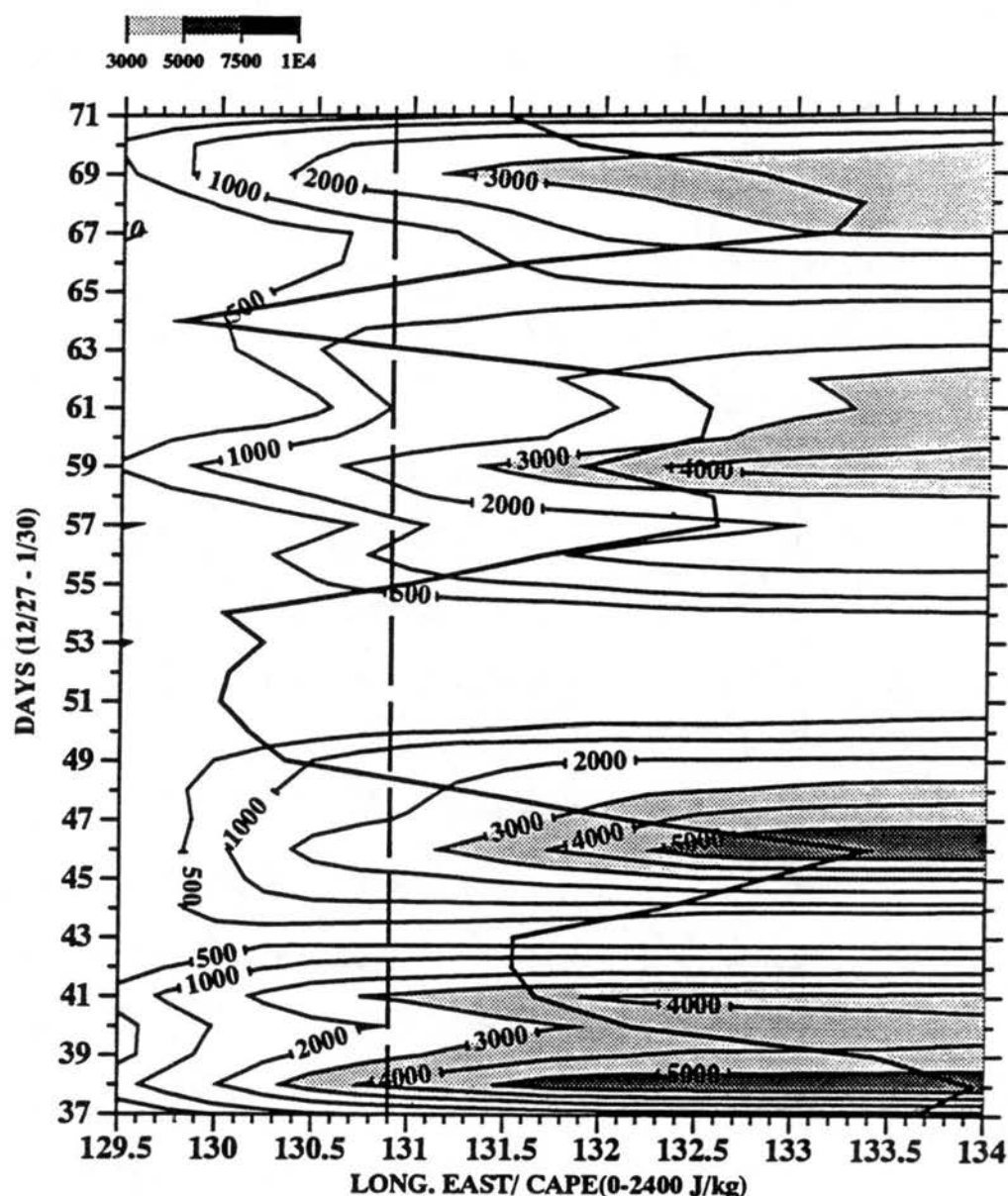


Figure 3.1 Time-longitude diagram of CG lightning from 11°S to 15°S for 27 December 1989 to 30 January 1990. Approximate longitude of Darwin is indicated by the dashed line. CG flash counts are contoured for values of 500, 1000 and every 1000 thereafter. Flash counts greater than 3000 are lightly shaded, and greater than 5000 are dark shaded. Relative values of smoothed CAPE are indicated by the solid line, increasing toward the right (e.g., maximum values $\sim 2400 \text{ J kg}^{-1}$) and decreasing toward the left (minimums $\sim 200 \text{ J kg}^{-1}$).

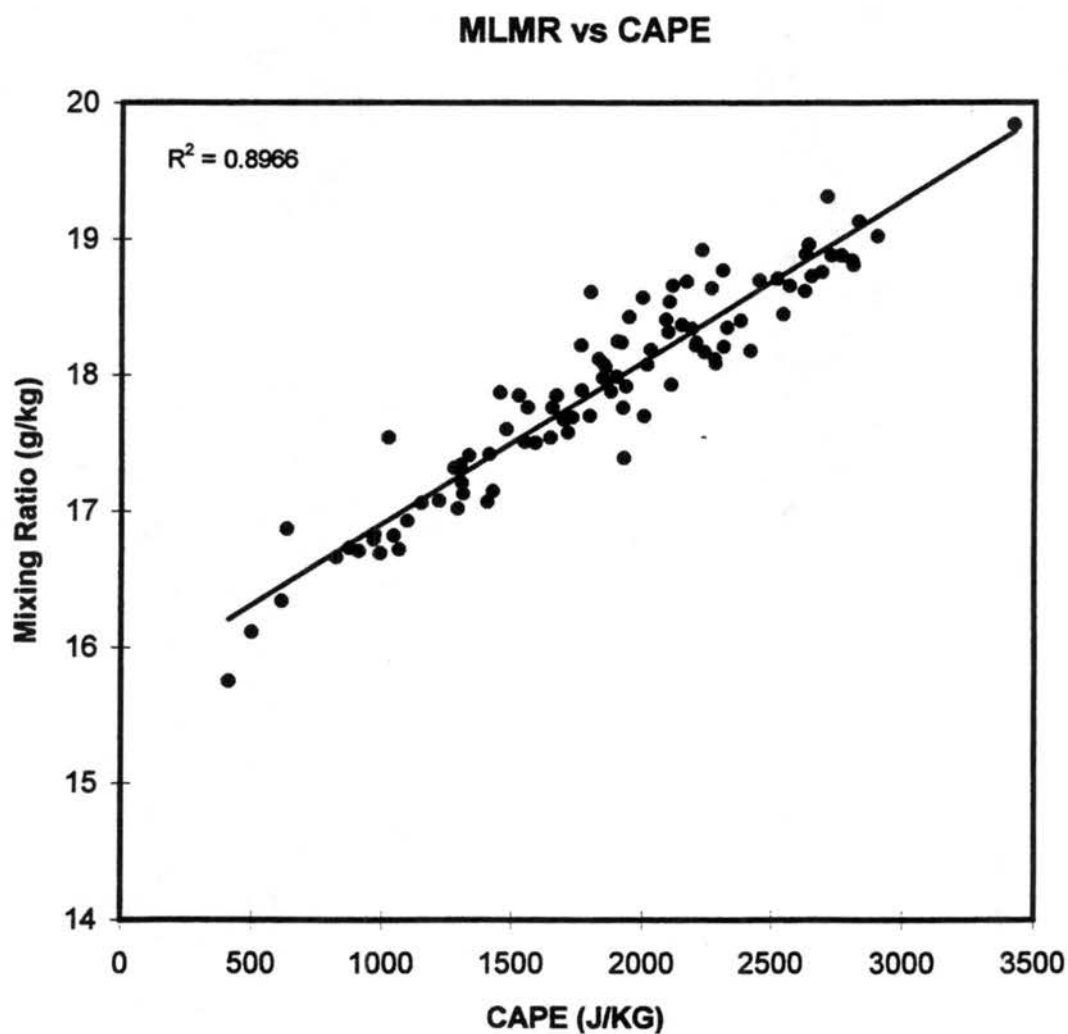


Figure 3.2 Plot of mean boundary layer mixing ratio (MLMR g kg^{-1}) vs. CAPE (J kg^{-1}) computed from six hourly soundings taken aboard the R/V Vickers during February 1993. The least squares regression line and R^2 (.8966) are also indicated.

a) **MXLYR TEMP vs WET-BULB POTENTIAL TEMPERATURE**

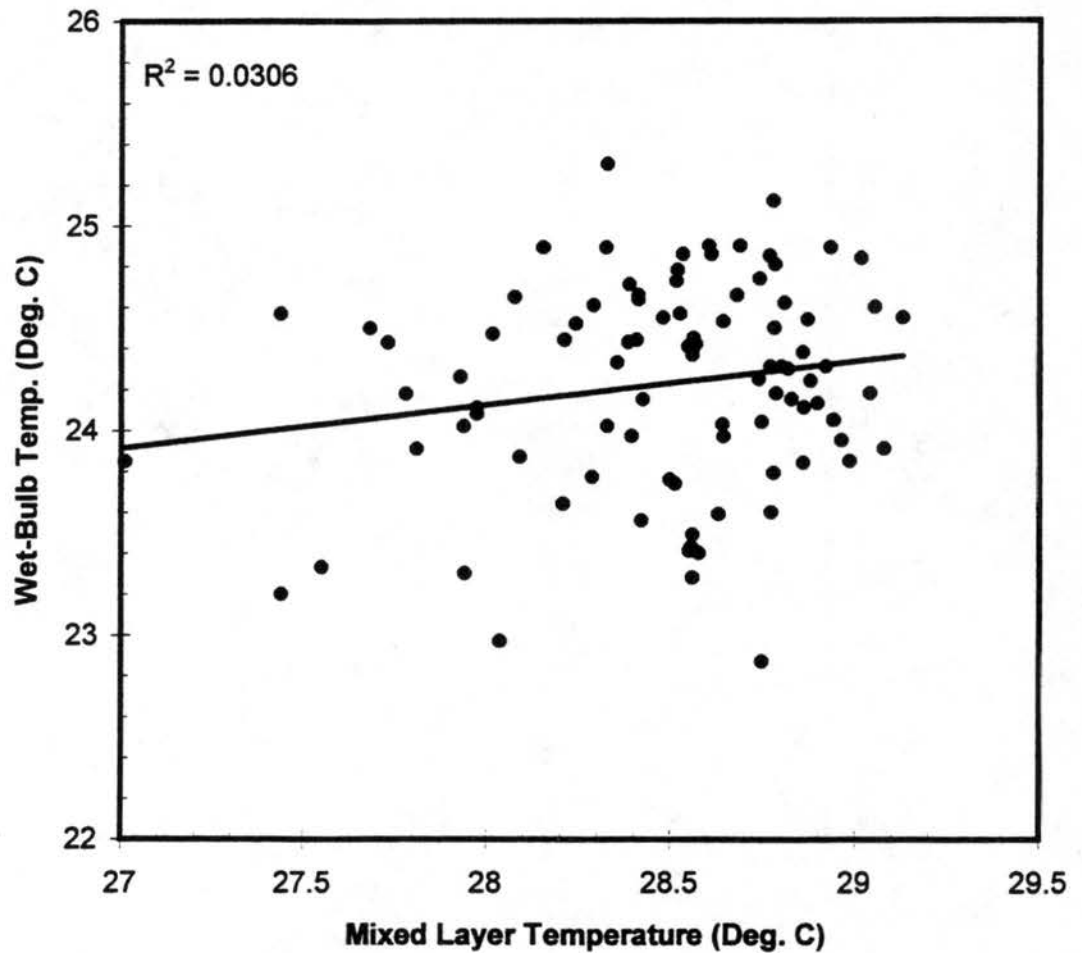


Figure 3.3 a) As in Figure 3.2 but mean mixed layer temperature vs. mean mixed layer wet-bulb potential temperature.

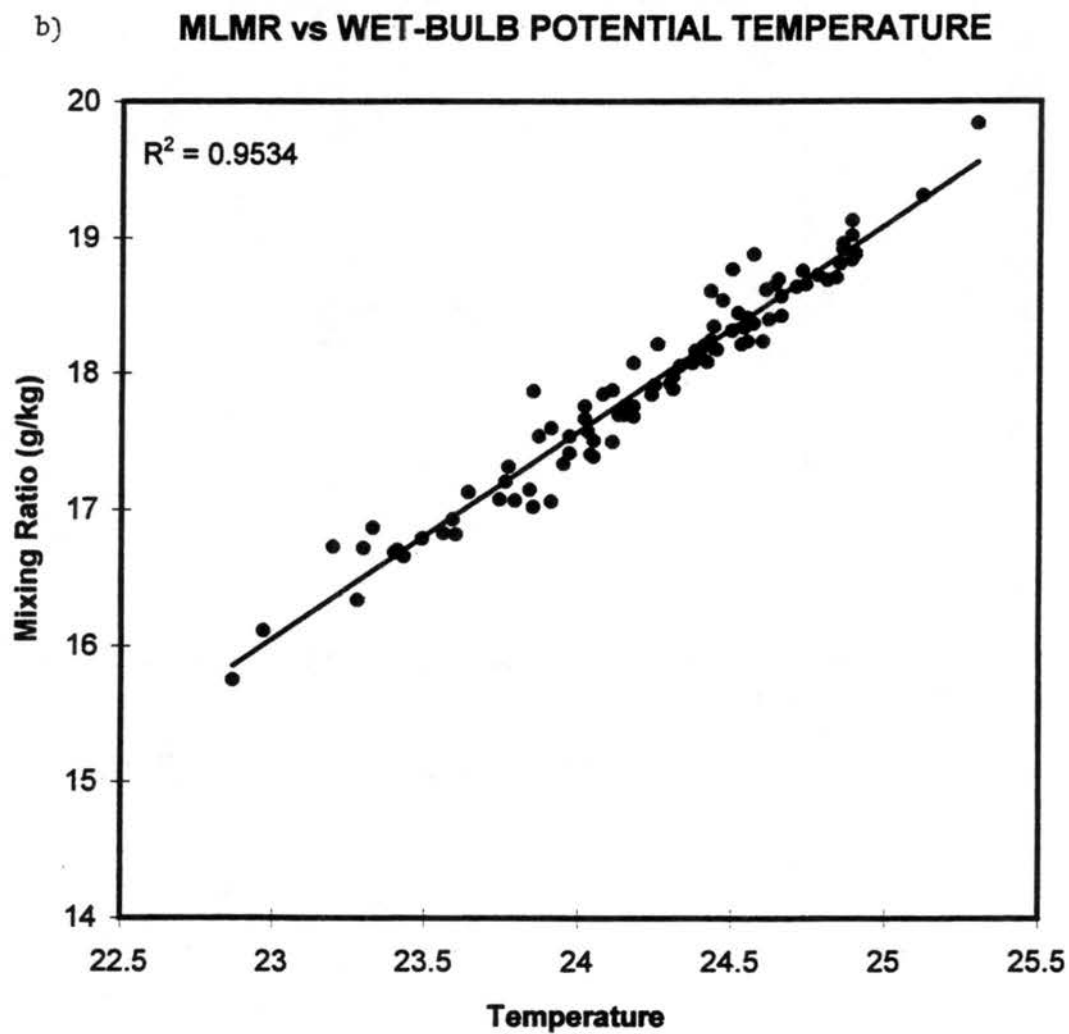


Figure 3.3 b) As in (a) but MLMR vs. mixed-layer wet-bulb potential temperature.

R/V Vickers CAPE vs. Wet-bulb Potential Temperature

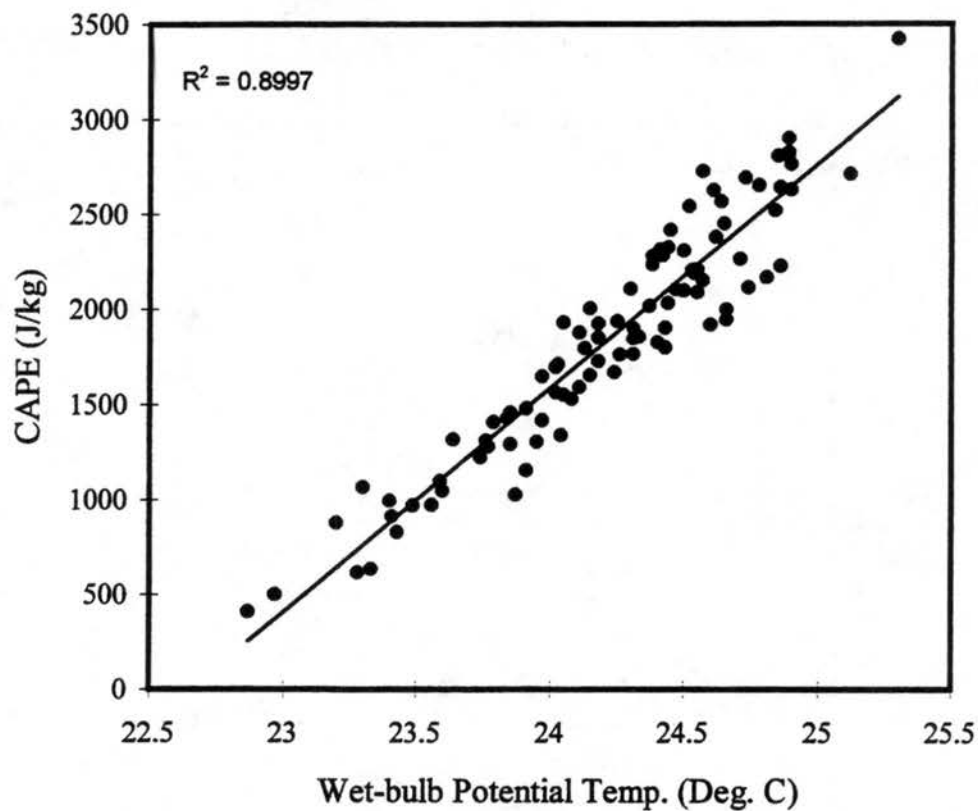


Figure 3.4 As in Figure 3.2 but CAPE vs. mixed-layer wet-bulb potential temperature.

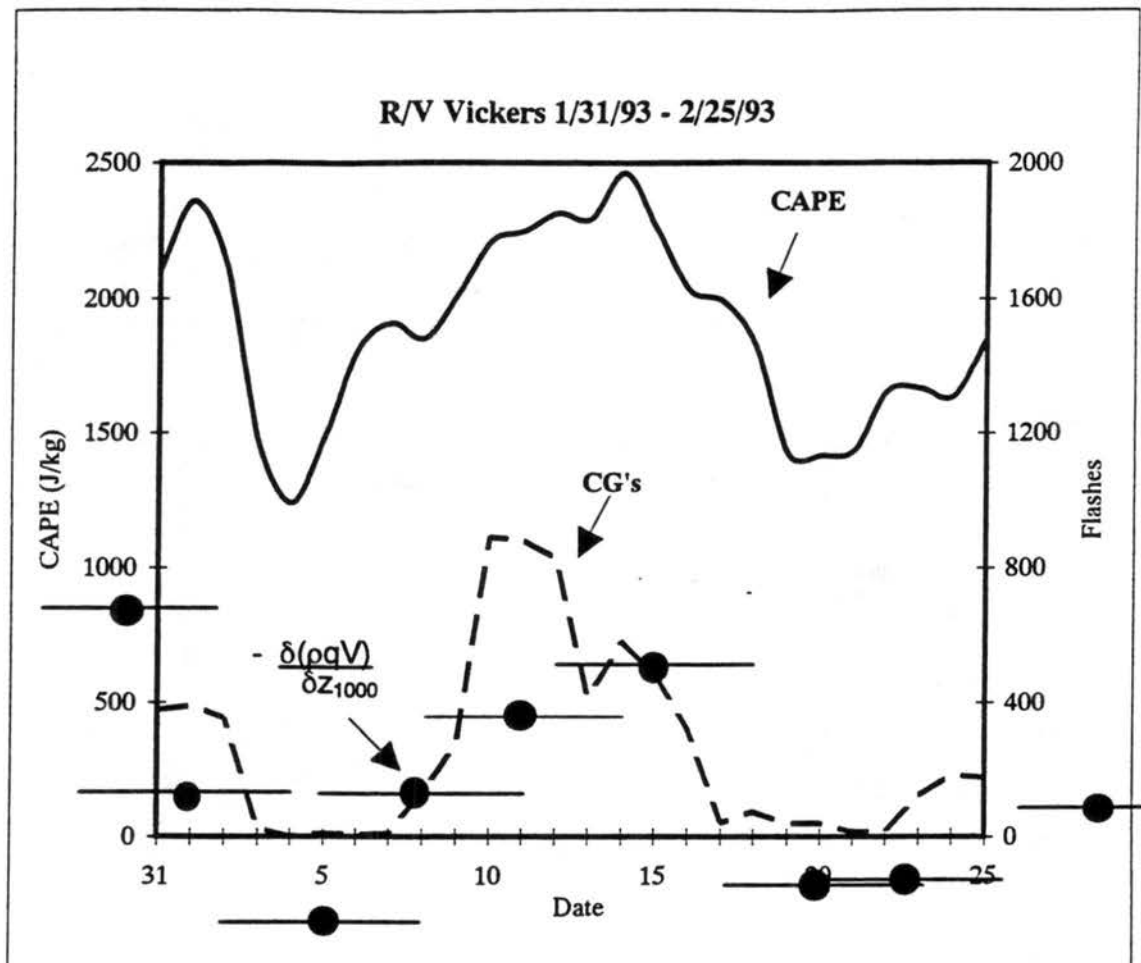


Figure 3.5 Time series of smoothed CAPE (left ordinate; R/V Vickers soundings; solid line), CG lightning flash counts (right ordinate; dashed line), and weekly *relative* moisture convergence values from the COARE gridded sounding data set at 1000 mb (large dots) for the nominal position of the R/V Vickers (1/31/93 - 2/25/93). Solid lines through the moisture convergence values represent the averaging period used to obtain the value.

Wet-bulb Potential Temperature Vs. CG Flash Number

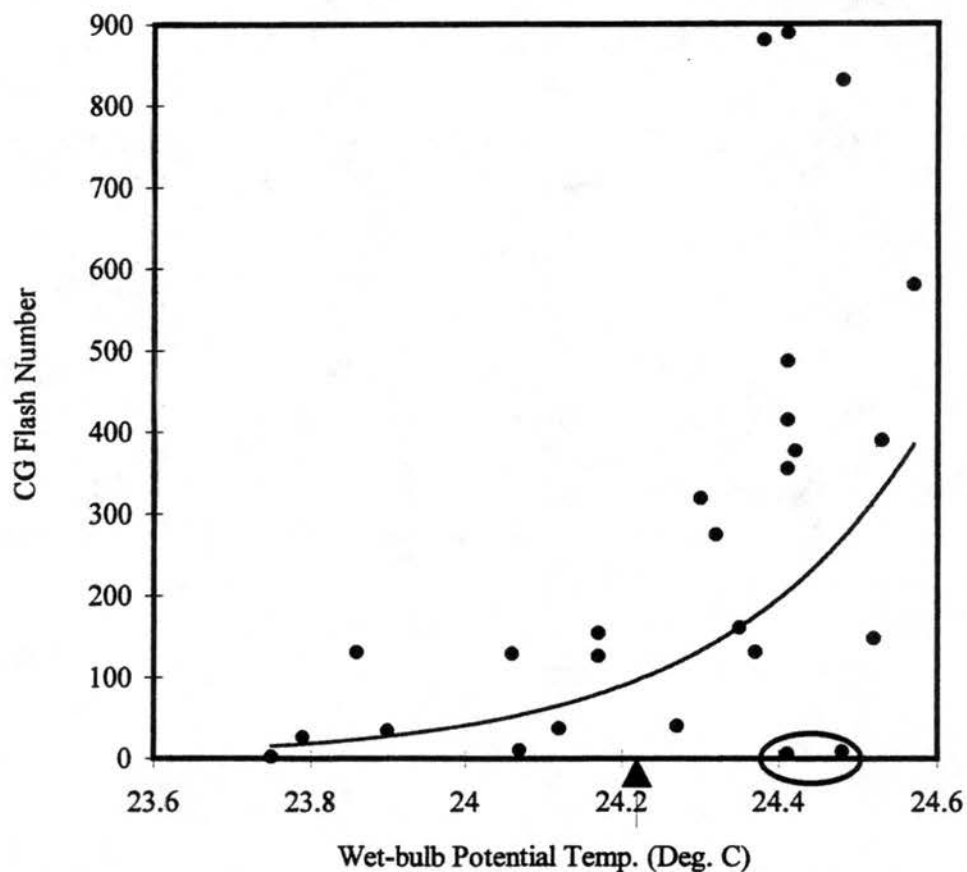


Figure 3.6 As in Fig. 3.2, but smoothed CG flash counts (ordinate) vs. smoothed 24-hour mean wet-bulb potential temperature (abscissa). The mean wet-bulb of the sample is indicated by an arrow. Two wet-bulb data points classified as outliers are circled (see text). The "best-fit" curve is indicated to illustrate the overall trend in the data.

**IMET: Time Series of 24 Hr. SST, Temp. and RH%
1/1/93 - 3/3/93**

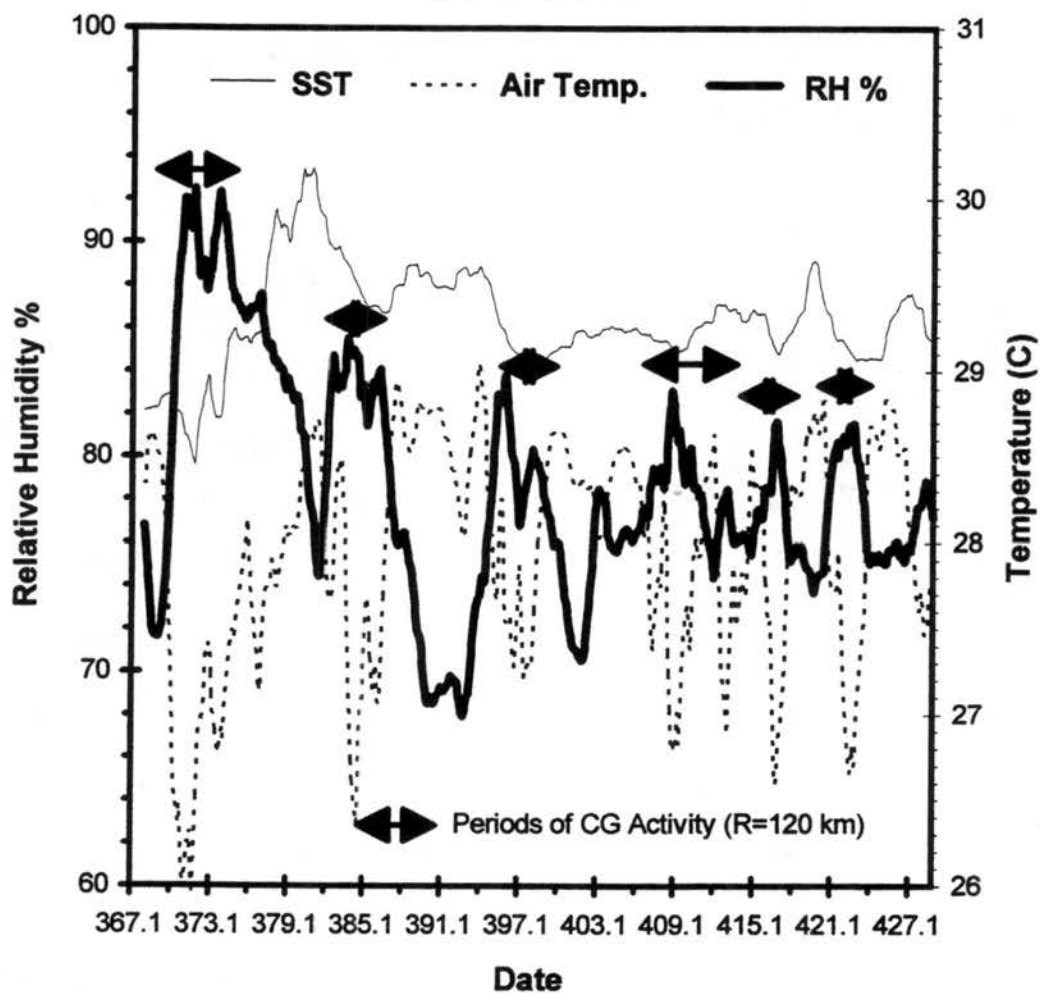


Figure 3.7 Time-series plot of IMET surface meteorological data (as indicated) and sea temperature at a depth of .45 m from 1/1/93 - 3/3/93. Also indicated are time periods associated with CG lightning that occurred within 120 km of the IMET buoy (indicated as bold arrows). Relative humidity is plotted on the left ordinate, temperature on the right ordinate. Time is indicated by COARE IOP Julian day (beginning with 1 January 1993 at day 367).

a) moisture conv. for days 93/ 1/ 1 - 93/ 1/ 7 at 1000 mb

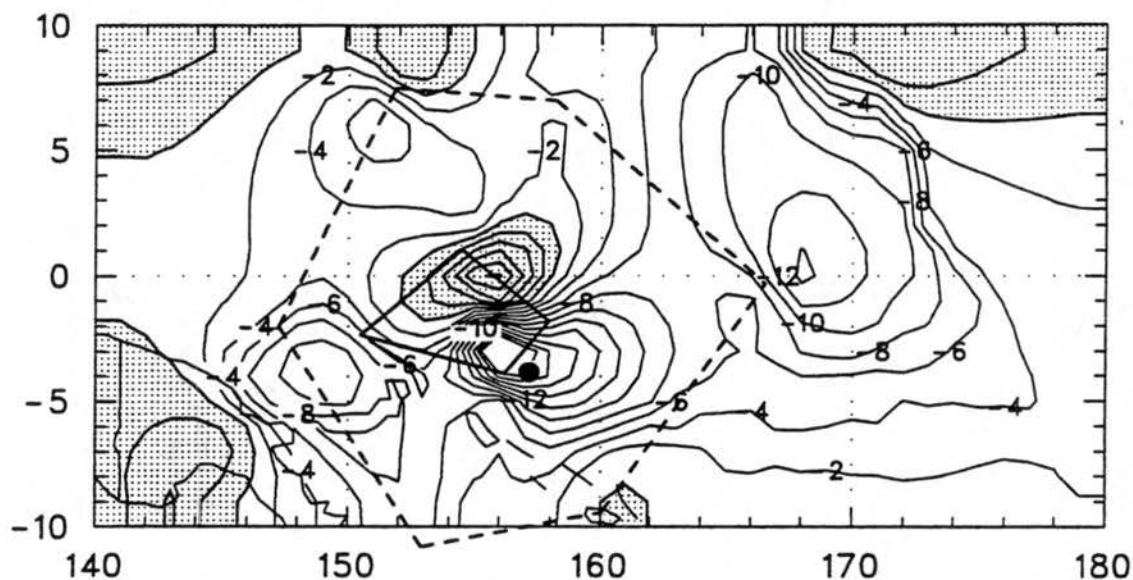


Figure 3.8 a) Mean moisture flux divergence ($\times 10^{-8} \text{ s}^{-1}$) over the western Pacific for the week of 1/1/93-1/7/93. Shaded areas represent regions of moisture flux divergence, white areas represent areas of flux convergence. Values are contoured every $2 \times 10^{-8} \text{ s}^{-1}$. The large black dot represents the mean weekly position of CG lightning over the ocean sector.

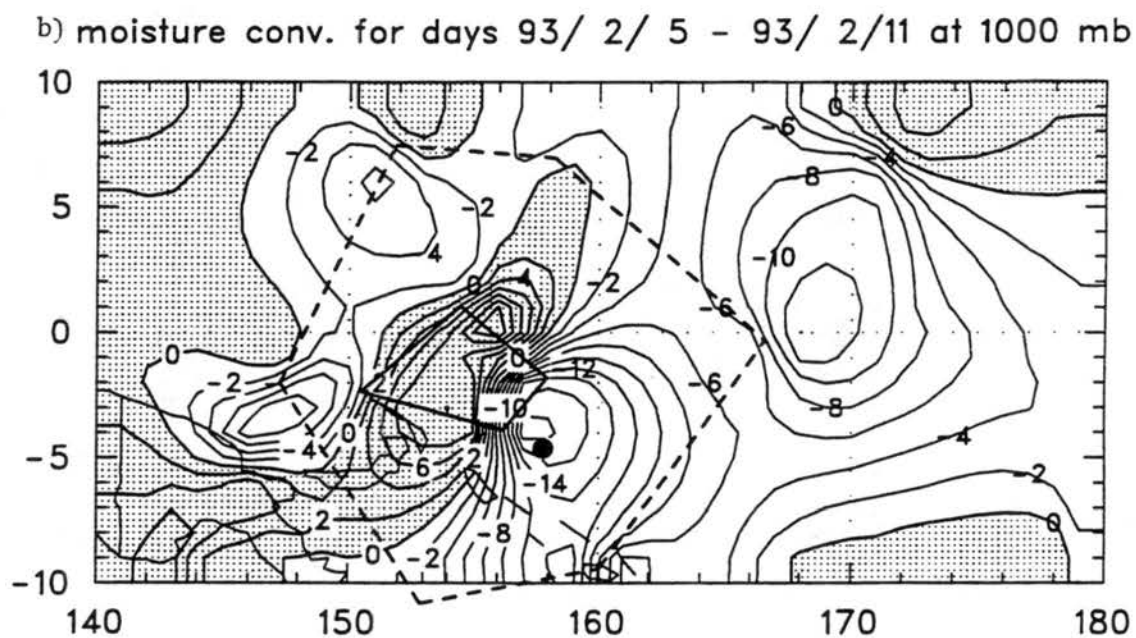


Figure 3.8 b) As in (a) but week 2/5/93 - 2/11/93

c) moisture conv. for days 93/ 2/12 - 93/ 2/18 at 1000 mb

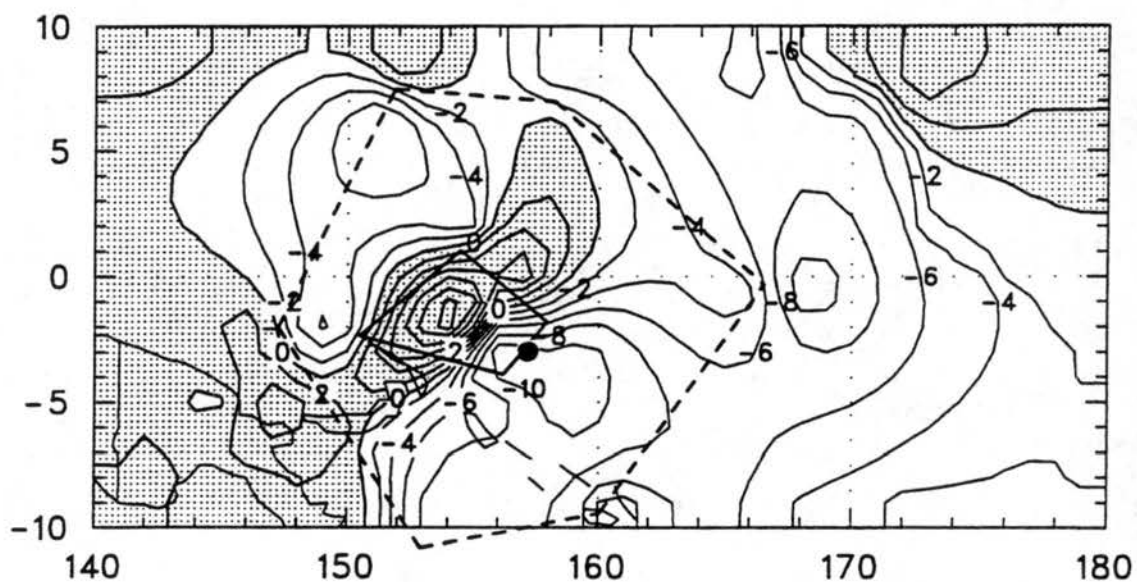


Figure 3.8 c) As in (a) but week 2/12/93 - 2/18/93.

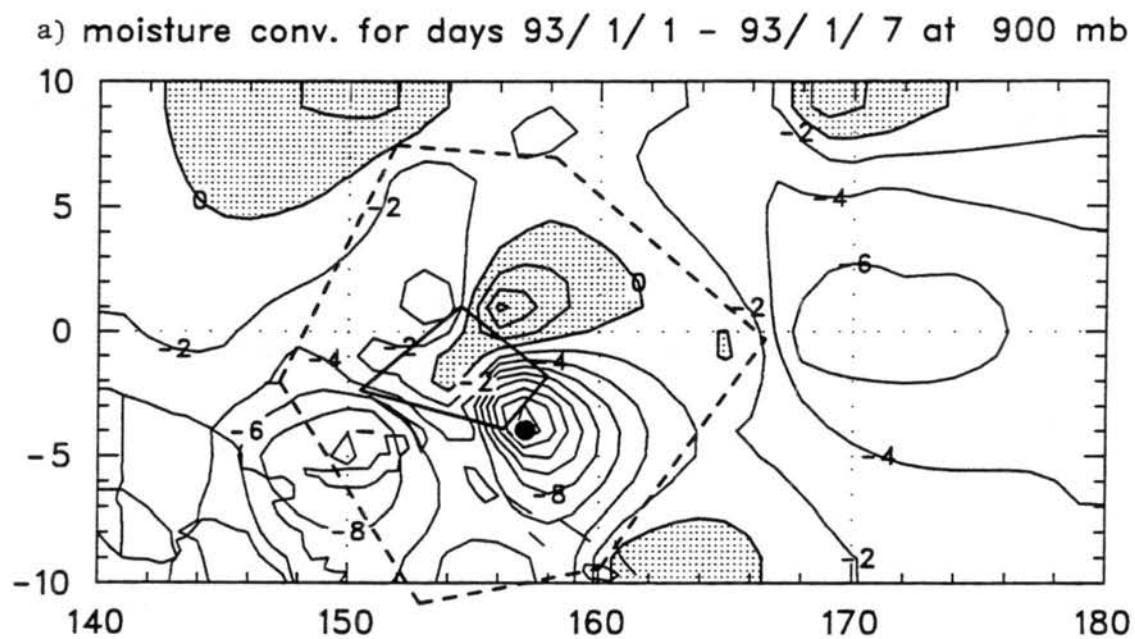


Figure 3.9 a) As in Fig. 3.8a except 900 mb.

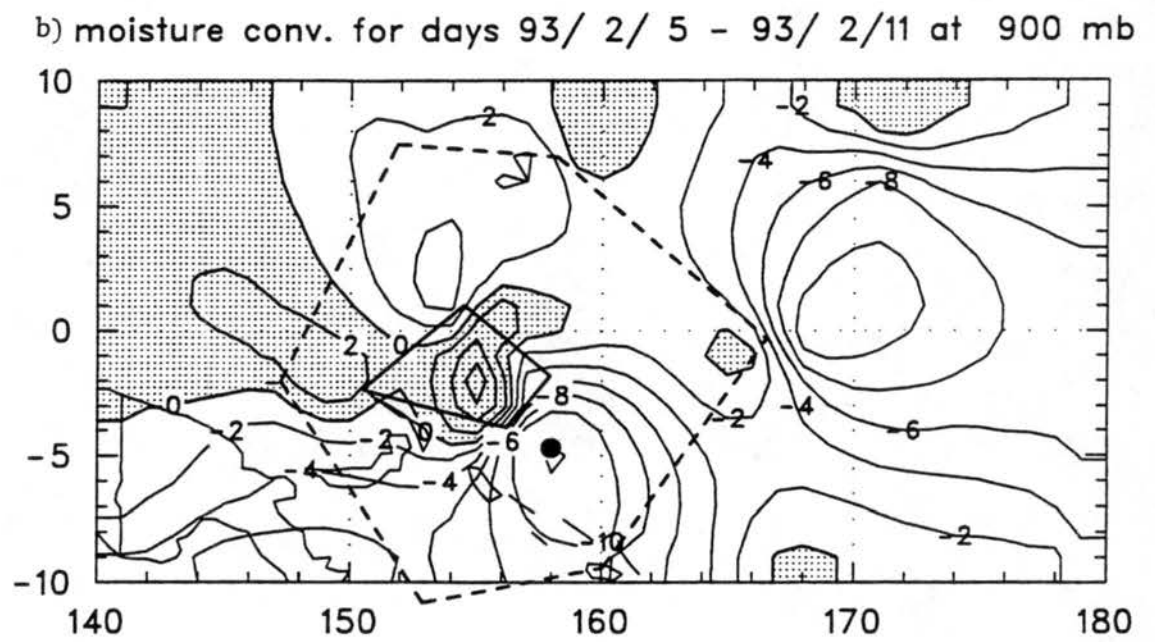


Figure 3.9 b) As in Fig. 3.8b except 900 mb.

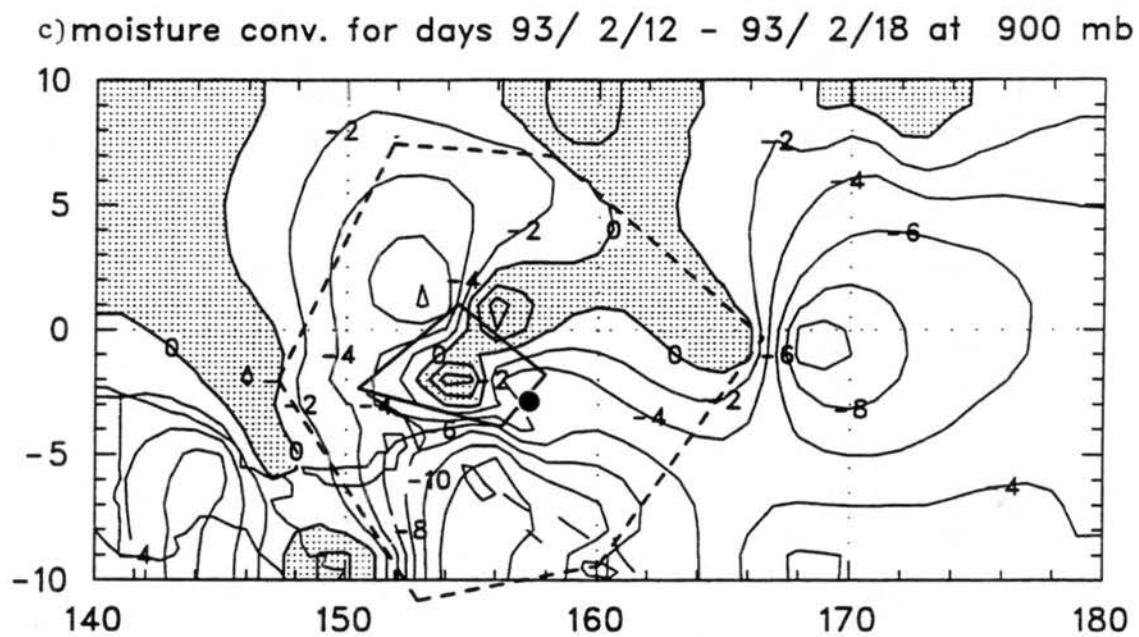


Figure 3.9 c) As in Fig. 3.8c except 900 mb.

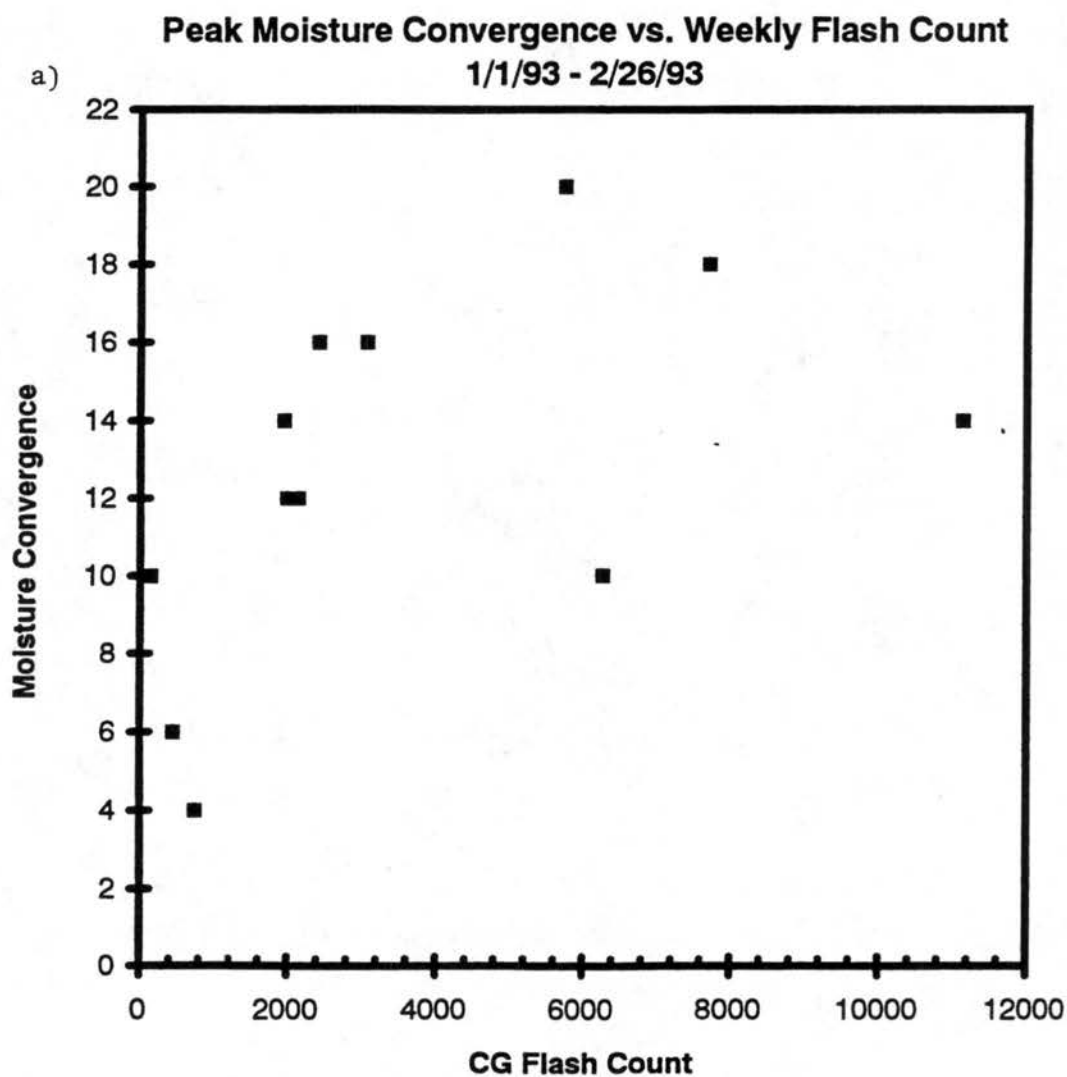


Figure 3.10 a) Peak moisture flux convergence (10^{-8} s^{-1}) associated with the weekly mean CG lightning location (e.g., Fig. 3.8-9) vs. the weekly mean CG flash count.

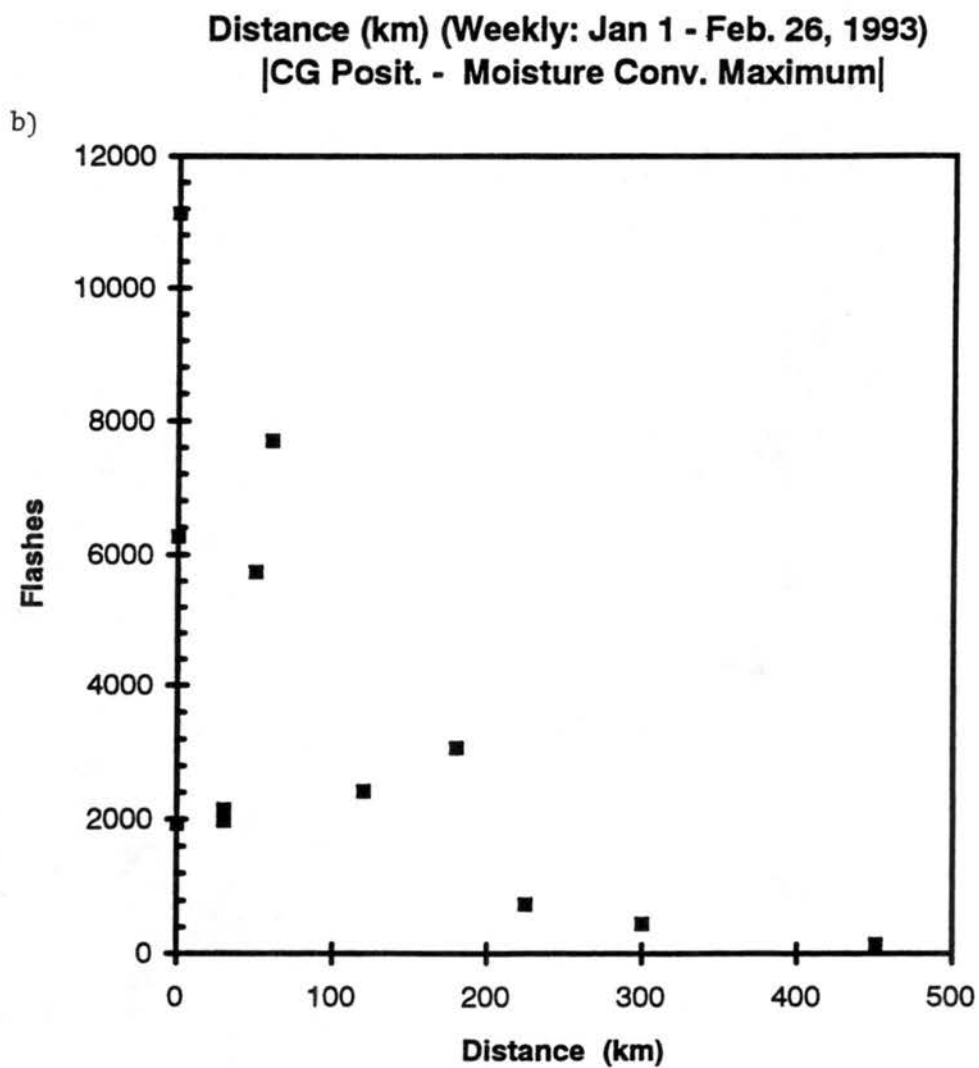


Figure 3.10 b) Distance between weekly CG flash location and the location of the moisture flux convergence maximum vs. the number of CG flashes counted each week.

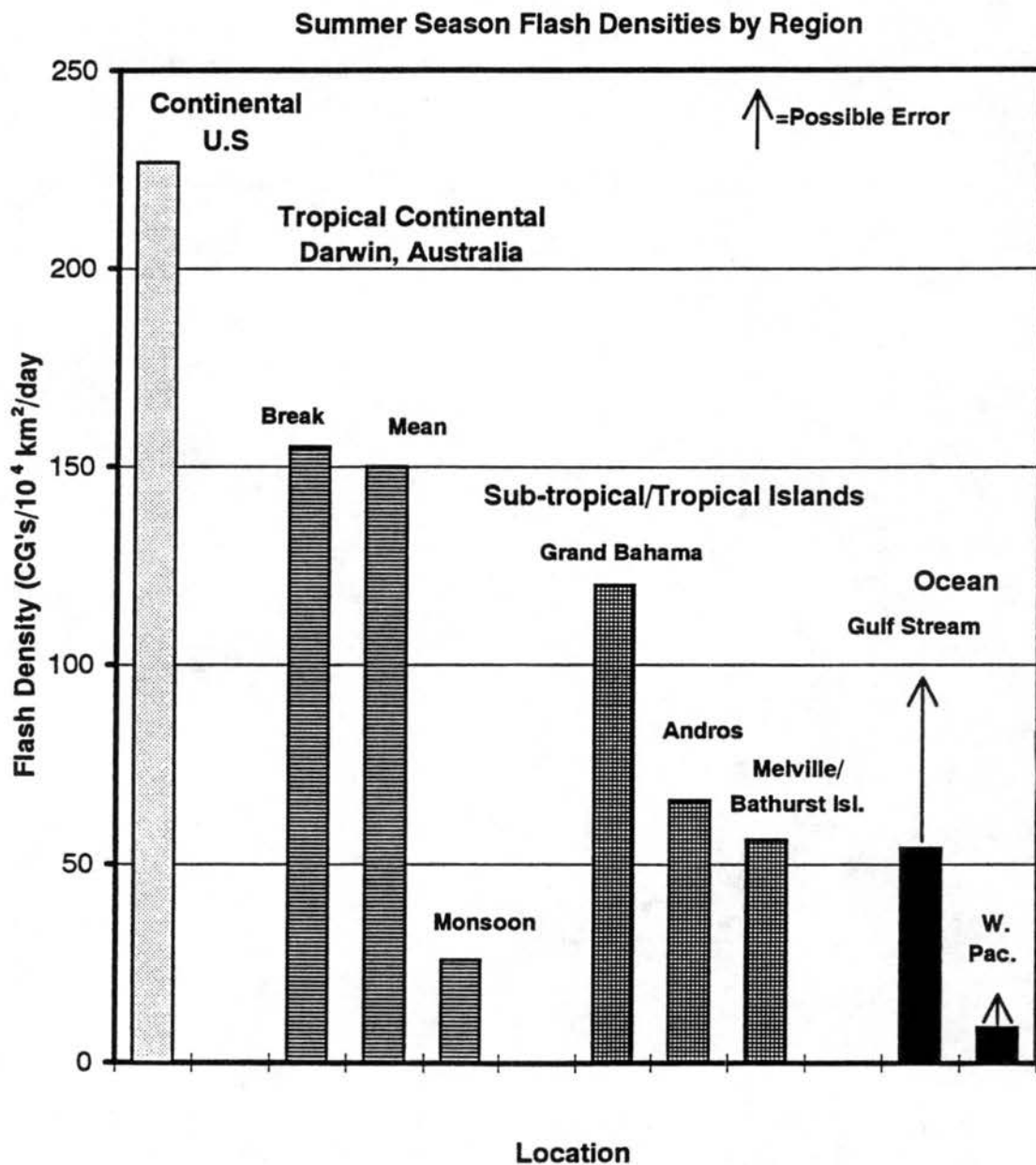


Figure 3.11 Histogram of CG flash density (flashes/10⁴ km²/day) for the continental U.S. (summer 1994-95), Darwin Australia (mean, monsoon and break periods), the Bahamas (summer 1994-95), Melville and Bathurst Islands, the Gulf Stream (to 400 km off the east coast of the U.S., summer 1994-95) and the W. Pacific warm-pool (TOGA COARE) ocean sector. The vertically oriented arrows on the Gulf Stream and warm-pool values indicates an estimated maximum flash density based on a 50% reduction in detection efficiency relative to the other networks.

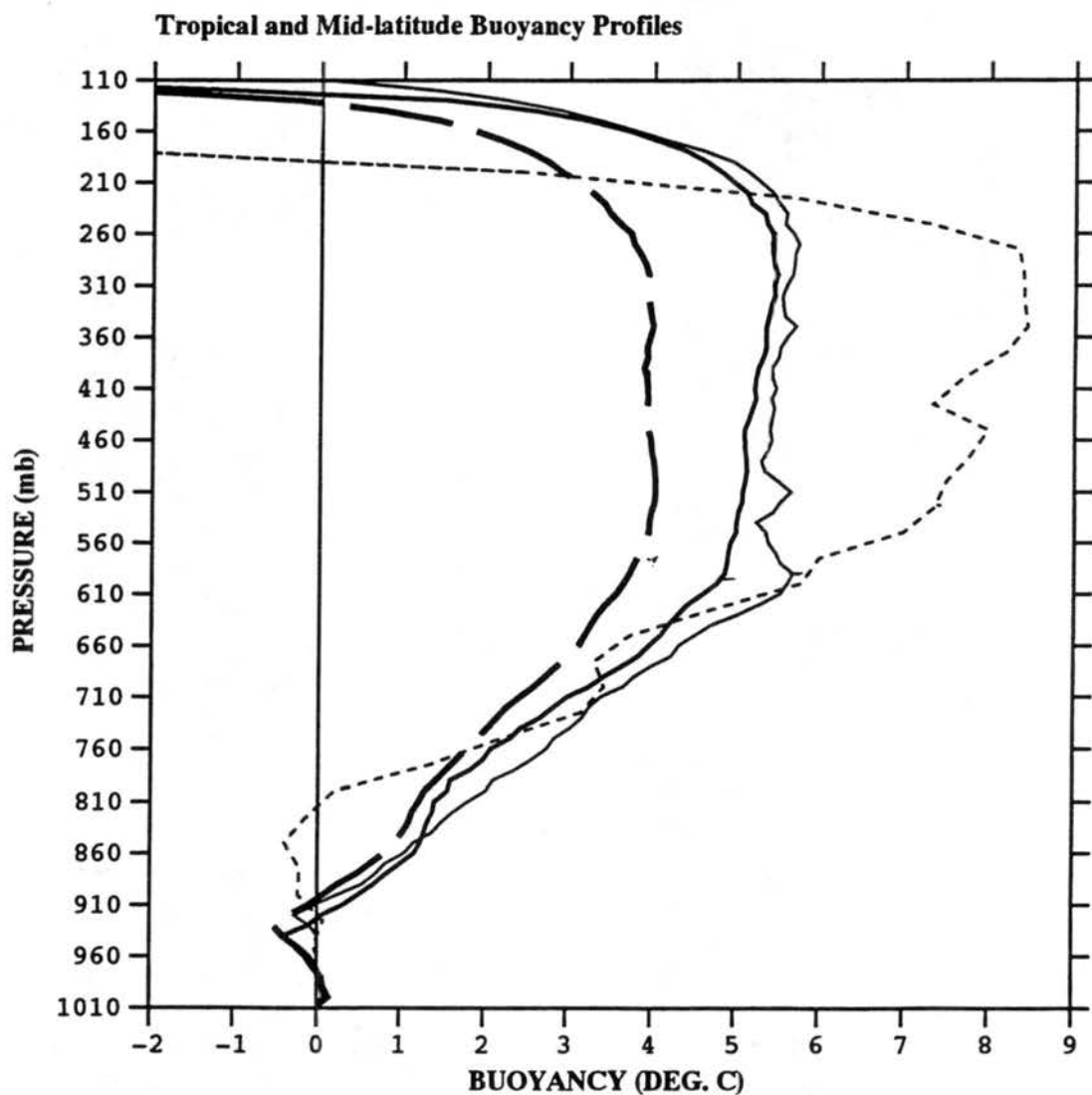


Figure 3.12 Tropical and mid-latitude thermal buoyancy profiles ($^{\circ}\text{C}$) vs. pressure (mb) for the COARE weakly electrified (long bold dash), COARE active (bold solid), DWN-Island (light solid) and mid-latitude continental (PRE-STORM; short dash) environments.

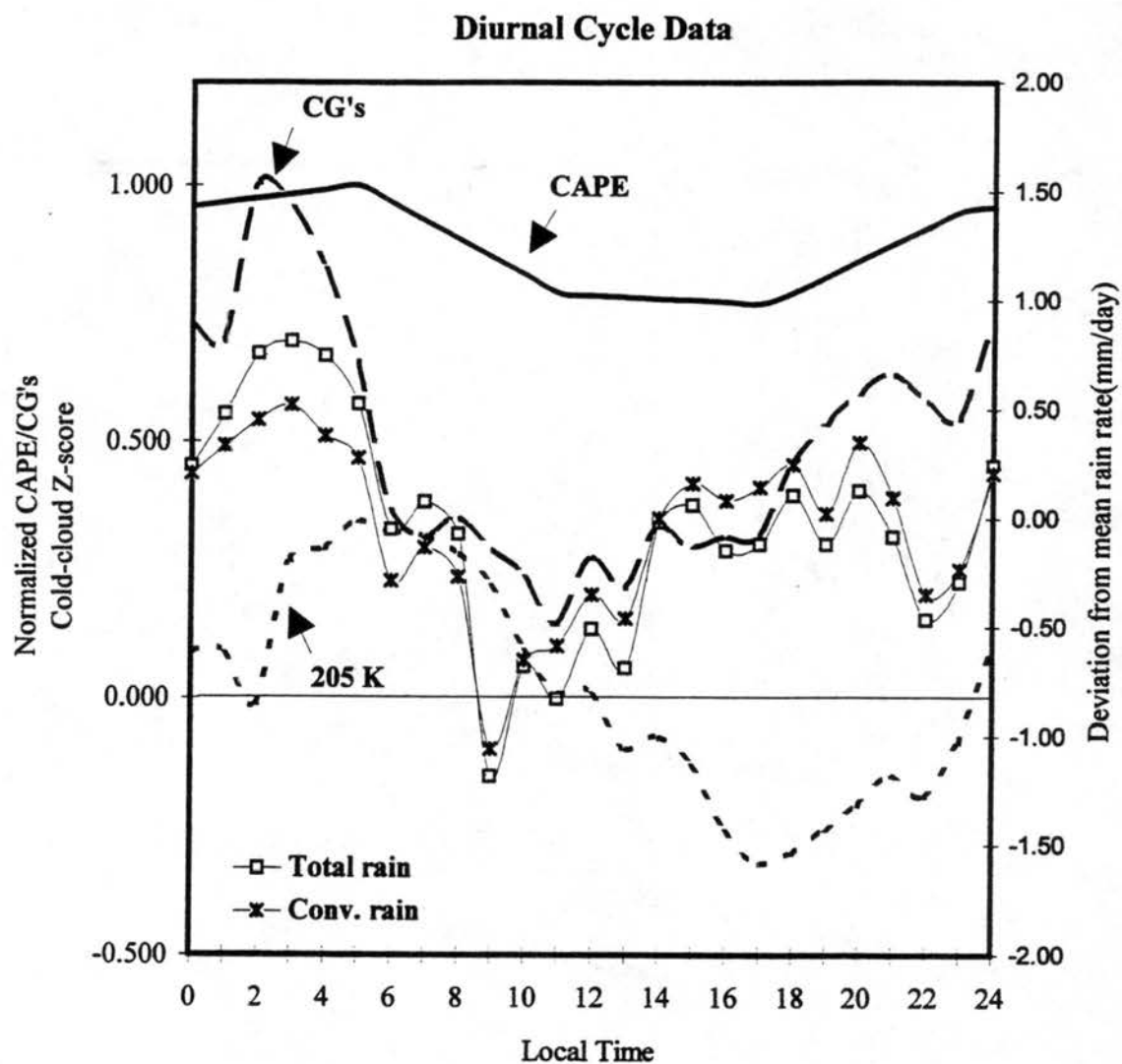


Figure 3.13 The diurnal cycle (30 days) of normalized CAPE (solid), CG lightning (dashed), and cold-cloud area with brightness temperatures < 205 K over the ocean sector (dotted line; plotted as a Z-score). Also plotted on the right ordinate are the diurnal cycles of total and convective rainfall, plotted as a deviation from the mean in mm day^{-1} (Rickenbach, 1996).

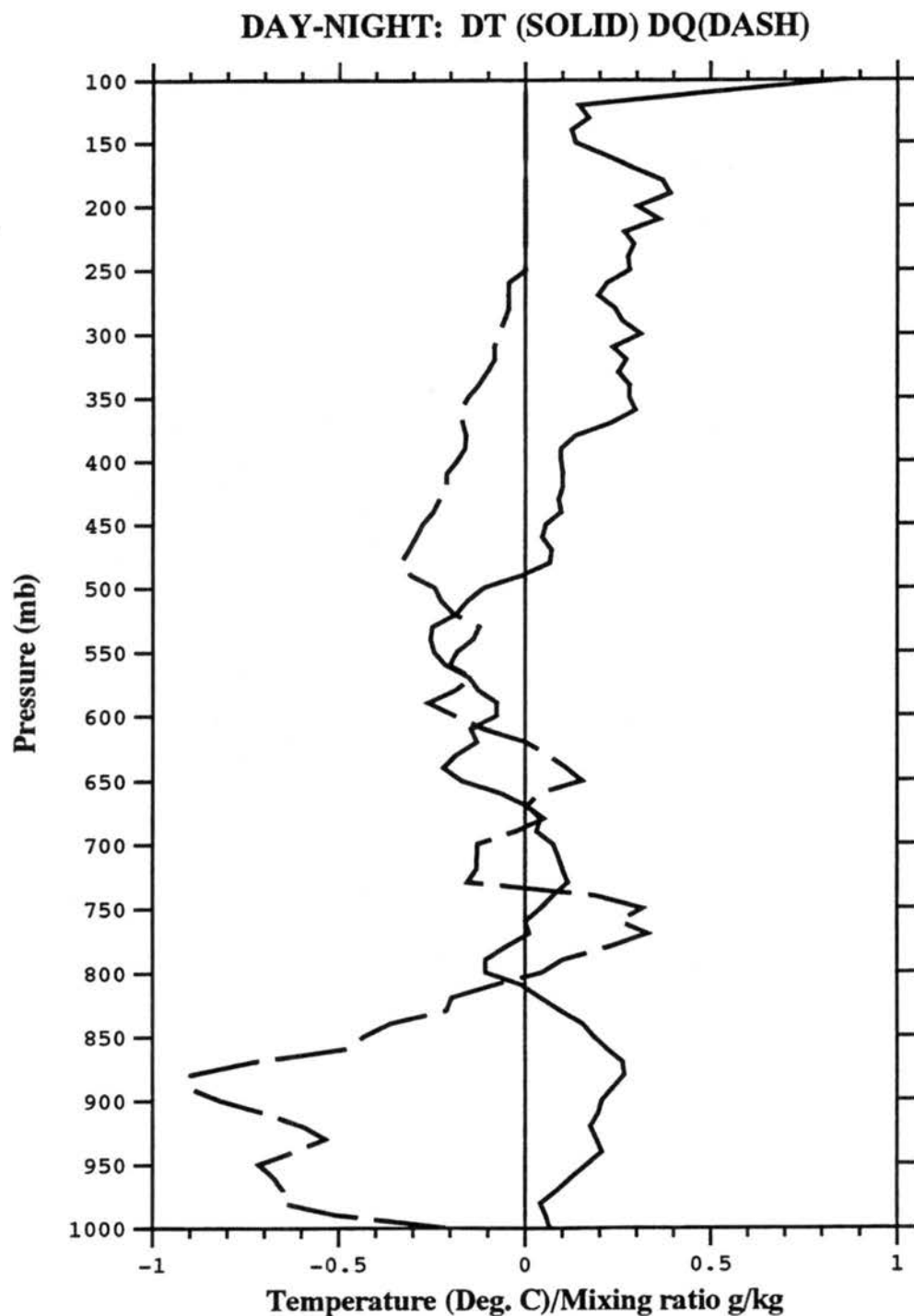


Figure 3.14 Difference ($^{\circ}\text{C}$ and g kg^{-1}) between day (11-17 Local) and night (23-05 Local) profiles of temperature (solid) and mixing ratio (dashed) in composited R/V Vickers soundings (1/28/93-2/26/93). Values > 0 imply a warmer, more moist atmosphere during the daylight hours.

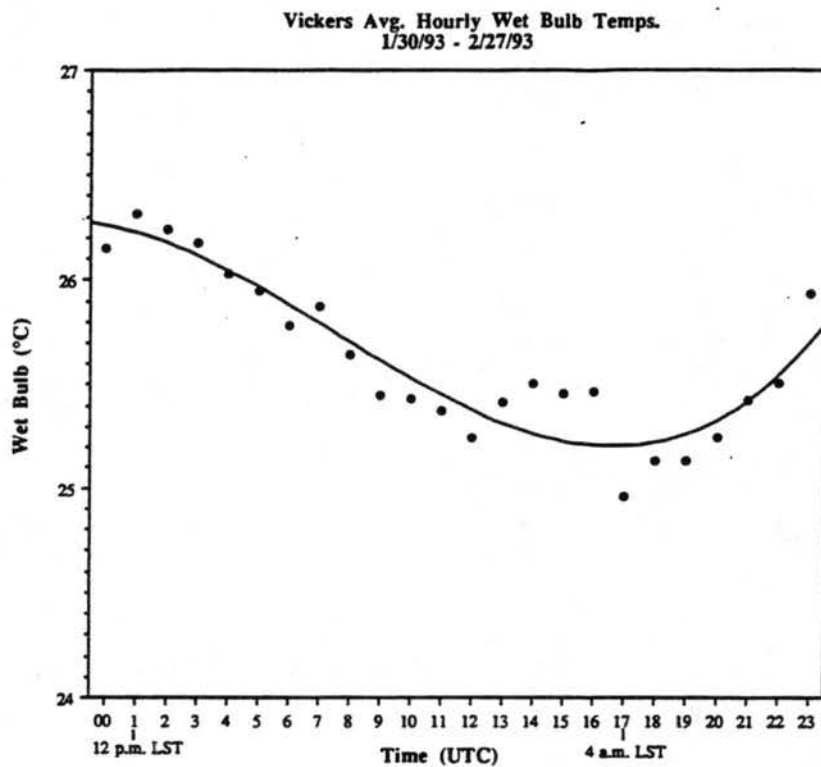


Figure 3.15 Diurnal cycle of R/V Vickers hourly wet-bulb ($^{\circ}\text{C}$; ordinate) temperatures (1/30/93 - 2/27/93). Time (UTC) is plotted on the abscissa with 1200 LST and 0400 LST indicated for reference.

Cold-Cloud Area vs. CG Flashes and Rainfall 1/1/93 - 2/25/93

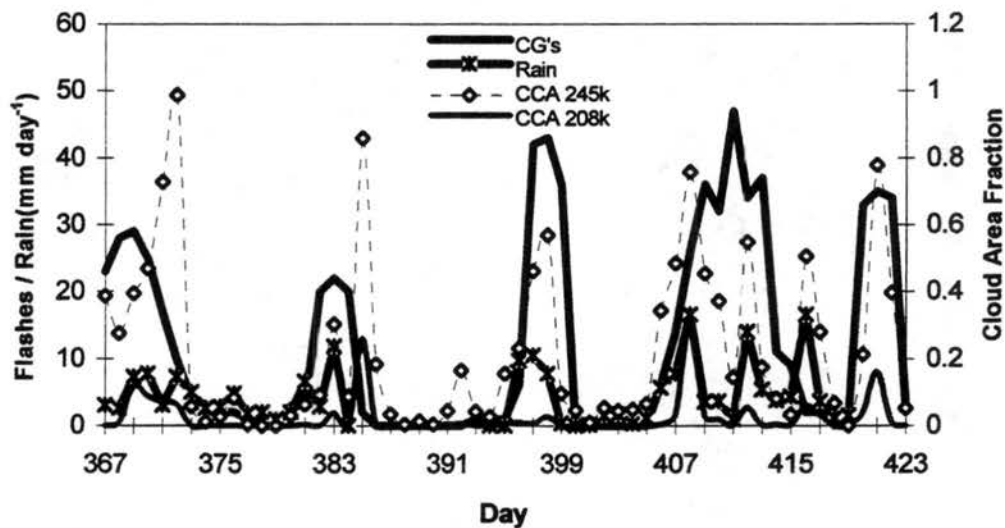


Figure 3.16 Time series (1/1/93-2/25/93) of smoothed daily CG flash count (solid) and rainfall (bold solid-x's; mm day^{-1}) on the left ordinate, and smoothed daily cold cloud area fraction at 245 K (dashed-diamonds) and 208 K (solid) on the right ordinate. Values were tabulated for a 150 km radius area centered on the R/V Vickers. Note daily rainfall values are missing for days 384-395, and after day 419.

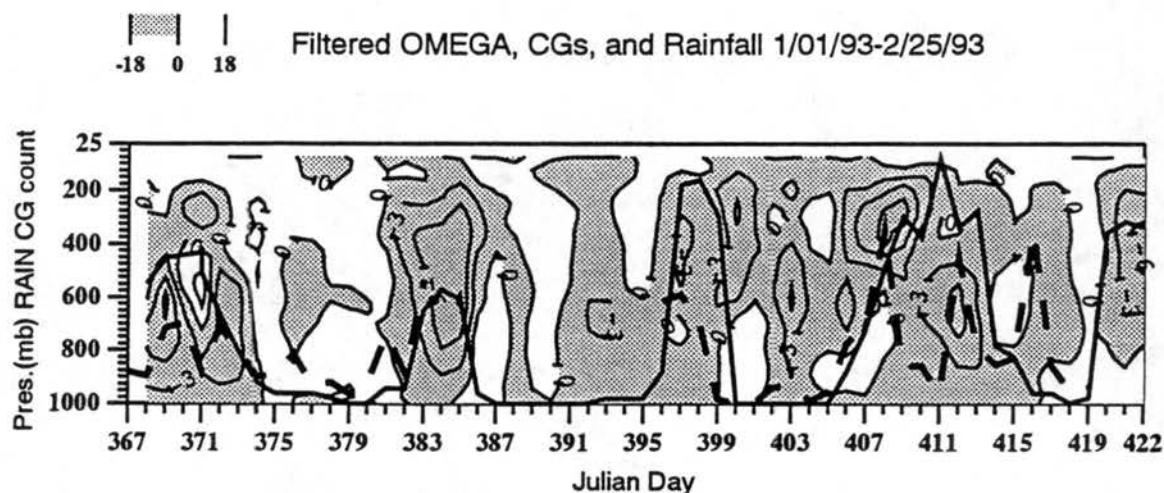


Figure 3.17 Smoothed time series of omega (mb hr^{-1} ; $1^\circ \times 1^\circ$ grid square centered on the R/V Vickers), normalized CG flash count (bold solid line) and rainfall (bold dashed line) within 150 km of the R/V Vickers (1/01/93-2/25/93). Omega is contoured every 3 mb hr^{-1} . Note daily rainfall values are missing for days 384-395, and after day 419. Sounding data from PRC ships 1, 3 and 5 were no longer available after day 417..

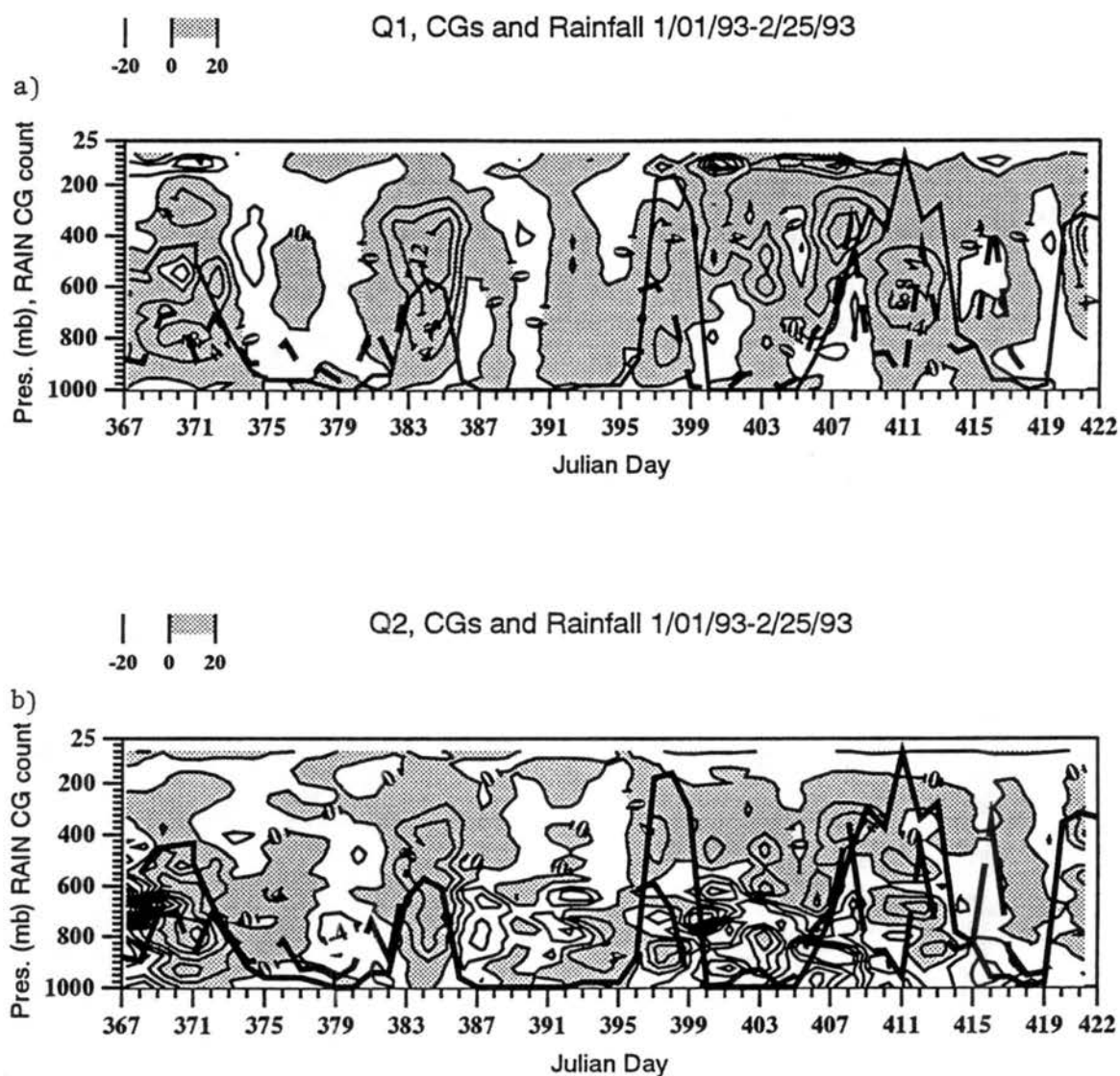


Figure 3.18 As in Fig. 3.17 but a) Q1 ($^{\circ}\text{C day}^{-1}$); and b) Q2 ($^{\circ}\text{C day}^{-1}$). (Q1 and Q2 values courtesy Mr. X. Lin, CSU).

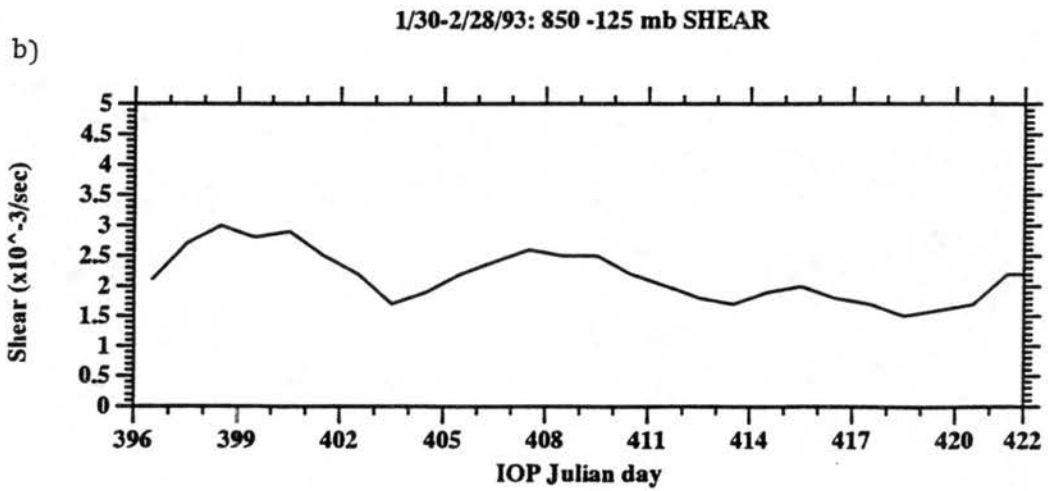
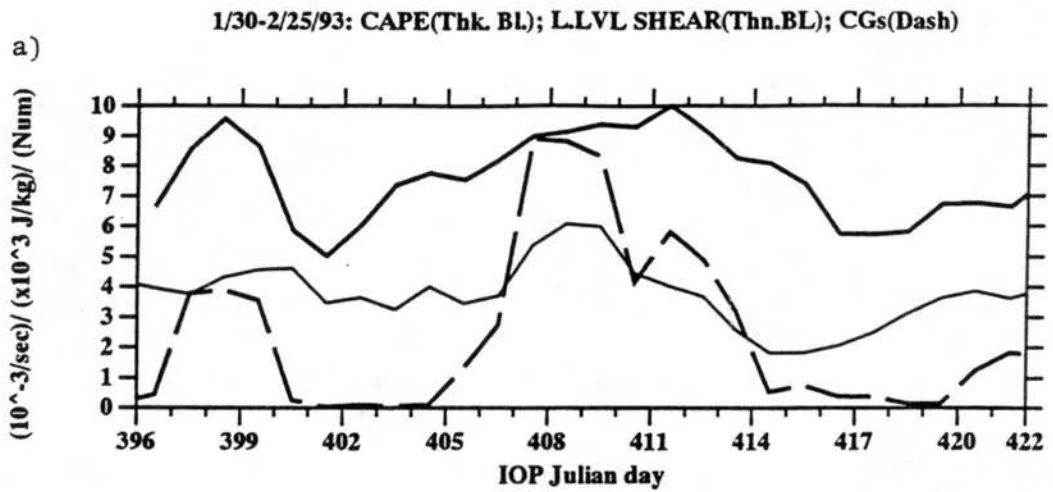


Figure 3.19 Smoothed time series of: a) low level wind shear (10^{-3} s^{-1} ; thin solid line), CAPE (J kg^{-1} ; thick solid line) and CG flash counts (300 km radius; dashed line) for 1/30/93-2/25/93 (IOP Julian-days 396-422); and b) 850-100 mb wind shear for the same time period as (a).

CG-LTG AND OLR: DAYS 1-59, 1993

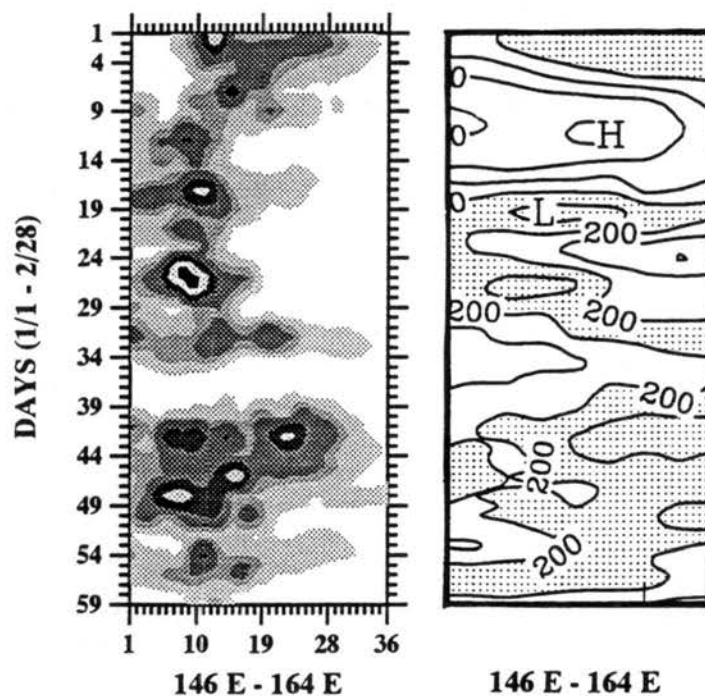


Figure 3.20 Filtered time-longitude plots of CG lightning flash count (left) and OLR (right) for latitudes 5°N-5°S, and time period 1/1/93 - 2/28/93 (Julian days 1-59, 1993). OLR values $\leq 200 \text{ W m}^{-2}$ are shaded. CG strokes are shaded in the following increments 10-50, 50-100, 100-300, 300-600, 600-900, and 900-1500. The abscissa in the CG plot is incremented in 0.5° intervals.

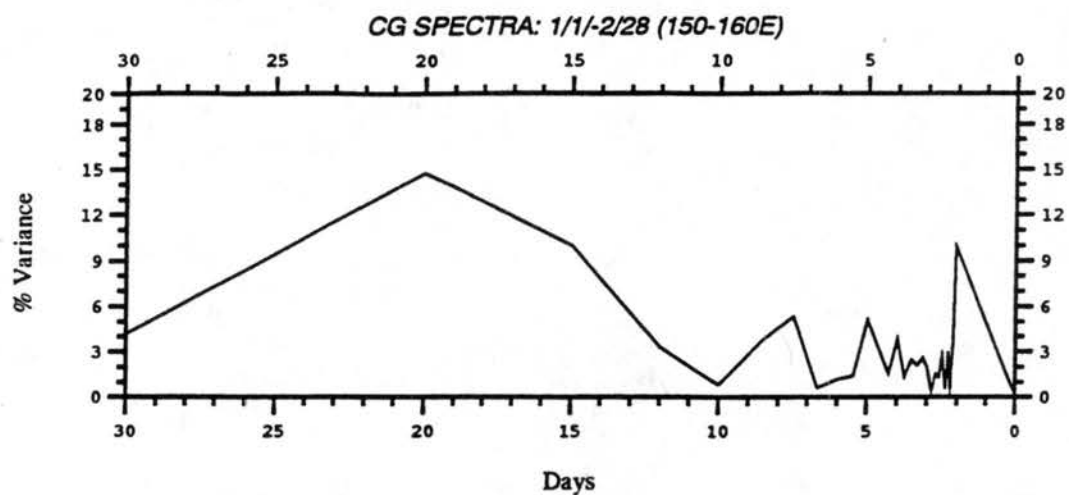


Figure 3.21 Spectrum of CG flash count between 5°N - 5°S and 150°E - 160°E for 1/28/93 - 2/28/93. Expressed as a percentage of the total variance (ordinate) explained by each period (days; abscissa).

CHAPTER 4

OBSERVATIONS OF LIGHTNING-PRODUCING OCEANIC CONVECTION

4.1 Overview of cases examined

Copious amounts of deep convection and heavy precipitation occur over the western Pacific Ocean, yet, lightning flash densities (flash counts per unit time normalized by area) over the oceans amount to approximately one tenth of that observed over land areas (e.g., Fig. 3.11; Turman and Edgar, 1982; Orville and Henderson, 1986). The *relatively* rare occurrence of lightning over the tropical ocean implies that significantly electrified oceanic convective cells are an anomaly. This point is further illustrated by considering the distribution of CG lightning activity and deep tropical oceanic convection shown in Fig. 4.1 which presents IR brightness temperatures and CG lightning locations observed on 9 February 1993 (1645 UTC) over the western Pacific warm-pool. Large areas of low IR brightness temperature in Fig. 4.1 indicate the presence of significant amounts of deep convection in the region, particularly to the south and east of the IFA. However, only a small percentage (~1-5%) of the cold-cloudiness (4°S; 159°E-160°E) in Fig. 4.1 is associated with CG lightning. From this example and many more not shown, it is clear that cloud-top temperature data alone, will not always yield “unambiguous” information regarding the *internal properties, organization or structure* of the deep convective oceanic clouds being observed.

In this Chapter we focus on different subsets (Table 4.1) of cloud/meso-scale observations of cloud-to-ground lightning, radar reflectivity, and electric field mill data for multiple cases of lightning and

non-lightning producing convection observed during COARE and CEPEX. Analysis of these cases provides important clues regarding the *distinct microphysical and cloud-scale structure* associated with lightning producing tropical oceanic convection. Kinematic analyses of two cases sampled by multiple Doppler radars (airborne and shipborne) are also contrasted as a function of their marked difference in electrical intensity (one of the systems is the lightning producing convection in Fig. 4.1). The observations of tropical oceanic convection are also contrasted with similar case study descriptions of lightning producing tropical island and mid-latitude arid-region convection. To conclude the Chapter, vertical velocity observations associated with deep convection sampled in multiple Doppler radar case studies are examined relative to their observed/expected electrical behavior.

Table 4.1: Survey of cases examined

Case Date	# Events Examined	Lightning	30 dBZ $\leq$ -10°C	Max. E-field kV m ⁻¹
Tropical Oceanic:				
3 January 93	1	Yes	Yes	11
17 January 93	1	Yes	Yes	6
31 January 93	3	1 No/ 2 Yes	No/Yes	0.8/9/NA
1 February 93	2	Yes/Yes	Yes/Yes	NA/10-15
9 February 93	3	No/Yes/Yes	No/Yes/Yes	1/5/NA
12 February 93	2	Yes/Yes	Yes/Yes	15/11
15 February 93	6	No/Yes x 5	No/Yes x 5	1/11/NAx4
19 February 93	1	No	No	-0.2
23 February 93	1	No* (*1 IC)	No	-3
Tropical Island:				
15 January 90	1 ("Hector")	Yes	Yes	NA
Mid-latitude: (SW Arid, U.S.)				
29 Sept. 94	Cells/Stratiform	Yes/Yes	Yes/No	NA
12 August 94	6	Yes	Yes	NA

4.2 The intensity and organization of deep tropical island convection as reflected in CG lightning observations

As a matter of procedural consistency, (similar to exploring the “observability” issue between CG lightning and CAPE in Sec. 3.1), we first establish a basis for expecting CG lightning in tropical oceanic convection to be related to the “convective intensity” and/or organization of oceanic convection by examining a strongly “forced” convective regime such as the maritime continent region surrounding DWN. We could rely on many previous observations of deep electrified convection in the mid-latitudes (e.g., Workman and Reynolds, 1949; Dye et al., 1986; Orville et al., 1988; Rutledge and MacGorman, 1988; Engholm et al., 1990; Rutledge et al., 1990; Schuur et al., 1991), and/or a more limited number of observations collected over tropical continental regions (e.g., Chauzy et al., 1985; Rutledge et al., 1992; Williams et al., 1992; Petersen and Rutledge, 1992), to demonstrate that lightning flash frequencies are correlated to both convective intensity and the organization of the convection. However, in an effort to 1) expand our knowledge of cloud electrical evolution relative to convective intensity/organization in the tropics; and 2) to present a clear example of what is meant by “intensity” and/or “organization” as related to the occurrence of CG lightning, we consider a particular example of the tendency for CG lightning to occur in patterns which reflect modes of convective intensity and organization of deep tropical island convection. This example is based on radar and lightning data collected during the DUNDEE over the Tiwi Islands (Melville and Bathurst; Fig. 2.7b).

First, consider Figs. 4.2a-b which show a 59 day sample of CG lightning flash location data detected over Melville and Bathurst Islands. The flashes were partitioned by time of occurrence relative to the peak in the diurnal cycle of lightning which occurs near 0600 UTC (1530 LST). Figure 4.2a shows CG activity on the fringe of the island between 0730 and 1130 LST (just prior to the initiation of a strong sea breeze), while Figure 4.2b shows the location of CG flashes over the island during the peak in

the diurnal cycle. If Fig. 4.2b is examined closely, it is apparent that meridionally banded structures appear in the lightning data. What phenomenon is associated with the banded structure?

Based on previous observations of the island convection (e.g., Keenan et al., 1988; Simpson et al., 1993; Keenan et al., 1994; Keenan et al., 1996), the banded structures observed in the lightning data (Fig. 4.2b) are likely associated with deep, highly electrified convective systems that form over the island in a regular fashion with a preferred linear organization in a north-south direction. Indeed, observations suggest the following: First, shallow isolated convective cells are generated over and along the fringes of the island along a sea breeze convergence "front" in the late morning or early afternoon. These cells are rather weakly electrified (e.g., Fig. 4.2a) but often produce heavy rain showers. The initial stage of shallow convection is subsequently followed by the development of deeper convective cells (more electrified) later in the afternoon and first order mergers of these cells form small lines and areas of deep, electrified convection. These smaller lines and areas of deep convection rapidly merge again (secondary merger process; e.g. Simpson et al., 1993), forming an intense, highly electrified squall line (often referred to as "Hector" by the island locals) that eventually assumes a more north-south shear-perpendicular orientation as storm outflows interact with the ambient low-level shear. During the second order merger and re-orientation to a north-south line, observations suggest Hector reaches its most intense stage with cloud tops and vertical velocities often exceeding 18 km and 30 m s^{-1} respectively (e.g., Simpson et al., 1993). Reflectivities of 30-40 dBZ often extend to elevations of 14 km or more (e.g., Simpson et al., 1993; Keenan et al., 1994). This regular pattern of development and organization of Hector occurs almost on a daily basis during the wet season, and has been the subject of several previous studies (e.g., Keenan et al., 1989; Simpson et al., 1993; Keenan et al., 1994; Keenan et al. 1996). It is logical to assume that the characteristic convective morphology associated with Hector should be revealed in lightning observations, and this may explain some of the banded structure shown in the afternoon lightning data (Fig. 4.2b).

Analysis of one particular case study which could be considered an “archetypical” Hector, illustrates the potential for banded structures in the lightning data (Fig. 4.2b). Figs. 4.3a-f show six Constant Altitude PPI’s (CAPPI’s) of radar reflectivity collected by the TOGA radar during DUNDEE (15 January 1990, 1400-1700 L). Coincident CG lightning positions are also indicated for the same three hour period (~30 minute intervals), covering the first and second order merger stage of Hector. The first figure (Fig. 4.3a; 1412 L), shows an initial pulse of isolated convection (no CG lightning detected) that began to develop along a sea breeze front situated along the edge of Cape Gambler over Melville Island (Fig. 2.7b). As time progressed, between 1438 L and 1509 L the convective cells on the eastern edge of the island merged to form short lines of deep, electrified convection with CG flash rates on the order of $1-2 \text{ min}^{-1}$. A similar merger process occurred between 1509 L and 1545 L for convective cells situated on the western edge of Cape Gambler (Figs. 4.2c-d), and the lighting activity subsequently increased during the intensification phase. Near 1545 L the convection became oriented in a zonal direction with CG lightning activity being concentrated in two clusters of intense convection (along $x = -10$ and $x = 20$ in Fig. 4.3c). During the next 40 minutes (1626 L) a peak in CG flash rate occurred ($\sim 2-3 \text{ min}^{-1}$) as both patterns of CG lightning and deep convection assumed a more linear orientation in the north-south direction. The system then began to weaken by 1657 L, reaching a length of $O[100 \text{ km}]$ as it rapidly dissipated into a weak region of stratiform precipitation by 1728 L. Note that approximately 60% of the CG lightning flashes occurred in a north-south oriented band associated with the merger and reorientation of the convection in a meridional direction. Given the regularity with which Hector forms and its characteristic development (e.g., Simpson et al., 1993; Keenan et al., 1988, 1994, 1996), it is reasonable to conclude that the appearance of banded structures in the lightning data shown in Fig. 4.2b is associated with the dominant morphology of the island convection.

The rapid electrical intensification of the system mimics the development of the deep convection rather well. This is illustrated in Fig. 4.4, which shows a time-height series of 30 dBZ radar reflectivity

evolution (volume at and above -10°C [~ 6.5 km] and peak 30 dBZ echo height), cloud echo-top and the number of CG flashes observed over ~ 30 minute time periods centered on each of the times shown in Figs. 4.3a-f. Trends in the height of the radar reflectivity shown in Fig. 4.4, while illustrative of the overall trend in vertical storm structure, can only be considered *approximate* as the convection sampled was located a distance of 100 km from radar (TOGA radar; Table 2.1), resulting in a coarse vertical resolution of the data (~ 1 -2 km). The data shown in Fig. 4.4 is representative of the general area (a circle of 30 km radius) centered on the two intense regions of lightning activity and deep convection that occurred at 1509 L (e.g., Fig. 4.3d; $x=0$ $y=100$). Referring to Fig. 4.4, note that the cloud-tops associated with isolated convection in that region at the beginning of the analysis (1347L-1412L) were already 11.5 km tall, yet no CG lightning had occurred. The general lack of CG lightning at this time is consistent with the relatively small volume of 30 dBZ echo (a proxy for hydrometeor mass) that existed at and above the height of the -10°C level (into the mixed phase region of the cloud). However, as the storms continued to develop, the 30 dBZ echo volume, peak 30 dBZ height (~ 15 km), cloud-top (~ 20 km), and CG flash count rapidly increased, reaching coincident peaks (Fig. 4.4) as the convection reached maximum intensity during the reorientation stage (1626 L; Fig. 4.3e). The weakening of the north-south oriented convective line at 1657 L is reflected in the decreasing 30 dBZ storm volume and lightning activity shown in Fig. 4.4.

We now have a sense of what is implied by the term “intensity” when referenced in discussions of the presence or absence of lightning in deep convection. Clearly, the term “intensity” refers to the vertical development of the convection (e.g., Fig. 4.4), which we will describe in terms of the height of the 30 dBZ surface and the volume contained by this surface (e.g., Larson and Stansbury, 1994; DeMott, 1996; Fig. 4.4). The volume contained within the 30 dBZ reflectivity “core” is assumed to be associated with the largest precipitation mass in the convective cloud. Additionally, the height of the 30 dBZ reflectivity core has been found to be well correlated to the location of IC flash initiation in non-severe

Florida thunderstorms (Murphy et al., 1996). Since significant precipitation mass situated above the height of the -10°C level is likely to be associated with a robust mixed-phase process, and this is conducive to efficient electrical charging of the cloud and the production of lightning (e.g., Murphy et al., 1996), the term “intensity” can be used to describe the “electrical activity” or the “core” (e.g., dBZ’s > 30) structure of the storm.

4.3 Observations of lightning producing and non-lightning producing tropical oceanic convection

In this section we present a subset of Doppler radar and electric field observations collected during TOGA COARE from five days listed in Table 4.1 (31 January, 1 February, 15 February, 19 February, and 23 February 1993). Cases from the first three days are associated with lightning producing oceanic convection that occurred during periods of high CAPE (e.g., Fig. 3.5), and the latter two are examples of very deep, but non (or marginally) lightning-producing oceanic convection that occurred during periods of lower CAPE (Fig. 3.5). For 15 February, radar and CG lightning data for four separate convective systems (no electric field data) have been composited. The composite is presented in the form of a time-height-reflectivity diagram with the average number of CG flashes that occurred in 20 minute time intervals also indicated. The data from this day are composited in order to clearly illustrate, *via a single diagram*, the vertical and electrical evolution of the storms that occurred on this day. Similar time-height diagrams for the individual case study days are also displayed.

4.3.1 Lightning-producing convection: 31 January 1993 (case 1)

Much of the convection observed in the IFA on 31 January 1993 (and during the following 24 hours) was associated with the passage of an ISO, high CAPE and moderate to large low-level vertical

shear of the horizontal wind (e.g., Figs. 3.5, 3.19). The presence of a dissipating nocturnal/early morning convective system (which was deep, but did not produce any CG lightning; e.g., Table 4.1; Petersen et al., 1993) produced a mesoscale boundary oriented in a northwest to southeast direction approximately 15 km to the northeast of the R/V Vickers. The boundary was evident on the MIT radar display as a line of weak reflectivity (0-10 dBZ). This line served as the focusing point for isolated deep convective cells for the next 12 hours. Convection began to intensify along the boundary after sunrise. Near 2300 UTC several cells began to rapidly grow along the boundary, and it was noted in the MIT radar logs that the reflectivity at the surface associated with one cell increased from 0 dBZ to 48 dBZ in approximately 5 minutes. Two cells in the vicinity of the radar produced a few IC flashes (as indicated by a hand held optical lightning detector) and were associated with radar reflectivities in excess of 30 dBZ that extended well above the height of the -10°C level. Echo-tops were approximately 12-14 km.

One relatively isolated cell developed approximately 15 km to the northwest of the Vickers near 2250 UTC. As the cell intensified, it maintained a virtually constant range of 10-15 km from the ship (Fig. 4.5a) and drifted along the boundary toward the southeast. Examples of the vertical radar reflectivity structure of the cell are shown in Figs. 4.5b-c. At 2332 UTC (Fig. 4.5b) the cell possessed a very "hard" top with 30 dBZ reflectivities extending to nearly 10 km (the highest elevation angle in the radar scan just barely "topped" the cell during this stage of growth) and an echo-top extending to approximately 12 km. At first glance, the structure of the 30 dBZ echo aloft seems somewhat suspect in that it is disconnected from the lower part of the 30 dBZ reflectivity core. It is possible that the radar captured the tail end of a burst in vertical development near this time, followed by the initial descent of the main core of the storm (cells were developing rapidly with 10 minute average echo-top rise rates on the order of $3\text{--}4\text{ m s}^{-1}$, consistent with the presence of high CAPE's in a strongly capped boundary layer; e.g., Fig. 3.5). Conversely, there was heavy precipitation noted in the lower levels of the storm near the 2 km level ($\sim 50\text{ dBZ}$) and it may be that the echo is an artifact of radar side-lobe contamination. However,

the 30 dBZ echo at 8-10 km elevation does have a distinct "shape" and does reside in a region of 20 dBZ echo that extended to the top of the storm (coincident with the maximum vertical extent of 40 dBZ echoes above the freezing level). Hence it is reasonable to conclude that the 30 dBZ radar reflectivity structure is real. Near 2351 UTC (Fig. 4.5c) the 30 dBZ echoes aloft were descending and 20 dBZ reflectivities reached their maximum vertical extent prior to a rapid collapse in the next 10 minutes (e.g., by 0000 UTC) as the storm began to dissipate.

Since this isolated cell maintained an approximately constant range from the ship (~15 km) as it intensified, the electric field mill on the R/V Vickers was ideally located for detecting coincident trends in the magnitude of the vertical component of the electric field (E-field; Fig. 4.6). Note that negative (positive) electric field values in Fig. 4.6 indicate downward (upward) movement of a positive test charge under the influence of the E-field. In a clear or relatively undisturbed atmosphere (no dust, electrified clouds etc.), the E-field will be maintained at the "fair-weather" value of approximately -100 V m^{-1} . For comparison to the E-field evolution, the time-height series of radar reflectivity development associated with this cell is also indicated in Fig. 4.6.

The initial fair-weather value of the E-field (-100 V m^{-1}) in this case remained unperturbed for the first 20 minutes of slow development (Fig. 4.6). It is not until 30 minutes into the development of the cell that the E-field began to build rapidly foul (e.g., positive values; upward directed E-field) coincident with the extension of reflectivities $\geq 30 \text{ dBZ}^1$ above the height of the -10°C level (e.g., deep into the mixed phase region of the cloud). Interestingly, the growth of the E-field at the surface seemed to mimic the growth of the 30 dBZ contour with height, with a lag of about 10-15 minutes. At minute 45 (2332 UTC; Fig. 4.5b), the height of the 30 dBZ echo reached a peak while the top of the storm continued to grow

¹ In a Rayleigh scattering regime, a monodisperse distribution of 1 mm diameter raindrops in a concentration of 1 L^{-1} will produce a reflectivity of 30 dBZ. As these millimeter diameter particles begin to freeze and rime in the presence of smaller cloud-ice particles, the precipitation ice-mass requirement for charging of the convection based on a continuous collection model is satisfied (e.g., Takahashi, 1984).

until 2342-2351 UTC (Fig. 4.5c). As the 30 dBZ echo began a rapid descent between 2332 and 2342 UTC, the electric field experienced rapid growth and two in-cloud lightning flashes occurred in the storm (indicated as transients in the E-field) near 2338 UTC (~min. 52) and 2345 UTC (~min. 59) near the time of the peak in the foul weather E-field (9 kV m^{-1}). The rapid growth in the E-field and associated lightning flashes occurred coincident with the separation of the descending 30 dBZ core and rapidly ascending upper portion of the cloud (e.g., reflectivities ≤ 20 dBZ were still increasing in altitude). One more flash occurred at 2358 UTC (min. 72) as the remainder of the storm collapsed and moved southeast (note that the decrease in the E-field after approximately 0005 UTC (min. 79) may be associated with storm advection to the southeast in addition to the storm decay). No CG flashes were detected by the ALDF network in association with this storm.

The evolution of the electric field when considered relative to the coevolving reflectivity structure suggests the following sequence of events with regard to the electrification of the convection and storm microphysics/kinematics:

- 1) The reflectivity core of the storm, indicated by reflectivities ≥ 30 , developed *upward* into the -10°C to -20°C region of the storm, indicating that raindrops, initially formed below the 0°C level, were lofted into the supercooled region of the cloud (e.g., Jameson et al., 1996; Ramachandran et al., 1996).
- 2) As the lofted raindrops began to freeze and rime, precipitation-ice developed in the mixed phase region of the storm (as indicated by the presence of reflectivities ≥ 30 dBZ above the height of the -10°C level). In association with the freezing of raindrops and subsequent accretional growth, ice-multiplication processes were likely initiated (e.g., rime splintering between -3°C and -8°C ; Hallett and Mossup, 1974). Hence, conditions were established for efficient non-inductive charging to take place.

- 3) A rapid separation of charge occurred on the cloud-scale as descending precipitation ice in the storm core (e.g., 30 dBZ contour) began to separate from ascending smaller snow and ice-crystals resident in the upper portions of the cloud. The separation of charge caused E-field magnitudes to reach dielectric breakdown strengths in upper regions of the cloud and in-cloud lightning occurred (e.g., Murphy et al., 1996). Generation of substantial layers of charge in the cloud caused the surface electric field to build rapidly foul (e.g., indicating the presence of substantial negative charge in the cloud).

In subsequent observations, we will see that this sequence of events is repeatable.

4.3.2 Lightning-producing convection: 1 February 1993

On 1 February 1993, deep widespread convection occurred within range of the MIT radar (in association with passage of an ISO as mentioned in Sec. 4.2.1) and was organized into eastward moving short lines with random orientations (area coverage increased relative to the convection discussed in Sec. 4.2.1). Cloud-to-ground lightning activity occurred in isolated clusters throughout the COARE domain on this day. An example of the electrified convection and CG lightning that occurred this day is shown in Fig. 4.7, which indicates that cloud-to-ground flashes were located in the vicinity of reflectivity maxima situated approximately 13 km and 64 km northeast of the MIT radar. A time series of vertical cross-sections through the deepest parts of the two sets of lightning producing cells are shown in Figs. 4.8a-f.

Note the change in height of the 30 dBZ reflectivity core with time in Figs. 4.8a-f. Similar to the 31 January case (Sec. 4.2.1) and previous observations of electrified convection over regions of the Maritime Continent (Williams et al. 1992; Rutledge et al. 1992; Zipser and Lutz, 1994), the reflectivity cores extended well into the mixed phase region of the clouds coincident with the development of large surface electric fields and lightning. This is an indication that updraft speeds and hydrometeor masses

were large enough to allow significant charge generation to occur in the cloud (e.g., Williams and Lhermitte, 1983). It is also interesting to note that the peak in CG activity and the maximum reflectivity core depth (as sampled by the radar every 10 minutes) appeared to be slightly out of phase in both sets of convective cells (Figs. 4.8 a-f), with cores reaching peak heights as much as 0-10 minutes prior to the peak in CG activity (after 0532 UTC for both sets of cells). The trend of increasing CG flash rate with the peak and subsequent descent of the reflectivity core in the mixed phase region of the cloud has been observed in several other cases from COARE, and is also consistent with the findings of many previous studies conducted in other regions of the globe (e.g., Workman and Reynolds, 1949; Lhermitte and Williams, 1984; Goodman et al. 1988; Williams et al. 1989; Carey and Rutledge, 1995).

Figure 4.9 shows analog electric field data for the period 0428-0548 UTC. The peak in the E-field ($10\text{--}15\text{ kV m}^{-1}$) occurred near 0447 UTC as a rapidly electrifying cell moved directly over the ship. Note also the rapid excursion to a fair weather E-field between 0450 UTC and 0455 UTC. These field excursions were likely associated with precipitation (e.g., Krehbiel, 1986) as rainfall rates at the ship exceeded 70 mm hr^{-1} near these times. There were some indications of lightning in the electric field trace from 0445-0450 UTC, though no thunder was reported at the ship. The first cloud-to-ground lightning flash (confirmed visually) occurred near 0515 UTC, and was associated with one of the cells shown in Fig. 4.7 (cell located 13 km northeast of the radar). This CG flash occurred nearly coincident in time with the first well-defined, electric field perturbation that was due to lightning and detected by the field mill (~ 0516 UTC). Since the field mill sensitivity decreased rapidly with range, it is likely that the higher amplitude perturbations detected by the field mill were associated with CG's that occurred near the ship (i.e., within a range of 20 km). After comparing the time and number of flashes detected by the field mill with those detected coincidentally by the ALDF network, it is apparent *for this case* that the network detected all of the CG flashes that occurred near the ship (this assumes that the higher amplitude perturbations registered by the low gain channel of the field mill were in fact CG's and not in-cloud

flashes). Between 0519 and 0534 UTC (when no CG's were detected by the network) the field mill detected many smaller amplitude perturbations in the electric field which were probably in-cloud (IC) flashes.

The coevolving electric field signatures, lightning, and reflectivity fields associated with the cells observed near the R/V Vickers on 1 February 1993 suggests that IC flashes occurred prior to the majority of the CG flashes (with the exception of the CG at 0516 UTC), during intensification of the cells aloft (similar to the 31 January case, with the exception that no CG's occurred in that storm). Only after the cells reached the mature stage (peak echo-top heights, descent of the echo core), did both the field mill and the COARE ALDF network indicate a switch to predominantly CG lightning. The observed tendency for the peak in CG lightning to follow that of the IC lightning is similar to observations previously collected in the mid-latitudes by Goodman et al. (1988) and Williams et al. (1989).

One important thing to note with regard to lightning producing convection observed during COARE is that the oceanic cells often produced heavy rainfall at the surface well prior to the occurrence of any lightning (similar to observations by Chauzy et al. 1985). This was likely due to the presence of efficient warm rain coalescence in the COARE convective cells, a phenomenon often associated with maritime convective rainfall (see Young, 1993). The occurrence of lightning (both IC and CG) in the majority of the cases examined from COARE (e.g., 31 January 1993, 1 February 1993), took place only *after* significant hydrometeor mass developed in the mixed phase region of the cloud above the -10°C level. This strongly links the occurrence of lightning in these cells to the development of moderate concentrations of precipitation sized ice particles aloft, but not necessarily to the presence or intensity of rainfall at the surface.

4.3.3 Lightning-producing convection: 15 February 1993

On 15 February, four lightning producing storms occurred within 100 km of the MIT radar. The radar reflectivity structure and observed cloud-to-ground lightning patterns associated with four of the cases were very similar. Furthermore, each of the four cases exhibited a life cycle of approximately 60 minutes. Because of the similarities in radar structure, lightning evolution, and lifecycle, we chose to composite observations of the four cases into one "representative electrified cell" (REC). The resulting composite (Fig. 4.10) represents the growth and decay with time of the REC in terms of averaged quantities such as the radar reflectivity, CG *flash* rate over a 20 minute period centered on the time of the radar scan, and rain volume (calculated from reflectivity at the 2 km level using the convective Z-R relationship reported in Thiele et al., 1994).

Several characteristics of the lightning producing convection observed during COARE are illustrated by the composite REC (Fig. 4.10). First, note that 30 dBZ reflectivities extended above the height of the -10°C level (~6.5 km) for approximately 20 minutes coincident with an increasing CG flash rate and increasing echo-top height (the latter defined by the 1 dBZ contour). Interestingly, a sharp increase in the number of CG's took place as the 30 dBZ contour reached its peak elevation. Furthermore, as the 20, 30 and 40 dBZ reflectivities began their descent, the cloud-to-ground lighting rate peaked, and then fell off as the REC began to dissipate. The volume of rainfall produced by the REC exhibited only minor variations relative to the cloud-to-ground lightning evolution, with only a slight peak in rainfall occurring coincident with the peak in CG frequency. These observations further demonstrate the close coupling between the vertical distribution of hydrometeor mass in the mixed phase region of a cloud and the CG flash rate.

4.3.4 Weakly electrified convection: 19 and 23 February 1993

The relationship between cloud-top height, satellite TBB and degree of electrification in tropical oceanic clouds is dubious (e.g., Fig. 4.1) because 1) expansive areas of cold TBB's observed in the tropics are often associated with stratiform cloud shields, raining or non-raining, that occur *after* the most intense period of convection in a cloud cluster and in general, produce considerably less lightning than their convective counterparts (e.g., Rutledge and Petersen, 1994); and 2) tropical oceanic convection is often very deep (i.e., cloud-tops > 15 km) but only weakly electrified (i.e., producing little or no lightning; Rutledge et al., 1992; Williams et al., 1992). In this section we illustrate the problem of "gauging" convective intensity based on using cloud-top brightness temperatures as a proxy. We present representative observations of deep (cloud tops in excess of 15 km) but weakly electrified convection (little if any lightning produced) associated with two different cloud systems that occurred on 19 February 1993 and 23 February 1993. When the following observations are contrasted with those of Secs. 4.3.1-3, the potential utility of lightning observations for indicating the bulk characteristics of the vertical structure *inside* of deep tropical oceanic convective clouds will become apparent.

We begin with a discussion of deep convection that occurred on 19 February 1993. Near 1200 UTC on 19 February a tall *but weakly electrified* cloud system occurred near and over the location of the MIT radar and electric field mill (Fig. 4.11a-b). These particular clouds were associated with very strong surface forcing concentrated along a propagating boundary (Fig. 4.11a), and possessed radar echo-tops of 15-17 km (Fig. 4.11b). The system occurred in an environment of relatively increased deep tropospheric shear, decreased low-level shear, and moderate CAPE (e.g. Fig. 3.19). The combination of reduced low level shear and increased tropospheric shear (westward direction) may explain the predominant westward sloping radar reflectivity structure associated with this system (Fig. 4.11b). No lightning was reported visually by the bridge crew for this system (it was dark so lightning would have been easily observed), though the COARE ALDF network detected two CG flashes approximately 100 km to the northwest of the R/V Vickers between 1320-1330 UTC. Figure 4.12 shows a time-height

series of the reflectivity associated with an area of cells that originally developed 40 km to the northwest of the MIT radar at 1111 UTC, and subsequently moved to approximately 10 km northeast of the MIT radar by 1321 UTC. It is clear in both Figs. 4.11b and 4.12 that the clouds reached great vertical extent (17 km in Fig. 4.12), however, the most significant hydrometeor mass in the cloud (as evidenced by -10°C level. That is, the radar data suggest that pronounced lofting of millimeter sized raindrops above the height of the -10°C level did not occur, and it is therefore likely that glaciated conditions consisting of smaller ice particles (small, low-density graupel, snow, cloud-ice) and little cloud-water existed above the height of the -10°C level. Thus conditions would not have been favorable for significant electrification to occur (Fig. 4.11b). Indeed, the E-field data for this case (not-shown) revealed only slightly enhanced (-200 V m^{-1}) fair weather fields in the vicinity of cells as they passed within 15-20 km (north) of the R/V Vickers (Fig. 4.11a).

A storm similar to the 19 February 1993 case occurred on 23 February 1993, but this system had a more interesting electric field signature. This system was comprised of complex broken lines of deep convection with an associated trailing region of stratiform precipitation (Figs. 4.13a-b, 4.14a-b) that was associated with both detrained condensate and decayed convection left in the wake of the rapidly moving squall-line. The squall-line moved in an easterly direction passing over the R/V Vickers near 1530 UTC. As the system moved over the ship the electric field (Fig. 4.15) showed very little deviation from fair-weather values. Once the system moved east of the ship, a single intra-cloud (IC) lightning flash was observed by the bridge crew. This flash was also detected by the R/V Vickers electric field mill at 1641 UTC (0341 L; Figs. 4.13a-b). These observations were approximately coincident with the time of maximum vertical development observed in the leading line of convection (e.g. Fig. 4.13b). Shortly after 1641 UTC, the E-field began to increase toward larger negative values (indicating the advection or development of net positive charge overhead). The E-field reached sustained maximum fair-field values

of -2 to -3 kV m^{-1} between 1700 and 1815 UTC (Fig. 4.15) as a region of deep stratiform precipitation developed over the location of the ship (Fig. 4.14a-b; 4.15). The development of the strong fair-field occurred coincident with the rearward advection (strong storm relative flow toward the west) of hydrometeors and decaying convection from the leading convective line. Prior to a gradual decay of the stratiform precipitation region, the field reversed to foul field values of $\sim 1 \text{ kV m}^{-1}$ (in the fashion of an end of storm oscillation; e.g., Krehbiel, 1986).

The growth of the strong fair weather field in the 23 February case (indicating the presence of a significant amount of positive charge aloft) when considered relative to the observation of a single IC lightning flash, suggests that the convection which occurred on 23 February may have been on the “verge” of becoming highly electrified. The cross section of radar reflectivity shown in Fig. 4.13b indicates echo tops of 18 km or more, 20 dBZ reflectivities that extended to altitudes approaching 13 km, and a 30 dBZ contour that tended to hover around -5° to -10°C (6-6.5 km). Hence, radar reflectivities indicate that the volume of 30 dBZ reflectivity above the -10°C level (and mixed phase mass) was relatively small. Though we can only speculate, the data suggest that there may have been *just* enough precipitation ice to produce the bare minimum of charge required to initiate a single lightning flash (and a subsequent manifestation of residual positive charge over the radar later in the period). It is interesting to note that this storm seemed to be on the “threshold” of significantly electrifying, while Fig. 3.5 suggests that the CAPE and the CG flash count during this time period (23-25 February) were on an increasing trend (the wind-shear also appeared to be more favorable for development during this time period; e.g., Fig. 3.19).

The observations of little or no lightning associated with these deep (cloud-tops exceeding 15 km) tropical oceanic convective systems suggest a note of caution is warranted when parameterizing the occurrence of lightning in large scale models based on cloud-top height (e.g., Price and Rind, 1994).

This also applies to monsoon systems (similar to oceanic convection) that occur over the tropical maritime continent region of northern Australia (e.g., Rutledge et al., 1992; Williams et al., 1992).

4.4 Mixed phase processes and electrification: Contrasting observations from the arid southwestern U.S.

In order to provide a distinct contrast to the above observations, observations of convection from central Arizona were analyzed using the Phoenix, AZ NEXRAD radar data and NLDN lightning data. These analyses provide an extreme contrast in radar-echo, precipitation, and electrical development between arid-region thunderstorms and tropical, oceanic convection. The cases analyzed are listed in Table 4.1. Herein we will summarize important characteristics of the observations analyzed for the 12 August 1994 cases. The 12 August 1994 cases consisted of six downburst producing thunderstorms that occurred in the Phoenix metro area. For the purpose of contrasting these cases with the tropical oceanic cases discussed in Sec. 4.3, the focus of the analysis in this section will be placed on the development and lifecycle of vertical precipitation structures (as indicated by radar) relative to observed CG lightning (electrical) tendencies. In particular, we examine one specific convective system with regard to its vertical profile of radar reflectivity and electrical evolution. The remaining observations were composited into a single time-series profile similar to the REC (Fig. 4.10) created for the electrified oceanic convection discussed in Sec. 4.3.3.

A vertical cross-section time-series of radar reflectivity is shown in Fig. 4.16a-f. These vertical cross-sections were collected during the mature to dissipating stages of development for a deep convective cell that occurred between 0600 UTC and 0700 UTC on 12 August 1994. The storm produced copious amounts of CG lightning, heavy rainfall and a strong divergent wind signature (macroburst) near the surface (Fig. 4.17). In contrast to the convection observed in the tropics (e.g.,

Figs. 4.5, 4.8, 4.10, 4.12, 4.13) where radar echoes initially appeared below the freezing level and later built upward into the supercooled region of the cloud, initial precipitation echoes in these clouds occurred at elevations near or above the height of the 0°C level.

The first portion of the “mature” phase of this particular storm (12 August 1994; Fig. 4.16a) was associated with the occurrence of light rainfall at the surface (as evidenced by the presence of 30–45 dBZ reflectivities) and the initiation of a few CG flashes (Fig. 4.17). Note that the precipitation development in this case was initiated in a volume of high reflectivities (45–55 dBZ) situated primarily above the height of the freezing level (Fig. 4.16a). The development of the precipitation relative to that of CG lightning occurred as follows: 1) Fig. 4.16a; a region of 45–55 dBZ echo (likely graupel and small hail) formed in the 0°C to -20°C region of the cloud and the number of CG lightning flashes began to increase (Fig. 4.17); 2) Fig. 4.16b; the precipitation region aloft intensified as graupel and hail rapidly grew by accretion (≥ 55 dBZ echo at 7 km in Fig. 4.16c). This was accompanied by a rapid increase in the CG lightning frequency and outflow winds at the surface (Fig. 4.17) as descending ice mass melted and liquid water-mass evaporated (e.g. Srivastava, 1987); and 3) The precipitation region aloft increased in volume (Fig. 4.16d) while descending to the surface (Figs. 4.16e-f) coincident with peaks in CG lightning, macroburst winds at the surface, and a rapid increase in surface rainfall (Fig. 4.17).

In *contrast* to the tropical oceanic convection, rainfall in the 12 August 1994 cases was generated predominantly by a mixed-phase ice-process (e.g., accretional growth of ice particles which later descended as larger precipitation-ice particles and melted to form rainfall; Young, 1993). In a fashion *similar* to the tropical oceanic REC (Fig. 4.10), peaks in CG lightning and rainfall occurred just after the echo core began to descend through the -10°C level (~ 6.5 km). Note that the mid-latitude case was also associated with a strong (40 ms^{-1}) divergent wind signature, the magnitude of which was *not* observed in the tropical cases (peak storm outflows at the R/V Vickers rarely exceeded 15 m s^{-1} in magnitude). An example of the characteristic process of upper level precipitation development (as indicated by trends in

radar reflectivity), followed by descent of the reflectivity core, coincident development of a divergent wind signature, increased CG lightning and heavy surface rainfall (reflectivity used as a proxy) is shown in Fig. 4.18 in composite form for four of six cases examined from 12 August 1994 (the case in Fig. 4.16-17 is not included in this composite).

From a cloud electrification viewpoint, the generation of precipitation-ice by a mixed phase process in mid-latitude continental clouds (Fig. 4.18) is conducive to efficient electrification of the cloud via non-inductive charging processes (i.e., riming of ice crystals to form graupel, collisions of graupel with ice, descent of graupel which becomes rainfall as it melts). In tropical oceanic clouds, a significant portion of the precipitation sized ice-particles must be initiated by first "lofting" liquid precipitation (e.g., raindrops formed by warm-rain coalescence) and cloud-water above the 0°C level, and then freezing the drops. Hence tighter initial constraints are placed on the updraft strength of tropical oceanic clouds relative to the electrification process (i.e., the updraft magnitude has to be large enough to lift and suspend millimeter sized raindrops above the 0°C level). In these two climatological extremes, the dominant precipitation process is fundamentally different, yet the examples shown in Figs. 4.5-4.18 illustrate the importance of a vigorous precipitation-ice phase in the development of lightning-producing convection, *regardless of the environment the clouds form in.*

4.5 Multiple Doppler observations of tropical oceanic MCSs

Considering the observations of lightning producing tropical oceanic, tropical island/continental and mid-latitude continental convection, the term "vertical development" as related to electrical intensity, is directly related to some degree of mixed-phase microphysical development. Since the observations presented in Sec. 4.2 suggest that the lofting of drops into the supercooled region of tropical oceanic clouds is related to their electrical development (this has also been observed in the electrification of

Florida cumulonimbi; e.g., Ramachandran et al., 1996), the development of a robust mixed-phase process (and hence lightning) in oceanic convection is likely to be a function of the updraft strength at present lower and middle levels in the cloud. Therefore, we turn our attention to the kinematic aspects of lightning and non-lightning producing tropical oceanic convective clouds.

In this section we discuss dual and quad-Doppler radar analyses (Jorgensen et al., 1994), together with electric field and cloud-to-ground lightning (CG) observations for two oceanic convective squall lines observed during TOGA COARE on February 9, 1993. Both systems were deep (maximum radar echo cloud tops of 16 and 18 km respectively), shear-parallel squall lines that occurred approximately 8 hours apart and were separated by a distance of over 400 km. However, one system was only weakly electrified as evidenced by the lack of lightning observations (system located near 2°S 155.5° E in the IFA, hereafter referred to as the “ship case”), and the other highly electrified (4.5° S 159°E and referred to as the “aircraft case”). Since both systems were sampled by multiple Doppler radars (ship case- shipborne dual Doppler; aircraft case- airborne quad-Doppler), the three dimensional wind fields could be determined and the observed kinematic structures and electrical properties of the two systems compared. In this manner we can examine relative differences in kinematics between the two systems and relate these differences to observed electrical activity.

4.5.1 Analysis Method

Fields of radar reflectivity and three dimensional wind velocity, derived from dual (ship case and aircraft case) and quad Doppler techniques (aircraft case), were examined in horizontal and vertical planes (see Sec. 2.1.5) together with CG lightning locations detected by the COARE ALDF network. In the horizontal plane, radar reflectivity and wind-fields were interpolated to a 1 x 1 km grid for the ship case, and a 1.5 x 1.5 km grid for the aircraft case. In the vertical plane, cross-sections of radar

reflectivity and vertical velocities were interpolated to a Cartesian grid with 0.5 km and 1 km grid spacing for the ship and aircraft cases respectively. Surface electric field data collected aboard the R/V Vickers were also examined for the ship case.

In order to more closely compare the intensities of the two systems, composited fields of vertical velocity and radar reflectivity were generated from six volumes of radar data for the ship case, four volumes for the aircraft case (e.g., Smull et al., 1994), for similar periods in each of the storm lifecycles (1012 UTC - 1212 UTC, ship case; and 1614 UTC - 1705 UTC, aircraft case). The composites were then partitioned into "total", "convective", and "non-convective" representations of the cloud-systems. The partitioning of the radar data is based on contoured frequency depiction's of volumetric reflectivity (e.g., Yuter and Houze, 1993) as applied in an algorithm developed by DeMott (1996) (c.f., Cifelli, 1995). The comparisons of vertical motion fields are complicated by the difficulty involved in selecting radar volumes from each cloud system that represent similar stages of convective evolution, and also by potential errors introduced through the use of different measurement platforms, sampling strategies and wind-synthesis techniques.

4.5.2 Storm Environments

Stream line analyses at sea level for 1200 UTC 9 February 1993 (not shown) indicated that a broad area of confluence existed in the environments of both systems, being strongest in the vicinity of the aircraft case (near 4.5°S, 159°E). Gridded sounding data indicated that vertical velocities in the mid-troposphere were weak, but directed upward in the vicinity of both systems. Similar to the low level confluence, low-level moisture convergence (Figs. 3.8b and 3.9b), and mass divergence fields, were stronger in the vicinity of the system sampled by the aircraft. Sounding data collected aboard the R/V Vickers at 0600 UTC (three hours prior to sampling by the shipboard radar), indicated the presence of

unidirectional shear with low level westerlies and upper level easterlies, and a CAPE of 2000 J kg^{-1} . The sounding taken for the aircraft case (dropsonde data collected by one of the NOAA P-3 aircraft) had no upper level data but the MLMR was computed to be 18.35 g kg^{-1} . The observed MLMR of 18.35 g kg^{-1} when used with eq. 3.1, yields a CAPE of $\sim 2200 \text{ J kg}^{-1}$, which is slightly larger than that of the ship case. While the CAPE in both soundings was similar, the aircraft sounding indicated higher humidities above the mixed layer to a pressure level of approximately the 700 mb. The deeper layer of low-level moisture in the aircraft sounding is consistent with the presence of deeper moisture convergence southeast of the IFA (e.g., Fig. 3.9b). Low-level (0-1.5 km) shear in the vicinity of both the aircraft and ship cases was rather modest ($\sim 5 \times 10^{-3} \text{ s}^{-1}$), as indicated by the 0600 UTC soundings taken aboard the R/V Vickers, and the PRC-#3 ships (ship closest to the aircraft system; later wind data were contaminated).

4.5.3 The weakly electrified ship case

The ship case (Figs. 4.19 a-c) consisted of a line of convection approximately 150 km long oriented in a NW-SE direction, approximately parallel (within 15°) to the deep tropospheric shear vector. The line of convection propagated to the N-NE at approximately 10 ms^{-1} along an outflow boundary that was initiated by another area of convection situated approximately 100 km to the south of the shipboard radars. The squall-line rapidly deepened as it entered the shipboard dual Doppler lobes near 0900 UTC but remained only weakly electrified. At 1042 UTC (Fig. 4.19a.) the convective line (near its peak in intensity) and gust front reached the MIT radar producing only a modest drop in surface temperature (3°C), wind gusts of $10\text{-}15 \text{ m s}^{-1}$, and moderate foul weather electric fields which peaked at 5 kV m^{-1} . The field mill indicated one possible flash detection at 1052 UTC though no lightning was noted in the R/V Vickers Bridge Weather Log, or detected by the COARE ALDF network. The reflectivities and storm-relative winds shown in Fig. 4.19a were characteristic of the storm. In general, the system was

associated with relatively weak convergence at low-levels that appeared to be situated only near the more intense cells. On the mesoscale, there appeared to be only very weak confluence in the derived low-level wind field.

In the vertical plane, reflectivities (Fig. 4.19b; 1112 UTC) were less than 50 dBZ below the 6 km (-5°C) level and less than 30 dBZ above the 6 km level (consistent with the observed weak electrification), even though the storm exhibited peak radar echo tops approaching 16 km in some cells on the northwestern side of the system. The weak convergence/confluence at low-levels coupled with the presence of weak reflectivity in the mixed phase region of the storm (0° to -40°C), were consistent with weak vertical velocities (e.g., $\leq 6 \text{ ms}^{-1}$; Fig. 4.19c) diagnosed near the 5 - 6 km level, though secondary vertical velocity maxima in excess of $8\text{-}10 \text{ ms}^{-1}$ were occasionally observed at elevations *above* 10 km in the storm (i.e., well above the mixed phase region).

4.5.4 The highly-electrified aircraft case

The aircraft case (c.f., Smull et al., 1994) was an intense, lightning producing squall line located on the southern edge of a large region of primarily stratiform precipitation (e.g., Fig. 4.1). The system propagated toward the NE (in the direction of the decayed system) at a speed of 8 m s^{-1} . The squall line (Figs. 4.20a-b) was approximately 200 km long and was sampled by NOAA P-3 and NCAR Electra aircraft. Similar to the ship case, the aircraft case was oriented approximately parallel to the ambient deep tropospheric shear vector, in a NW-SE direction. Both the eastern (Fig. 4.20a) and western ends (Fig. 20b) of the squall-line were sampled by airborne radar, however the most extensive sampling was performed on the eastern end of the system (Fig. 4.20a), while only two quad-Doppler volumes were collected for the extreme western end of the system (Fig. 4.20b).

region of the storm (Fig. 4.21b), with 30 dBZ cores approaching 15 km in elevation. Similar to the eastern side of the system (e.g. Fig. 4.21a) a distinct local maximum occurred in the reflectivity data (>45 dBZ) near the -10°C level (~ 6.8 km, Fig. 4.21b), *perhaps* indicative of the presence of wet ice or significant concentrations of riming frozen drops and graupel. During the short time that the two volumes of radar data were collected on this side of the system, the center of the CG lightning activity shifted to the northwest and increased in frequency as the convective cores exhibited a slight decrease in intensity aloft.

In cross-sections of vertical velocity for both ends of system (Figs. 4.21c-d), updraft velocities at elevations of 5-7 km equaled or exceeded 6 ms^{-1} near local reflectivity maxima in the most intense cells and in the vicinity of the CG lightning. Note in Fig. 4.21c (1645 UTC) that the accompanying 30 dBZ reflectivity core aloft had descended nearly 2 km from the time of the reflectivity cross-section shown in Fig. 4.21a (1625 UTC), and at 1645 UTC was situated just above the height of the 6 m s^{-1} vertical velocity maximum ($x=12$; Fig. 4.21c) at an elevation of 7 km. Intense convective cells on both ends of the convective line also exhibited elevated vertical velocity maxima (e.g., Fig. 4.21d) of approximately $8\text{-}12\text{ m s}^{-1}$ near the 10 km level.

4.5.5 A kinematic comparison of the lightning and non-lightning producing aircraft and ship cases

Both squall lines occurred in environments of similar CAPE and shear. Yet, the aircraft case exhibited a larger degree of low-level forcing (both in scale and magnitude), substantially more vertical development within the cloud (i.e., considerable mass in the mixed phase region) and a relatively high degree of electrification as evidenced by lightning observations. Convection in the ship case, while producing cloud top heights on the order of 16 km, was associated with only weak surface forcing of limited spatial extent, weak reflectivities above the -5°C level, moderate electric fields at the surface, and

On the eastern end of the convective line (Fig 4.20a), pronounced storm relative confluence occurred on the mesoscale, with flow entering the system from both sides of the convective line. To the rear of the convective line, the flow accelerated to create a small bow in the reflectivity pattern. Smull et al. (1994) determined that the inflow air in this more fully developed portion of the line originated primarily on the northern side of the line (consistent with the direction of propagation). This is an interesting observation as the air entering the system from the north originated in a region of decayed convection and stratiform precipitation, and hence would have been expected a priori to exhibit fairly low moist-static energy.

Cloud-to-ground lightning on the eastern end of the system (all negative in polarity) occurred in areas where the convection was deep and vertically developed (Fig. 4.20a). In these areas radar echo-tops of 16-18 km were common, 30 dBZ reflectivities extended deep into the mixed phase region of the storm, and secondary reflectivity maxima (≥ 40 dBZ) occurred in the 5 to 8 km layer (Fig. 4.21a; 1625 UTC). Interestingly, one intense cell located near $x=33$ in Fig. 4.21a (30 dBZ to nearly 14 km at 1625 UTC) was observed to produce significantly fewer total lightning flashes both visually (as noted by aircraft scientists) and by the ALDF network (an order of magnitude fewer CG's), when compared to convective cells just 30 km to the west that were of similar size and intensity. The majority of the CG flashes seemed to occur between 1625 and 1645 UTC during the descent of 30 dBZ reflectivity cores in convection situated near $x=10$ in Fig. 4.20a.

At the extreme western end of the system the convection and electrification appeared to be more intense (Fig. 4.20b). Enhanced low level convergence in this area was present, as indicated by the nearly head-on confluence present in the horizontal wind field. The pronounced bowed structure in the convective line resembled a line echo wave pattern and was consistent with the pattern of strong storm relative winds that nearly resembled an anticyclonic vortex. The cloud-to-ground lightning (all negative CG's; Fig. 4.20b) in this section of the storm appeared to concentrate in one of the apexes of the bowed reflectivity pattern. Reflectivities in the electrified cells indicated substantial mass in the mixed phase

virtually no detectable lightning. These observations are consistent with the kinematic differences found to exist between the two storms and are most evident when composited fields of radar reflectivity and vertical motion are created.

To maintain consistency in our comparison of storm kinematics, wind-fields for the aircraft case composite were computed using conventional *dual-Doppler* techniques. In this manner artifacts due to different analysis techniques (e.g., quad vs. conventional dual-Doppler) are reduced, making the ship-aircraft case comparison more meaningful. Composited profiles of convective reflectivity, convective updraft and total system vertical velocity are shown in Figs. 4.22a-b. The marked difference in the mean convective reflectivity between the two systems (Fig. 4.22a) is consistent with the differences noted in electrical intensity between the two systems and is also similar to differences noted by Zipser and Lutz (1994) for tropical continental (highly electrified) and monsoon (weakly electrified) mean vertical profiles of radar reflectivity. Furthermore, the difference in reflectivity between the two systems diverges with height, the aircraft case possessing reflectivities ~5 dBZ greater than the ship case at 5 km, and ~10 dBZ greater at a height of 8 km. The increasing difference in reflectivity with height above 5 km in Fig. 4.22a suggests that a significant fraction of the convection sampled in the aircraft case was associated with a larger ice-mass in the mixed-phase region of the cloud.

With regard to the vertical velocity profiles, Fig. 4.22b suggests that larger vertical velocities occurred in the mid and upper troposphere of aircraft case relative to the ship case. In particular, the convective updraft composite of the aircraft case exhibits an increasing vertical velocity profile with height (above 5 km), 50% greater vertical velocities between 5 and 12 km relative to the ship case, and a peak in the convective updrafts occurring near a height of 12 km. The ship convective updraft composite suggests the presence of a bimodal updraft structure, with relative maxima in the updraft situated near 5-6 km and 12.5 km respectively. The aircraft case total vertical velocity profile exhibits a very broad peak in the middle to upper troposphere, whereas the ship case total vertical velocity field retains the

bimodal structure of the convective updraft profile. Importantly, the “mixed” quad/dual analyses (not shown) from the aircraft case (1614 UTC, 1625 UTC- dual; 1645-1705 quad) reveal a vertical velocity profile similar to the dual-Doppler profile shown for the aircraft case in Fig. 4.22b, however the quad-Doppler composite upper level vertical velocities (from 8-12 km) are almost a factor two larger than the dual-Doppler composite (consistent with Jorgensen et al., 1994). Hence the aircraft case dual-Doppler vertical velocity composites shown in Fig. 4.22b are likely a *conservative* estimate of the *true* mean vertical velocities present in the aircraft case.

Observations indicate that the ship and aircraft cases differed markedly in their electrical properties. This is consistent with the differences found to exist in the low level forcing (confluence/convergence), radar reflectivity, vertical motion, and by inference, their respective microphysical properties (e.g., Figs. 4.19, 4.21, 4.22a). Both systems exhibited local reflectivity maxima at lower elevations (i.e., below the freezing level), likely associated with coalescence processes. However, only the electrified system exhibited significant radar reflectivity, and hence significant hydrometeor mass, in the mixed phase region of the clouds. Differences in the mixed phase masses and electrical activity between the two storms, are hypothesized to be the result of the different vertical velocity structures (e.g., Zipser and Lutz, 1994).

For example, the diagnosed updraft magnitudes near the freezing levels (5 km) of the two systems differed by perhaps $2\text{--}3\text{ m s}^{-1}$ in the mean (e.g., Figs. 4.19, 4.20), yet this appeared to be a significant difference from both a microphysical and electrical viewpoint. In the electrified cells of the aircraft case, peak updrafts at the base of the mixed phase region (-5°C to -10°C) appeared to be $\geq 6\text{ m s}^{-1}$. Conversely, vertical velocities at the base of the mixed phase region in the ship case approached but remained generally $\leq 6\text{ m s}^{-1}$. Thus millimeter sized supercooled and frozen raindrops with terminal fall speeds on the order of $4\text{--}6\text{ m s}^{-1}$ could have been lofted higher into the mixed phase region in the aircraft case (and maintained at higher elevations as the particles grew, given general increases in vertical

velocity with height; e.g., Fig. 4.22b). Furthermore, the larger updrafts present in the aircraft case should have generated more supercooled cloud water, creating mixed-phase conditions conducive to the riming growth of graupel and frozen drops, and hence more efficient precipitation based charge generation in the storm (e.g., Takahashi, 1978, Saunders et al., 1991)

Given the generally weaker vertical velocity and radar reflectivity observed in the mixed phase region of the ship case, millimeter-sized liquid and frozen drops were likely confined to temperatures between 0°C and -10°C in the clouds, which is also a favored region for secondary ice-particle production through the Hallett-Mossup rime-splintering mechanism (Hallett and Mossup, 1974). Large concentrations of ice produced in this fashion may have favored a rapid glaciation or “overseeding” of the tropical oceanic convective clouds observed near the -5°C level in the ship case (e.g., Houze and Churchill, 1984; Gamache, 1990). The presence of weak updrafts, large concentrations of smaller ice particles (providing an efficient means for removal of cloud water), coupled with the fallout of frozen raindrops and graupel, *may* explain the relative decrease in radar reflectivities observed near the -5°C level in the ship case and the absence of lightning in the storm. Considering the difference in vertical velocities between the two systems near the 6 km level and the terminal fall speeds of millimeter diameter liquid/frozen raindrops, it may be that 6 m s⁻¹ updrafts are close to the “*threshold*” value required at this level for significant electrification of tropical oceanic convection to occur. In this respect, the ship case may have been on the *verge* of electrifying (which is consistent with the occasional extension of 30 dBZ reflectivities to 6.5 km, but not above). This “*threshold*” vertical velocity is similar to that suggested for oceanic thunderstorm electrification by Zipser (1994).

The deep convective cores in the aircraft case suggest the presence of moderate concentrations of graupel and/or frozen drops, and the local maxima in reflectivity observed between 6 and 7 km suggests the presence of larger ice particles and/or higher particle concentrations. It is interesting to compare the overall reflectivity structure (and reflectivity maxima at 5-7 km) in the aircraft case to multi-parameter

radar and microphysical observations collected in Florida by Bringi et al., (1996; B96), Jameson et al. (1996; J96) and Ramachandran et al. (1996; R96). The electrified storms studied in Florida by B96, J96, and R96 exhibited reflectivity structures similar to the aircraft case though the core reflectivities were typically larger. Multiparameter radar data and microphysical data the B96, J96 and R96 studies indicated that the reflectivity cores were collocated with positive differential reflectivity (Zdr) columns (associated with rain drops), that extended to nearly 6 km in depth. The top of the Zdr columns (near local reflectivity maxima) were collocated with LDR (linear depolarization ratio) caps. In turn, LDR caps in convection have been attributed to the presence of wet tumbling or irregularly shaped ice (Herzogh and Jameson, 1992), and/or oscillating partially frozen raindrops (e.g., Smith et al., 1995; B96, J96, R96). Hence, the reflectivity and vertical velocity structures in the aircraft case (to the extent that the reflectivity structures are comparable to those of the Florida studies) *may* suggest the presence of raindrop freezing and ice particle riming near the height of the -10°C level (~6.8 km). In turn, the presence of mixed phase microphysics near and above the -10°C level should be conducive to efficient non-inductive charging of the clouds (e.g., Takahashi, 1978; Saunders et al., 1991).

4.6 Vertical velocity observations in tropical oceanic convection: Evidence for a threshold

To further examine the idea that a “*threshold*” in vertical velocity near the 0°C to -10°C levels must be exceeded prior to the occurrence of significant electrification of tropical oceanic cumulonimbi (similar to Zipser, 1994; Zipser and Lutz, 1994), approximately 30 different dual-Doppler observations (over different cases, times and convective cells) of vertical velocity and 30 dBZ echo heights were composited from TOGA COARE radar data (ship and aircraft). It was then assumed that 30 dBZ echo tops situated in the mixed phase region of *growing or mature* clouds represented a “balance level” (i.e., where updraft velocity equals the particle terminal fall speed; Atlas, 1966; Lhermitte and Williams,

1985). In this framework, vertical velocities sampled at the 30 dBZ contour level are indicative of a mean terminal fall speed for precipitation in the analysis volume (graupel, liquid/frozen drops etc.). Fall speeds for different sized raindrops and frozen drops/hail (0.8 mm, 1 mm, and 2 mm diameters) were computed using $V_t = aD^b$ fall speed relationships (D = diameter). The fall-speed relationship for rain was taken from Ferrier (1994), and the relationship used for frozen drops/hail was based on the observations of Lozowski and Beatty (1979) and theoretical calculations by Bohm (1989). Due to a combination of sparse measurements, and extrapolated results that were originally intended to explain "hail" fallspeeds, (particle diameters ≥ 5 mm), some degree of uncertainty (indicated by vertically oriented arrows in Fig. 4.23) exists in the computed terminal fall speeds associated with frozen drops ≤ 4 mm in diameter. The rain/frozen drop fallspeeds were then plotted as a function of height together with the observed "mean" and "maximum" balance vertical velocities observed in each cell at the top of 30 dBZ reflectivity cores (Fig. 4.23).

The results (Fig. 4.23) suggest that the *lightning-producing* oceanic cells (those cells with 30 dBZ echo located at temperatures colder than -10°C), had mean/maximum vertical velocities that would support the lofting of liquid or frozen drops with a diameter of ~ 1 mm above the height of the -10°C level ($4.5 < w < 6 \text{ m s}^{-1}$). Though not shown in Fig. 4.23, vertical velocities of $4.5\text{--}6 \text{ m s}^{-1}$ would also be able to support graupel particles (that might have originated with millimeter sized frozen drops) of approximately 5 mm in diameter due to their reduced fall speeds relative to that of frozen drops. Now recall that millimeter sized drops (supported by vertical velocities between 4.5 and 6 m s^{-1}) in a monodisperse distribution of concentration 1 L^{-1} , will yield a reflectivity of 30 dBZ. Further, a concentration $O(1 \text{ L}^{-1})$ of mm sized ice particles (e.g., frozen drops and/or graupel) is thought to be a necessary condition for significant non-inductive charging to occur (Takahashi, 1984). It is important to realize that the vertical velocities used in Fig. 4.23 were diagnosed from dual-Doppler syntheses and are likely a *conservative* estimate of the true updraft strength in the core (e.g., quad-Doppler results

suggested 50% greater updraft magnitudes in some instances). It seems possible that maximum updrafts may have been large enough to loft 2 mm diameter drops into the mixed phase region of the cloud (Fig. 4.23). Note that the uncertainty in the 2 mm diameter frozen drop fall speeds, which decreases as the particle roughness or oblateness increases (e.g., note the possibility of lower V_t 's for 2 mm drops indicated by the arrow), might also permit the updraft to levitate those particles once frozen (especially if they were slightly rimed).

Observations of convection in Florida summarized by Jameson et al. (1996) and Ramachandran et al. (1996) suggest that the lofting of mm sized drops above the -10°C level, followed by drop freezing and the onset of glaciation, occurs coincident with rapid electrification of the clouds. To the extent that the "balance-level" assumptions (e.g., Atlas, 1966; Lhermitte and Williams, 1985), observations, and fall speed calculations made herein are valid and representative respectfully, the observations provide some limited circumstantial evidence to support the notion of a "*vertical velocity threshold*" of $\sim 6 \text{ m s}^{-1}$ (linked to the fall speed of mm sized rain/frozen drops) that must be exceeded in order for significant electrification to take place in an oceanic cumulonimbi (e.g., Zipser and Lutz, 1994). Interestingly, this vertical velocity is also approximately representative of the upper 10% of vertical velocities observed in tropical oceanic convection (e.g., LeMone and Zipser, 1980; Jorgensen and LeMone, 1989).

4.7 Summary

Similar to observations of electrified convection in mid-latitudes (Sec. 4.3 of this Chapter; c.f., Williams and Lhermitte, 1983; Lhermitte and Williams, 1985; Dye et al., 1986; Goodman et al., 1988; Williams 1989; Williams et al., 1989; Dye et al., 1989; Marsh and Marshall, 1993; Carey and Rutledge, 1996) and over tropical continents (e.g., Chauzy et al., 1985; Rutledge et al., 1992; Williams et al., 1992; Zipser and Lutz, 1994), observations presented in Secs. 4.2-4.4 suggest that lightning producing tropical

oceanic convection is associated with the presence of significant hydrometeor mass in the mixed phase region of the clouds (as indicated by the presence of reflectivities ≥ 30 dBZ; e.g., Table 4.1).

Based on radar observations (e.g., Figs. 4.5-4.10), it appears that an increase in the mixed phase ice-mass in electrified oceanic convection occurs in a multi-stage process. First, updrafts above a certain threshold intensity ($\sim 6 \text{ m s}^{-1}$; Fig. 4.23) loft millimeter sized raindrops into regions of the cloud with temperatures colder than -5°C to -10°C . This stage is followed by the freezing of the drops, which is likely accompanied by the rapid production of ice particles (e.g., Koenig, 1963) due to primary (e.g., Meyers et al., 1992) and secondary (e.g., rime-splintering, contact nucleation, coalescence freezing etc.; see Young, 1993) ice nucleation processes. Increases in ice-water content in the presence of supercooled liquid water, produce a rapid electrification of the cloud (e.g., B96, J96 and R96) due to efficient non-inductive charging processes (e.g., Takahashi, 1978; Saunders et al., 1991). Observations in at least two cases over the warm-pool (31 January and 1 February 1993; Fig. 4.6), coupled with observations collected in the mid-latitudes (e.g., Goodman et al., 1988; Williams et al., 1989; Carey and Rutledge, 1996) suggest that in-cloud flashes are initiated during the rapid growth phase of the convection (i.e., as the 30 dBZ contour extends rapidly into the mixed phase region; consistent with Murphy et al. 1996). Following the initiation of IC flashes, as the precipitation ice core (≥ 30 dBZ reflectivities) peaks in its vertical growth and begins to descend, gravitational power is converted to electrical power in the lower mixed-phase region of the cloud (e.g., Workman and Reynolds, 1949; Williams and Lhermitte, 1983), and cloud-to-ground lightning begins to increase (i.e., an electrical "bias" for CG lightning is created; e.g., Fig. 4.10).

Over the tropical ocean, deep convective clouds possessing updrafts less than $5\text{-}6 \text{ m s}^{-1}$ (e.g., Fig. 4.20) in levels of the cloud warmer than -10°C (true of $\sim 90\%$ of tropical oceanic convection; LeMone and Zipser, 1980; Jorgensen and LeMone, 1989; Lucas et al., 1994) are not capable of lofting the required rain and liquid water mass into the mixed phase region of the cloud. Therefore 30 dBZ echoes

rarely appear in these clouds (e.g., Figs., 4.11-4.13) at temperatures colder than -10°C . Note that radar echo statistics collected during COARE (DeMott, 1996) suggest that ~90% of the total “convective-echo” area is associated with 30 dBZ reflectivity features that occur at temperatures warmer than -10°C (which is consistent with the vertical velocity threshold hypothesis; e.g., Fig. 4.23). Regardless of the depressed 30 dBZ height, once the liquid and frozen hydrometeor mass is unloaded from the updraft near the 0°C level and the latent heat of fusion is released during the freezing process, the clouds are still capable of reaching great depths in the troposphere while generating copious amounts of cloud ice and snow aloft (e.g., the 9, 19, 23 February 1993 cases). Hence tropical oceanic clouds are often very deep, but only weakly electrified.

Based on the “convective” or “cloud-scale” observations presented in this chapter, and the discussion of CAPE, buoyancy and lightning in Chapter 3, specific hypotheses/questions can now be tested in modeling experiments to be discussed in Chapter 5. The questions to be addressed in the modeling experiments include the following:

- 1) To what extent does the profile of thermal buoyancy control the electrification of tropical convective clouds and how does this differ from a mid-latitude environment.
- 2) With regard to cloud microphysics and electrification, what is the role of frozen drops (vs. lower density graupel) in the electrification process?
- 3) Given the anomalous vertical structure of lightning producing tropical oceanic convection (e.g., 30 dBZ echo's which extend well above the height of the -10°C level), do lightning producing oceanic clouds produce a distinct heating and/or moistening profile and if so, how does this profile compare to electrified convection in other regions of the tropics.
- 4) Does the precipitation process (e.g., warm-rain coalescence vs. ice-processes) change appreciably in lightning-producing oceanic convection relative to that observed in deep, but non-lightning producing convection.?

GMS-4 satellite data for 1993.02.09 16:45

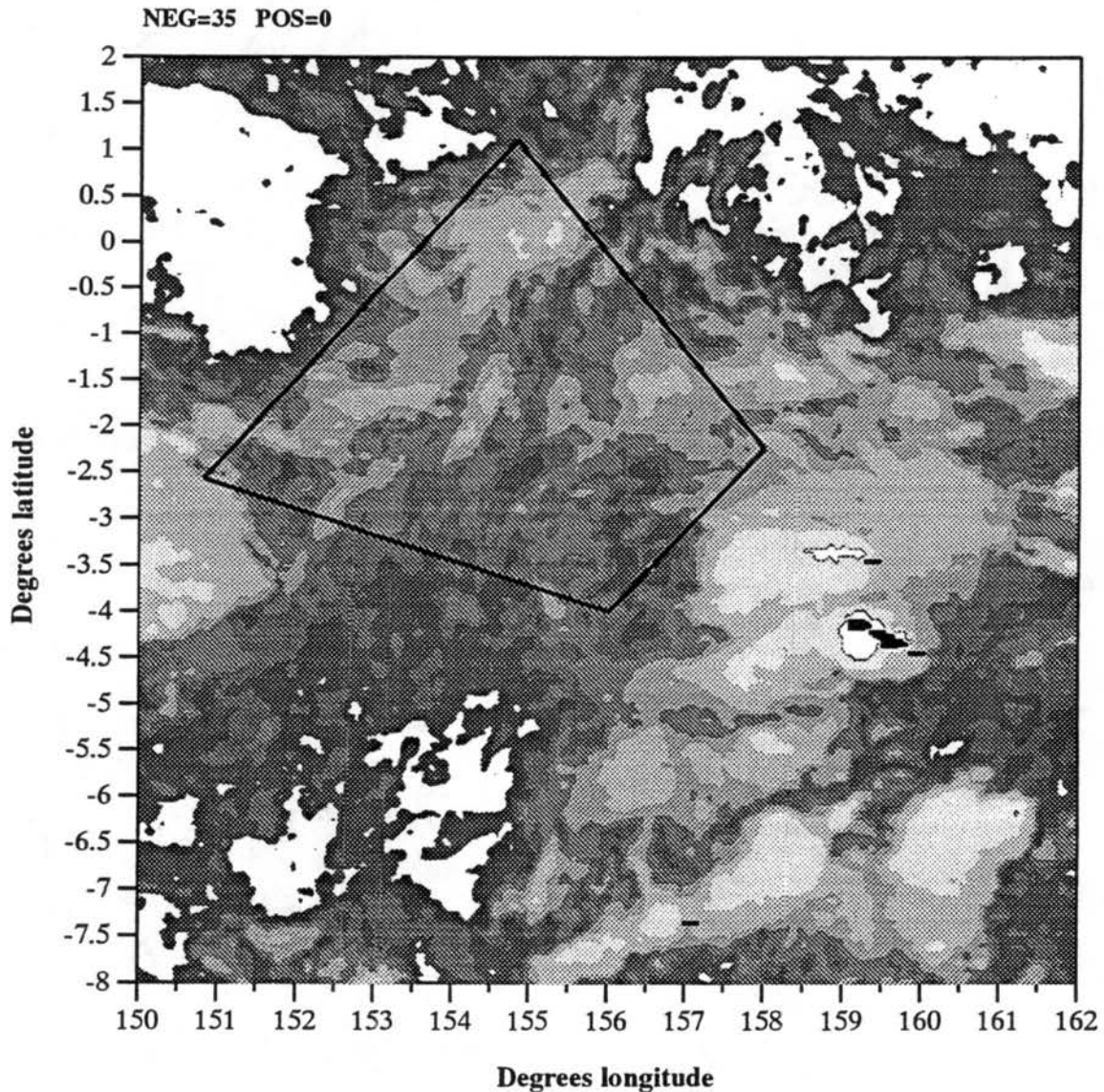


Figure 4.1 GMS IR satellite imagery and CG flashes for 1645 UTC 9 February 1993. Shading intervals from dark to light indicate temperatures less than 15°C, 0°C, -15°C, -30°C, -45°C, -65°C, and -80°C. The COARE IFA is enclosed by the dark solid lines. CG flashes of negative [positive] polarity are indicated by a (-) [+] and are plotted for a 30 minute interval centered on 1645 UTC. The number of negative and positive CG's that occurred in the time period is indicated in the upper left hand corner of the figure.

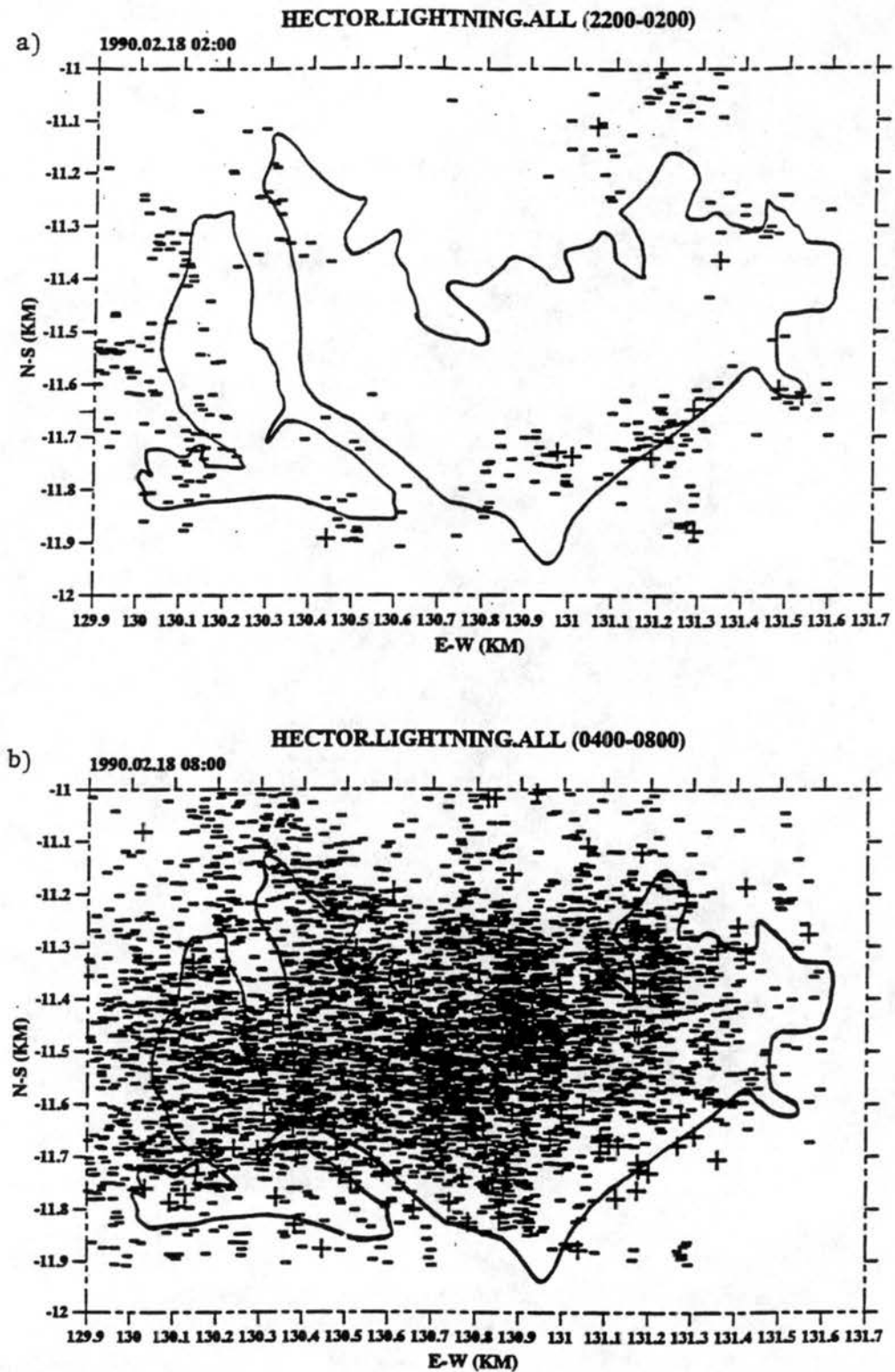


Figure 4.2 Cloud-to-ground lightning flashes detected over the islands of Melville and Bathurst from Dec. 1989 to Feb. 1990, for times between: a) 2200 - 0200 UTC (0730 - 1130 L) and b) 0400-0800 UTC (1330 - 1730 L). Negative [positive] flashes indicated by a (-) [+].

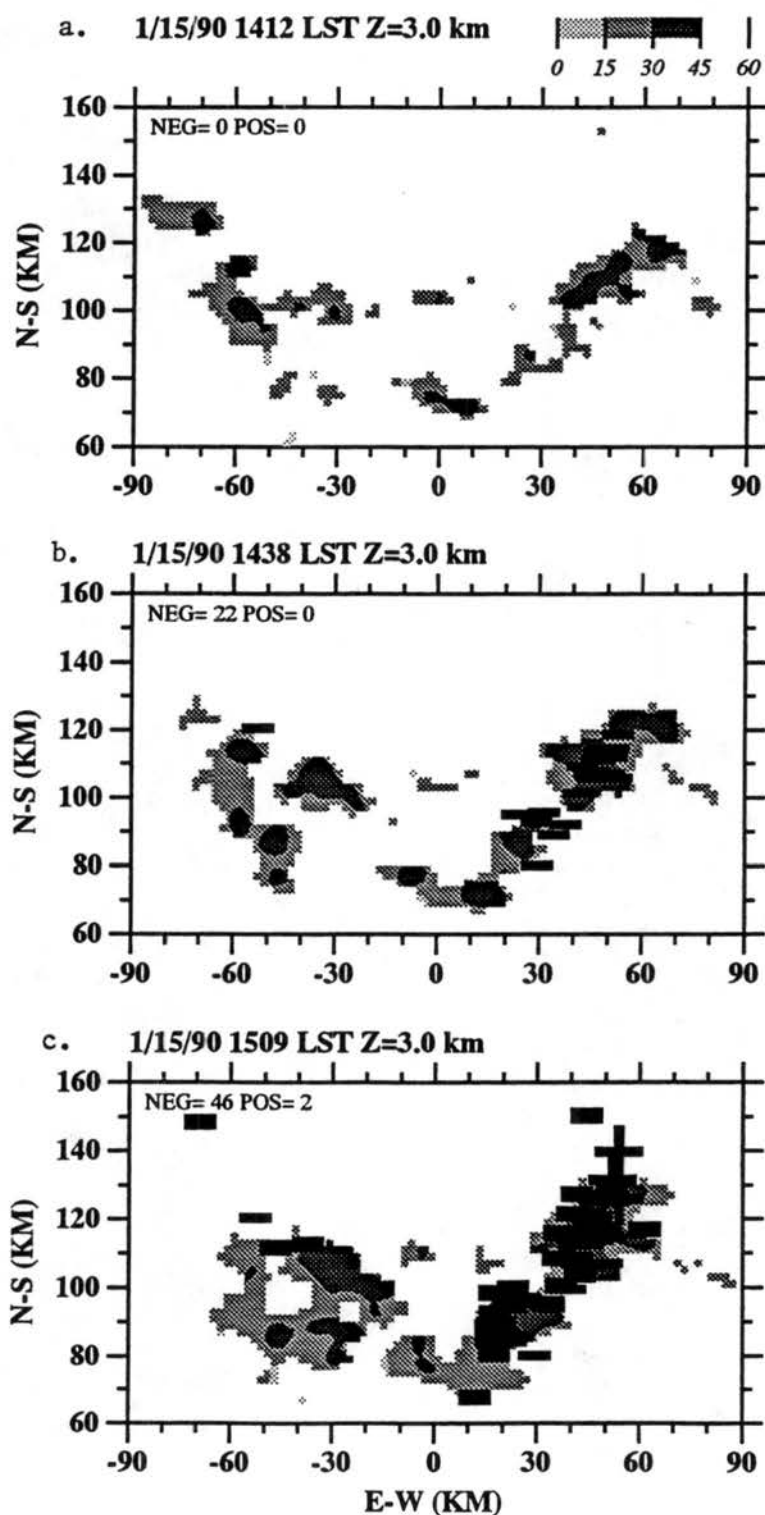


Figure 4.3 Horizontal cross-sections of radar reflectivity ($Z = 3$ km; 15 Jan. 1990) and CG lightning locations for: a) 1412 L; b) 1438 L; and c) 1509 L.

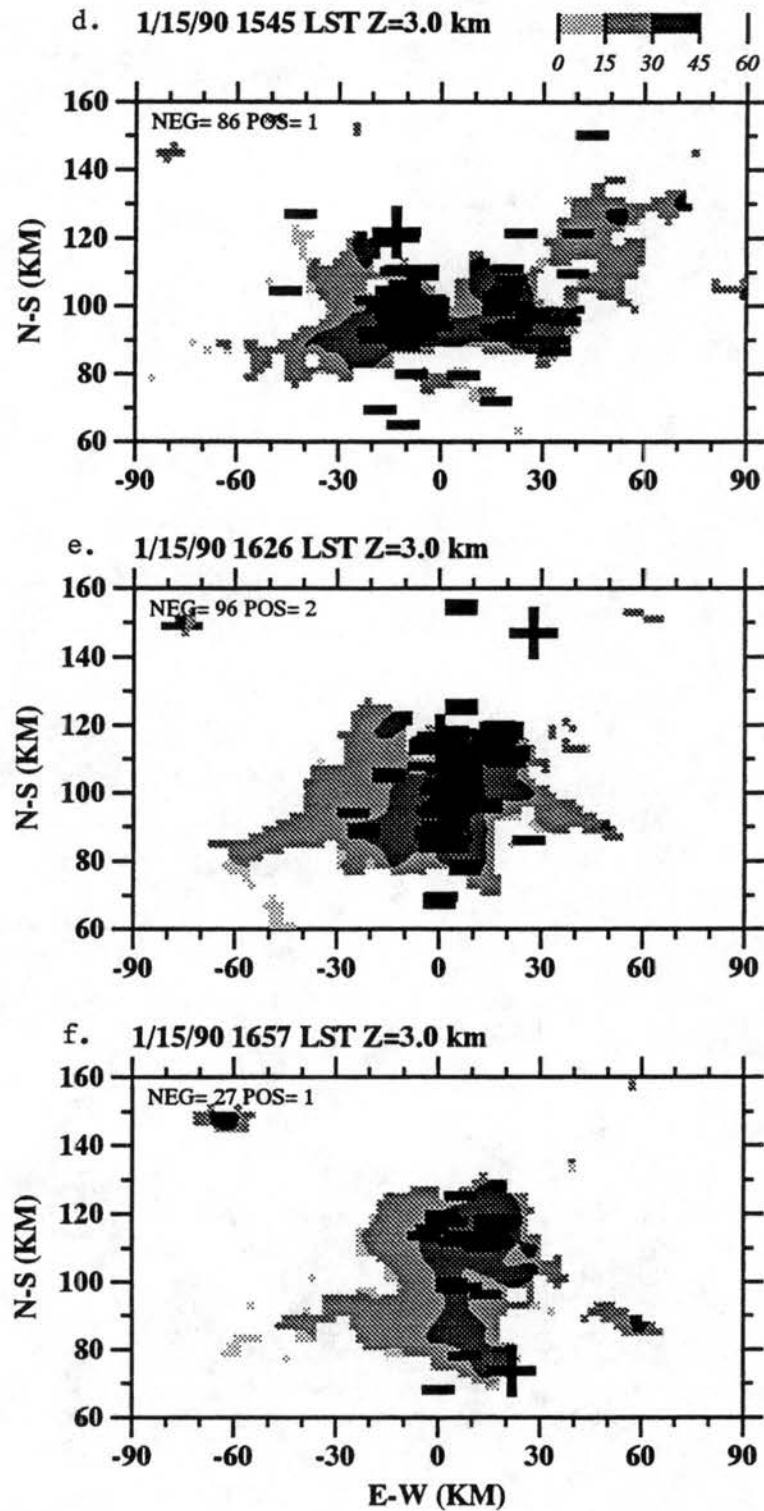


Figure 4.3 As in Fig. 4.3a but d) 1545 L; e) 1626 L; and f) 1657 L.

**30 dBZ Volume/top, 1 dBZ Echo-top, and CG Flashes
Tiwi Islands: 15 January 1990**

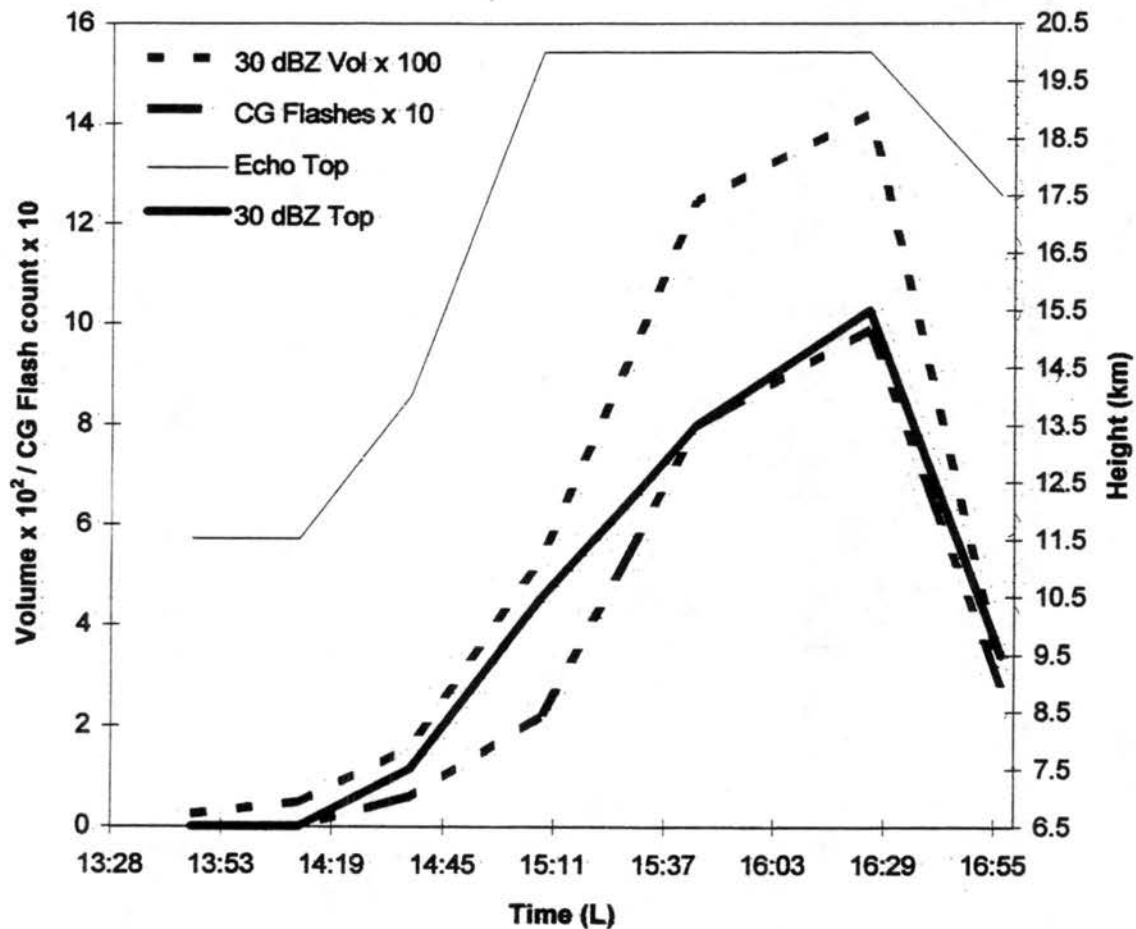
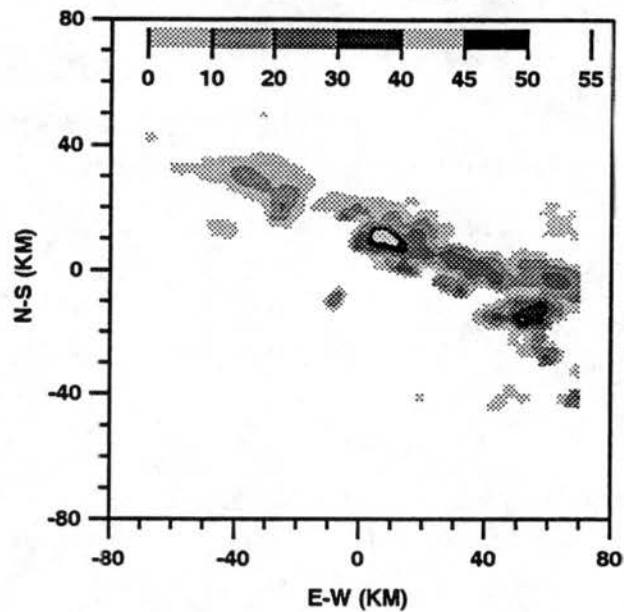
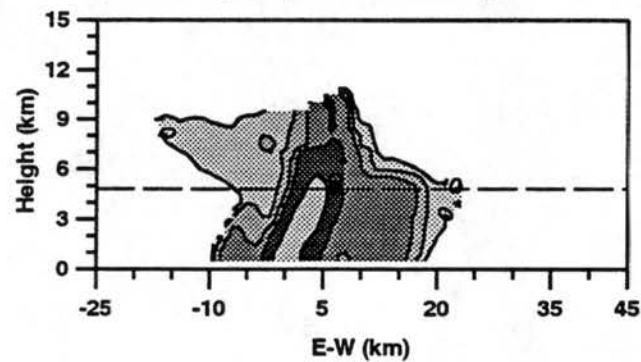


Figure 4.4 Time-height series of 30 dBZ volume ($\times 100 \text{ km}^3$), echo-top (km), 30 dBZ contour top (km) and the number of CG flashes observed in each time interval indicated on the abscissa for 15 Jan. 1990 (1328-1655 L). Flashes and reflectivity tops/volume are indicated on the left ordinate, height on the right.

a. MIT 31 JAN 93 2342 UTC (dBZ) Z= 2 Km



b. 2332 UTC (dBZ) XSCT: 90 Deg. Y= 11 km



c. 2351 UTC (dBZ) XSCT: 90 Deg. Y= 11 km

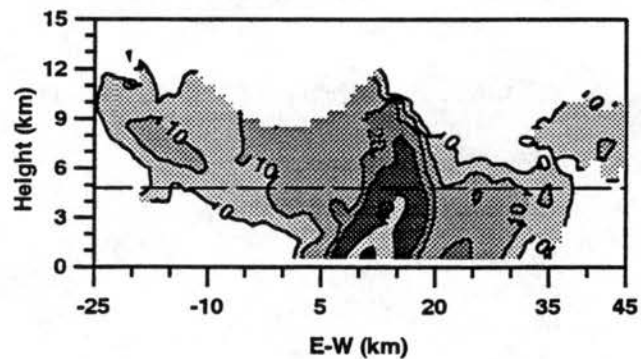


Figure 4.5 MIT Radar 31 Jan. 93, a) 2342 UTC, horizontal cross-section of radar reflectivity at $Z = 2$ km, radar located at (0,0); b) 2332 UTC, E-W vertical cross-section of radar reflectivity taken through the convective line shown in (a), reflectivity values shaded as in (a) approximate height of the 0°C isotherm indicated by dashed line; c) same as (b) but 2351 UTC.

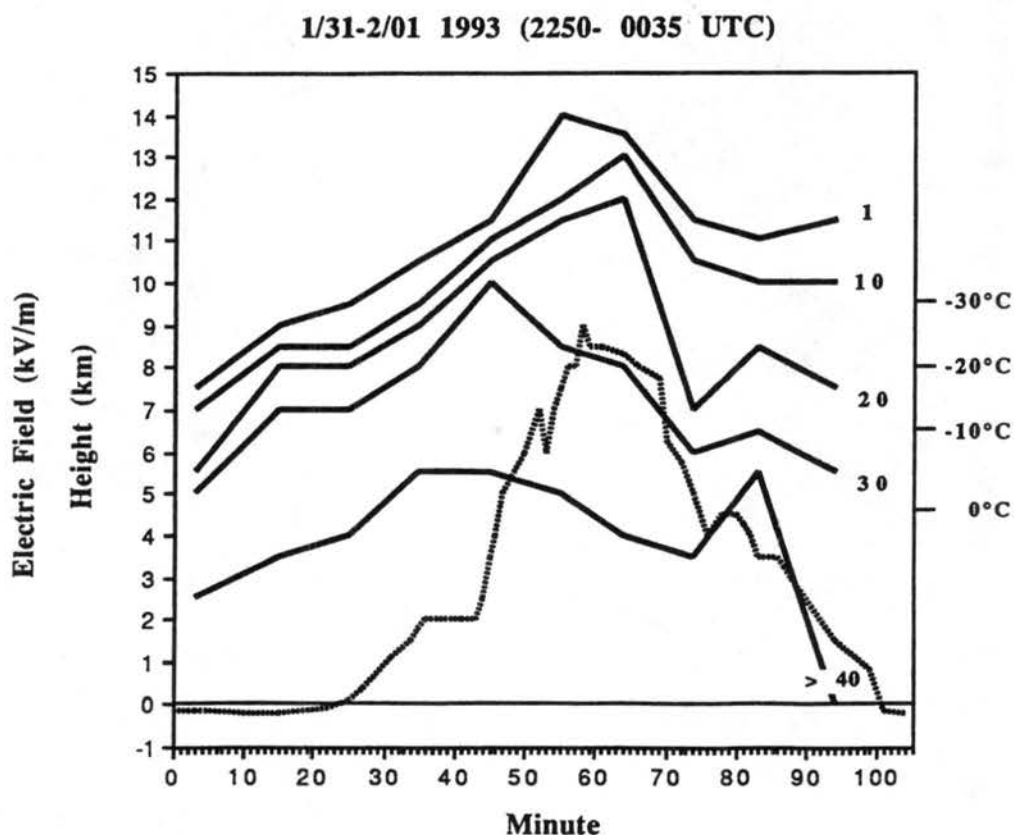


Figure 4.6 MIT Radar, R/V Vickers electric field mill, 31 Jan. 1993 (2250 UTC) to 1 Feb. 1993 (0035 UTC). Time series of electric field (kV m^{-1} ; 1 min intervals) plotted with a time-height series (10 min. intervals) of radar reflectivity echo-top (km) for values of 1, 10, 20, 30 and ≥ 40 dBZ. Approximate heights for the 0°C , -10°C , -20°C and -30°C isotherms are indicated on right ordinate.

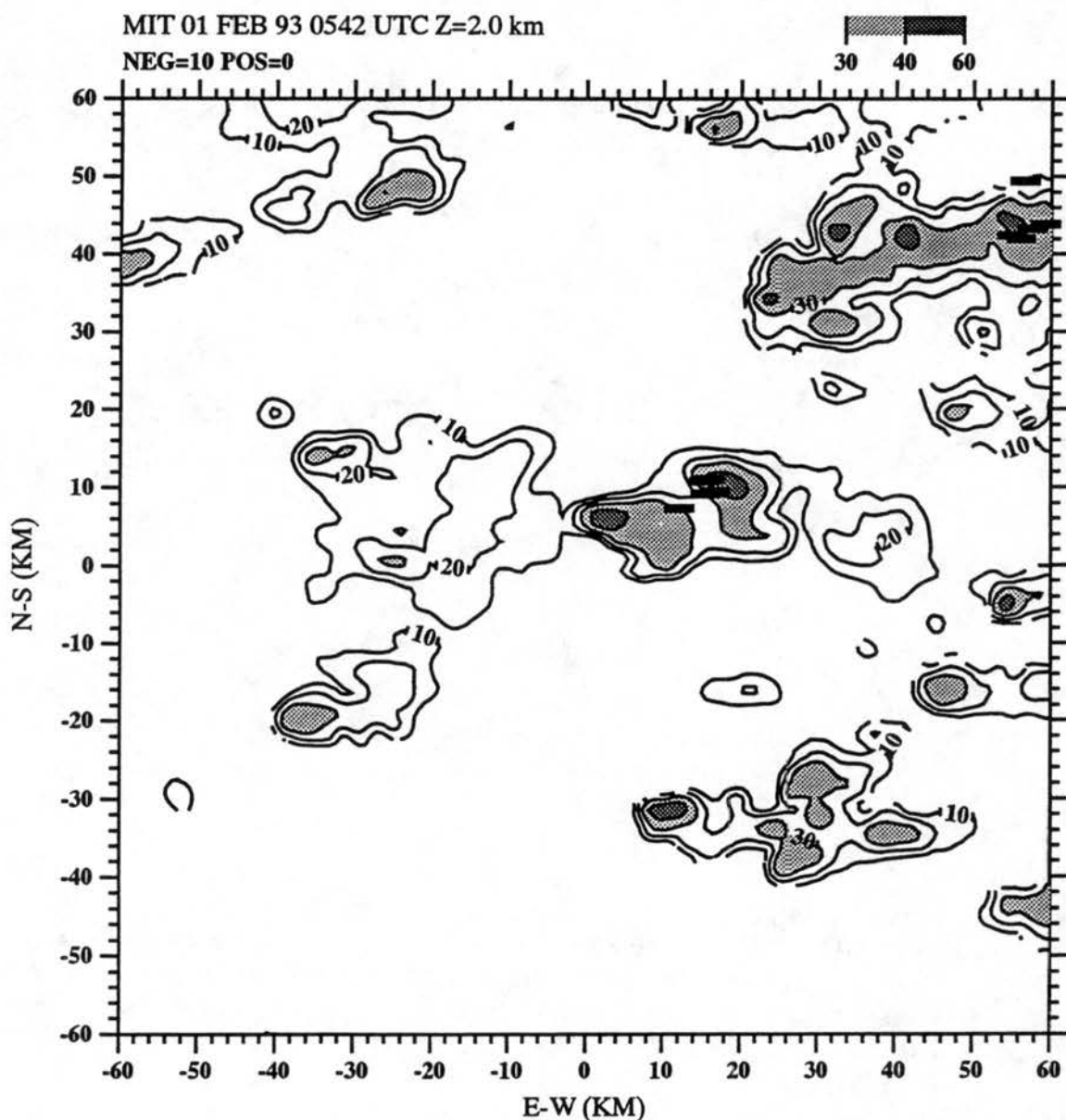


Figure 4.7 MIT Radar 01 Feb. 1993 (0542 UTC), horizontal cross-section of radar reflectivity at $Z = 2$ km, and CG return strokes (all negative in polarity) for a 20 minute period centered on the time of the cross-section. The radar is located at (0,0). Reflectivities are contoured every 10 dBZ, and shaded for values ≥ 30 dBZ.

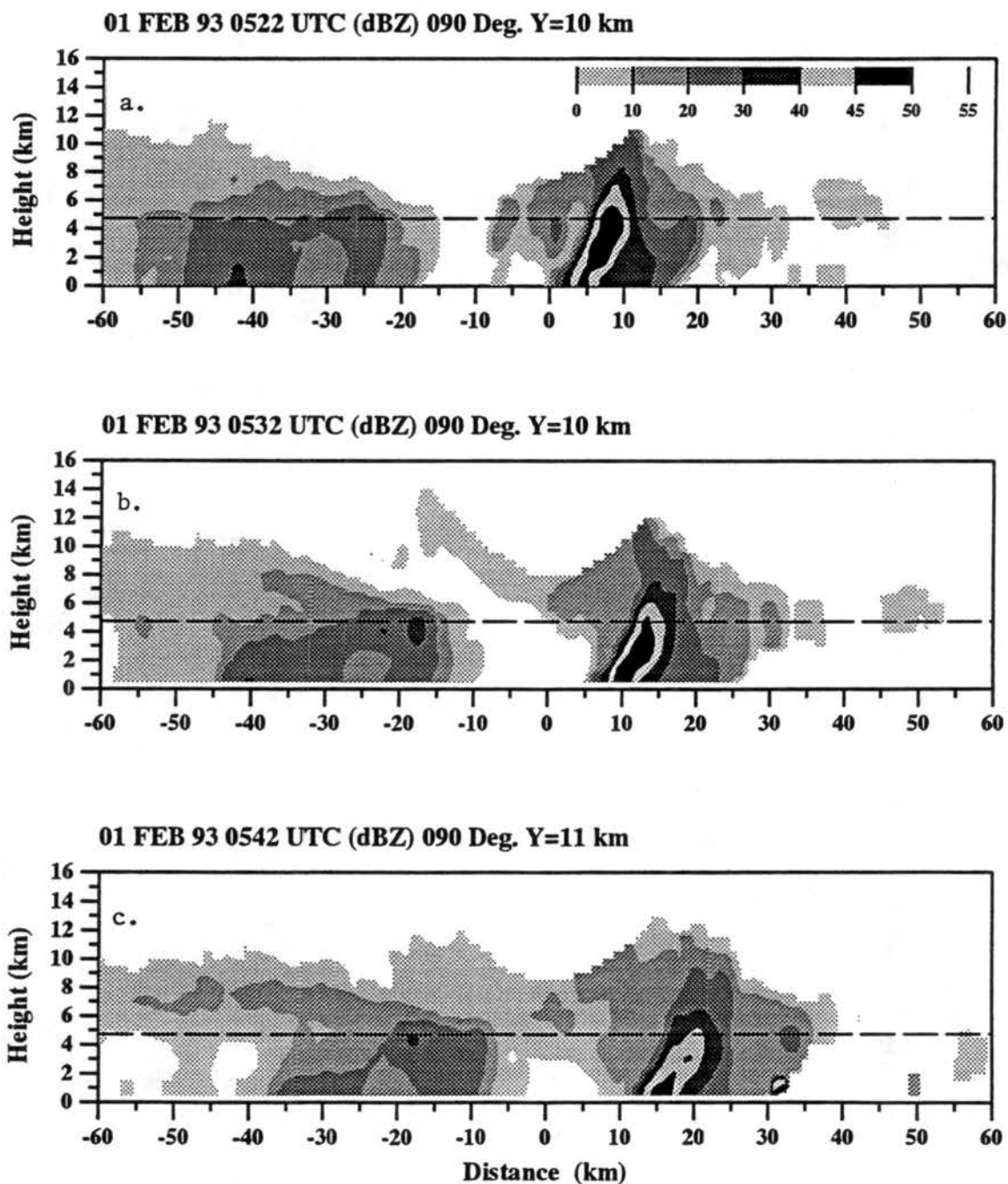


Figure 4.8 MIT Radar 01 Feb. 1993. Vertical cross-sections of MIT radar reflectivity parallel to the X-axis in Fig. 4.7 at: a) 0522 UTC, Y = 10 km; b) 0532 UTC, Y = 10; c) 0542 UTC, Y = 11.

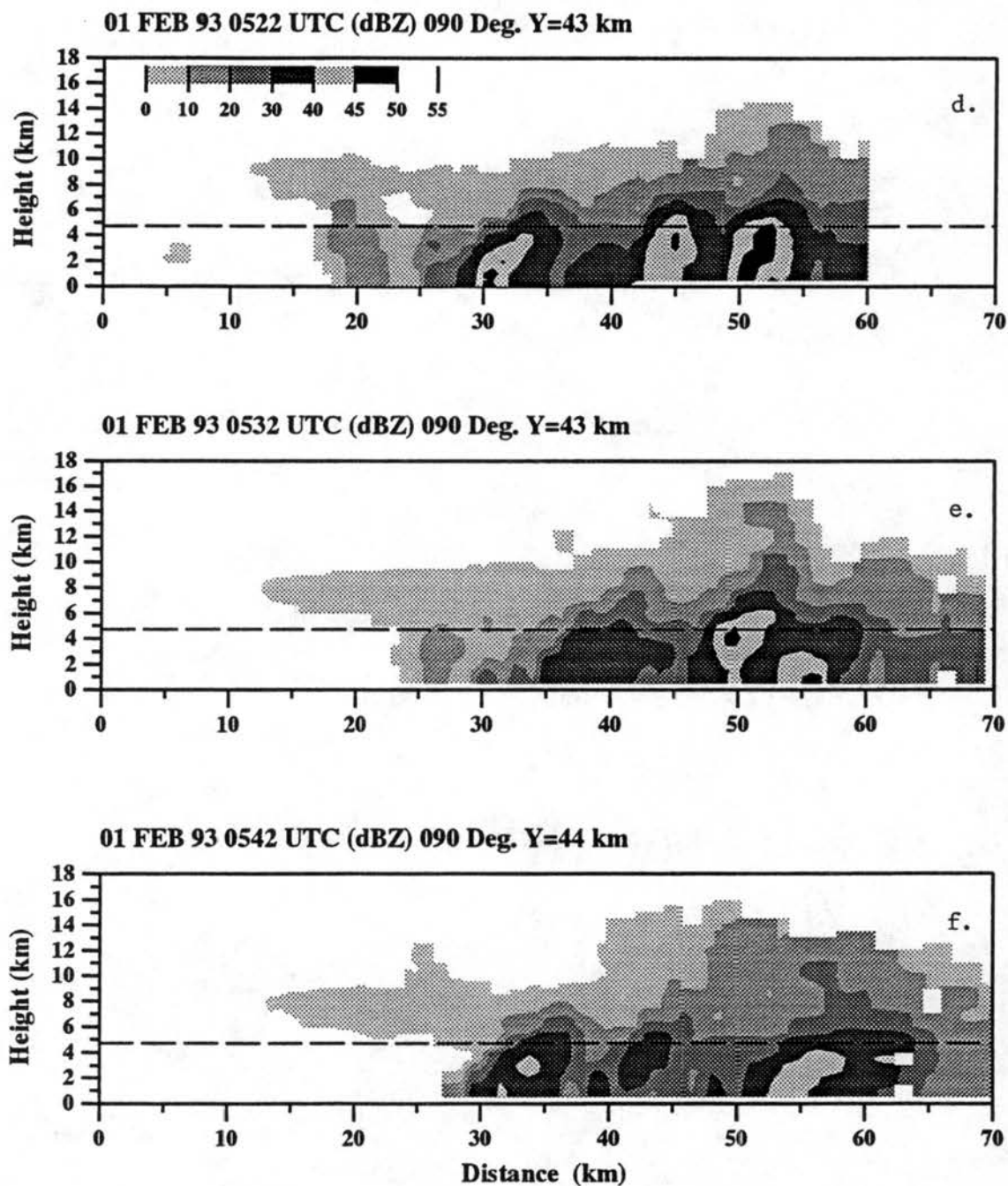


Figure 4.8 As in Fig. 4.8a but d) 0522 UTC, Y = 10; e) 0532 UTC, Y = 43 km; f) 0542 UTC, Y = 44 km.

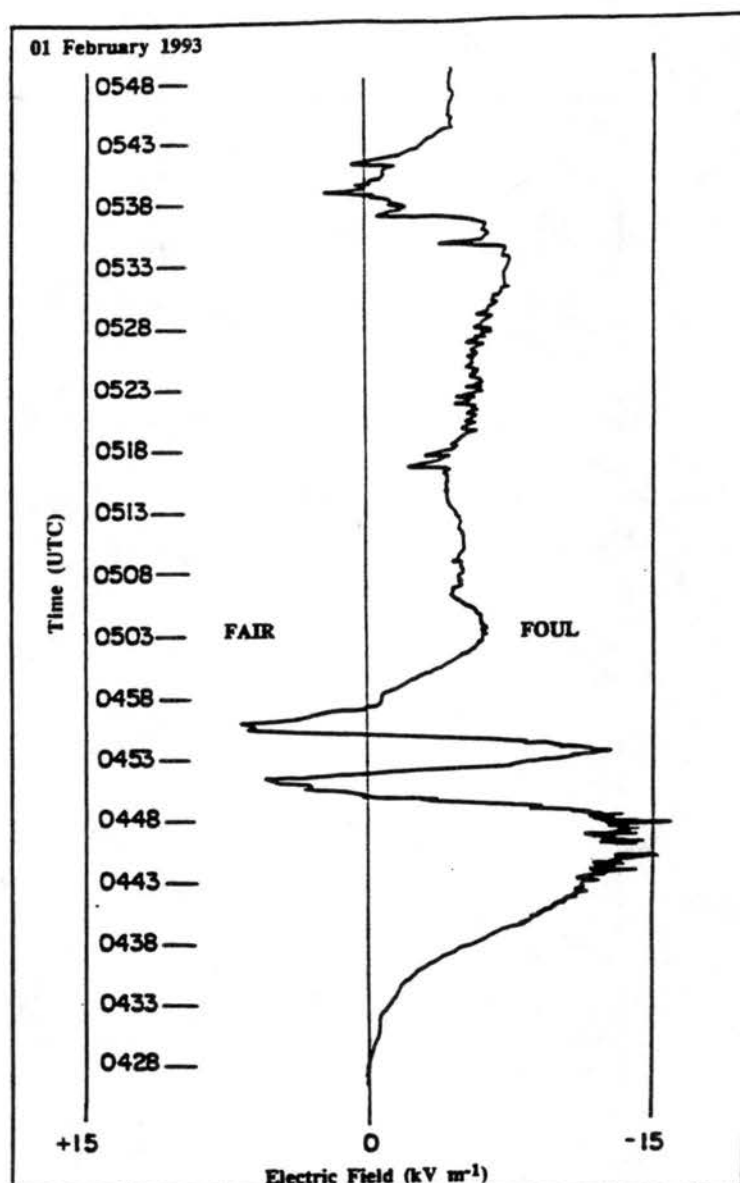


Figure 4.9 Electric field data (low-gain) for 1 February 1993, 0428 UTC - 0548 UTC. The magnitude of the electric field (kV m^{-1}) is plotted on the abscissa and time (UTC) is plotted on the ordinate.

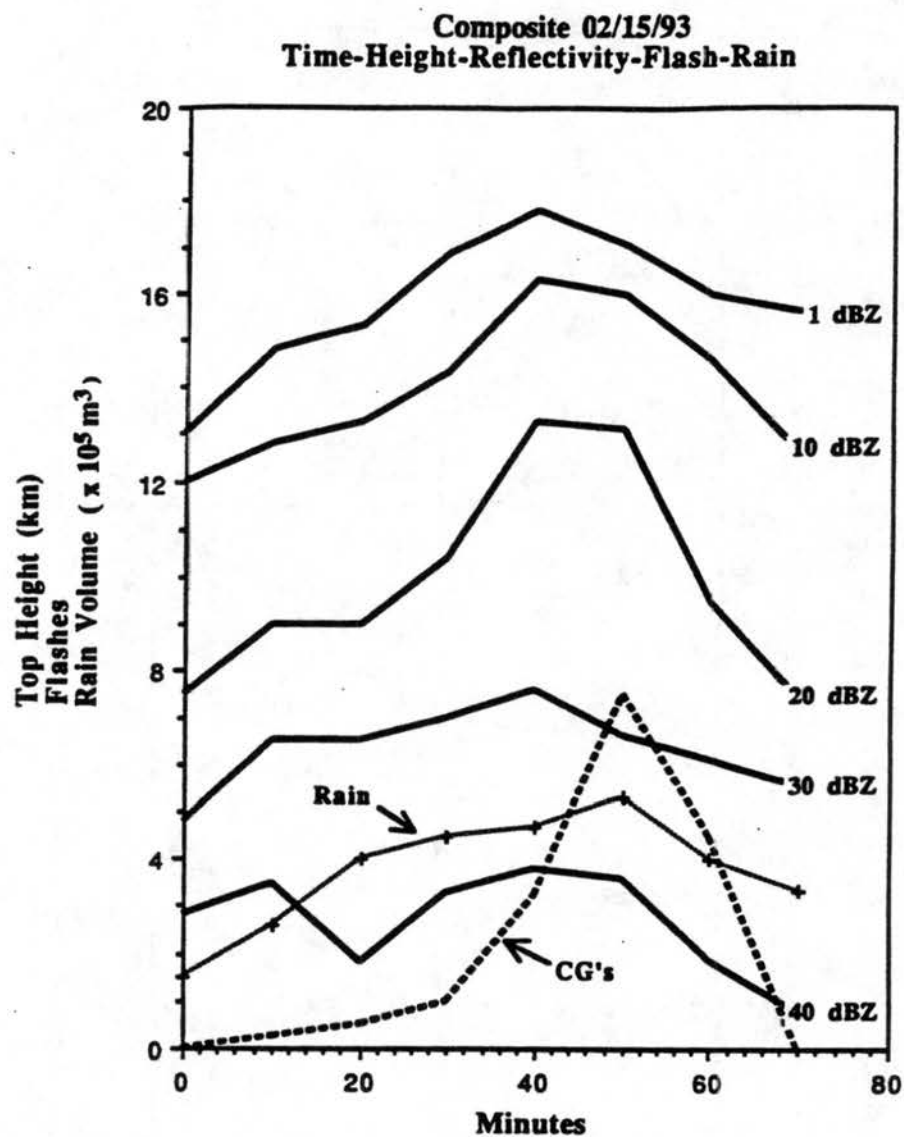


Figure 4.10 Composited time-height, reflectivity (dBZ), rain and lightning frequency for 15 February 1993. Values of reflectivity echo-top height (km; solid lines), CG flash number (dashed line), and rain volume ($\times 10^5 \text{ m}^3$, dotted line with (+) markers).

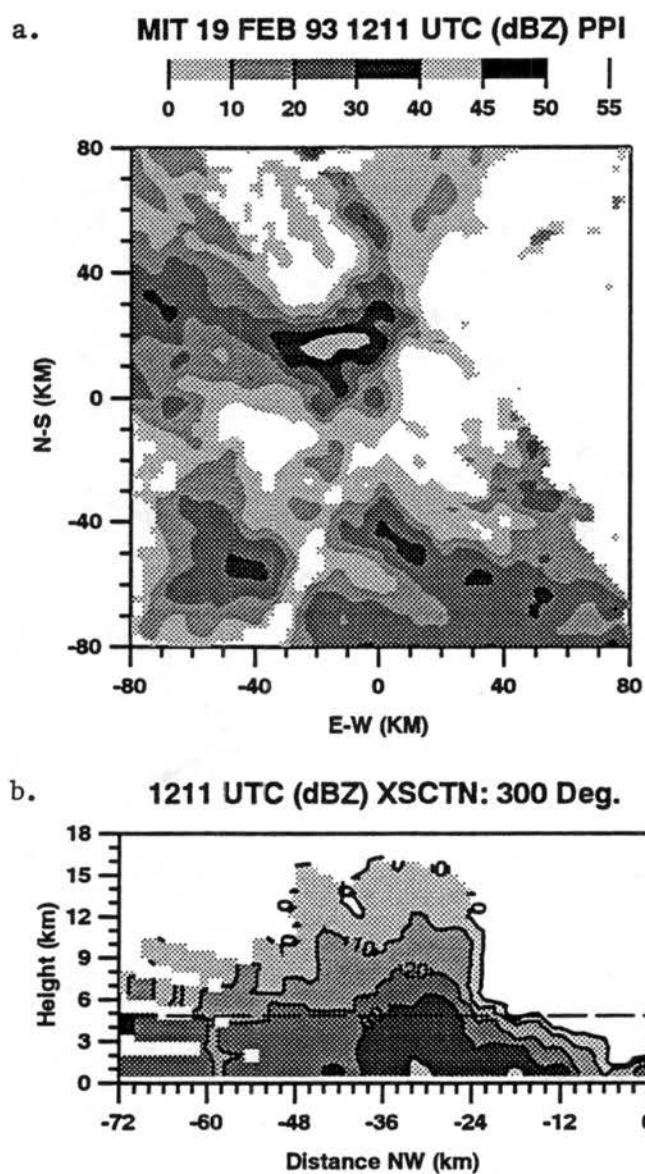


Figure 4.11 a) as in Fig. 4.5a, but 1211 UTC 19 Feb. 1993 b) as in Figs. 4.5b-c but 1211 UTC 19 Feb. 1993, azimuth of 300°.

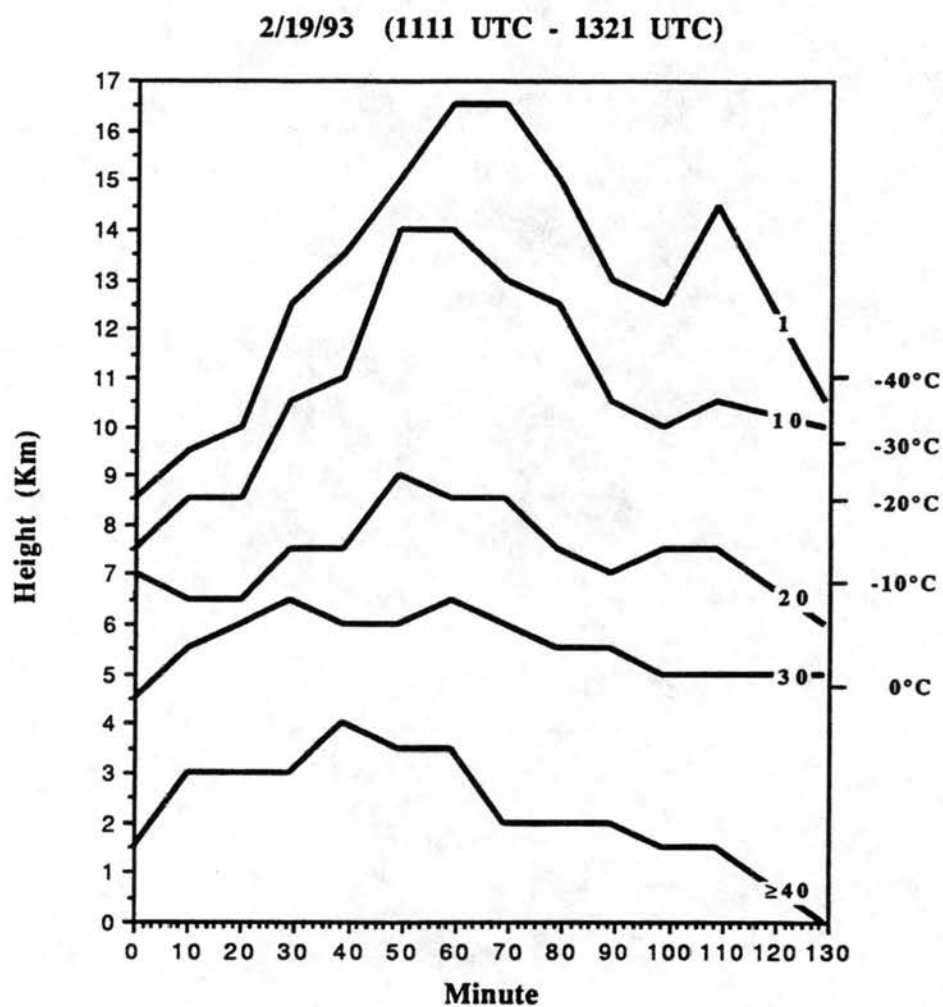


Figure 4.12 19 Feb. 1993, (1111-1321 UTC) time-height (km; left ordinate) series of radar reflectivity for echo-tops of 1, 10, 20, 30, and ≥ 40 dBZ. Approximate heights of the 0°C, -10°C, -20°C, -30°C and -40°C isotherms are indicated on the right ordinate.

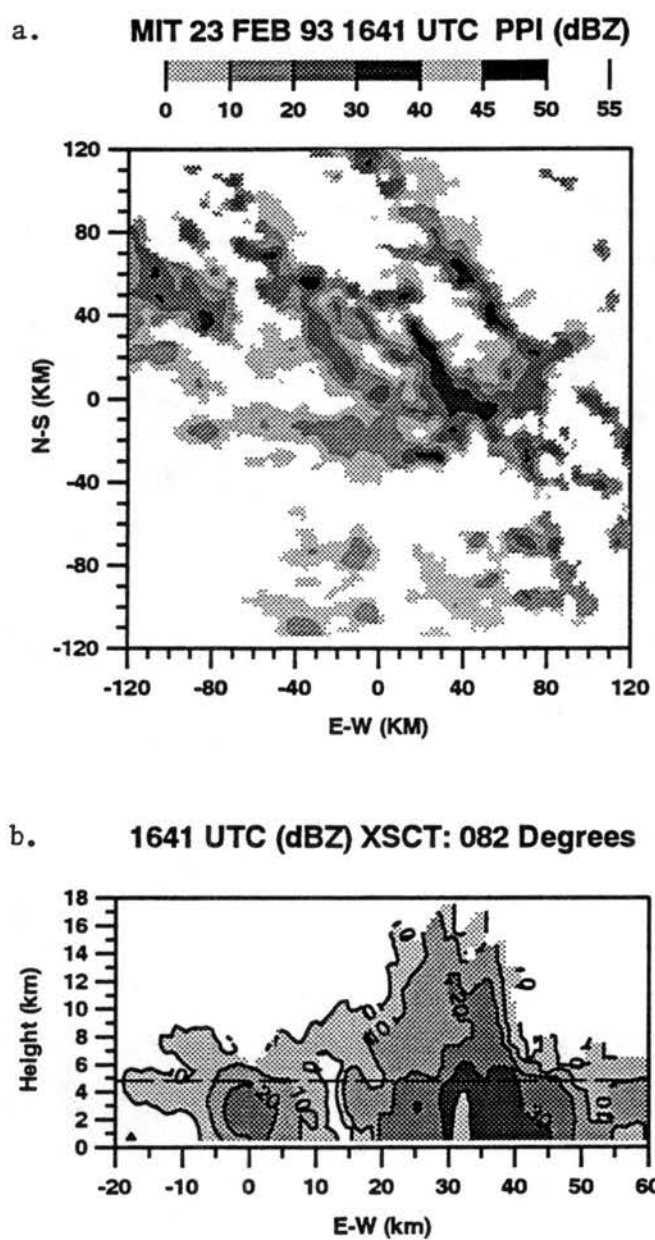


Figure 4.13 a) As in Fig. 4.5a, but 1641 UTC 23 Feb. 1993; b) as in Figs. 4.5b-c but 1641 UTC 23 Feb. 1993, azimuth 82°.

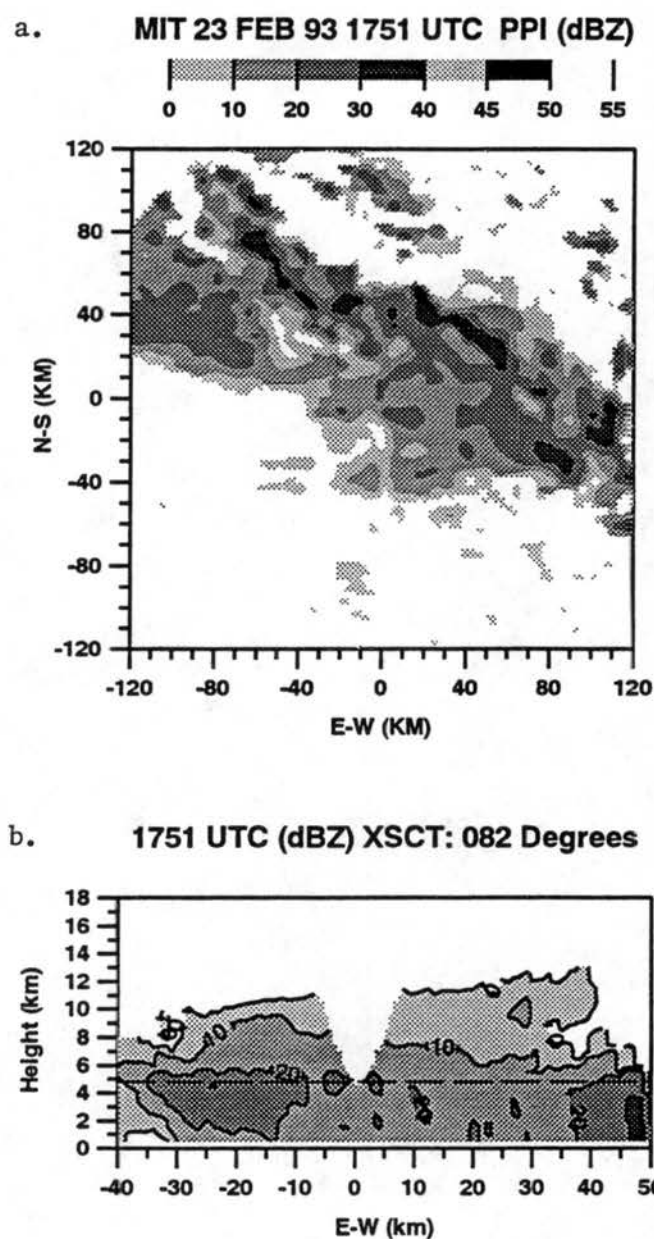


Figure 4.14 a) As in Fig. 4.13a, but 1751 UTC; b) as in Fig. 4.13b but 1751 UTC, azimuth 82°.

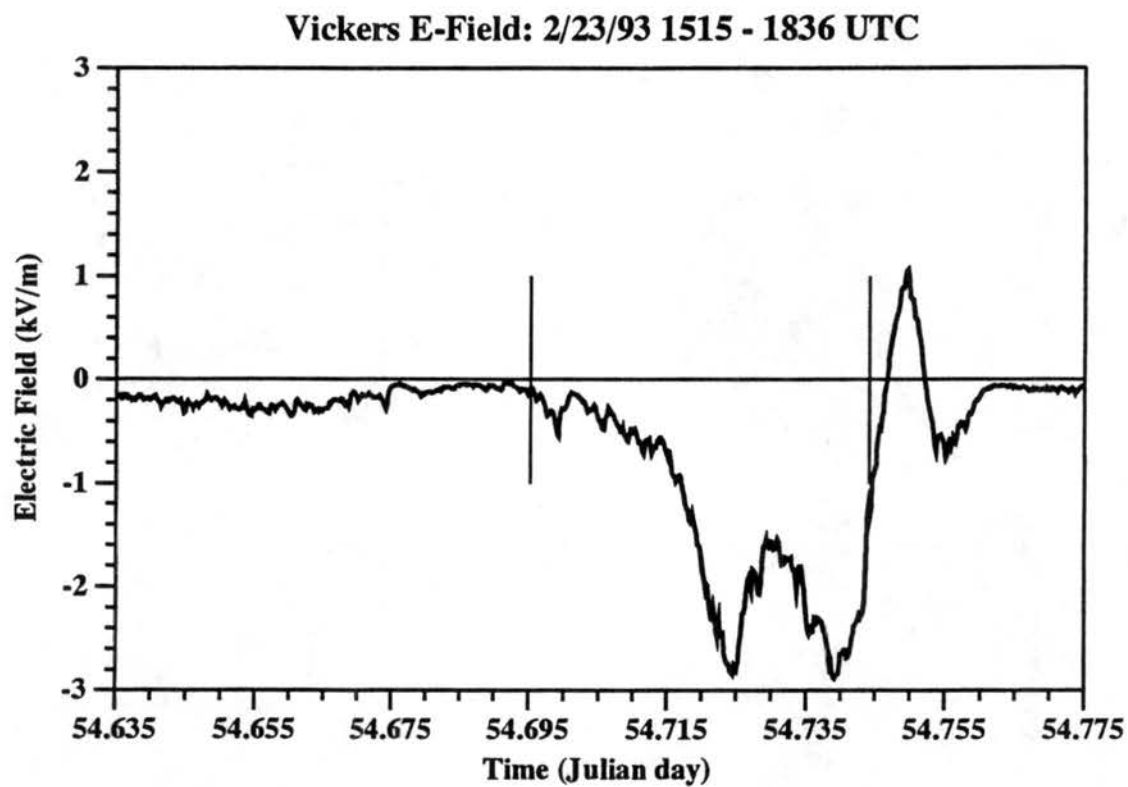


Figure 4.15 Electric-field (kV m^{-1}) evolution with time for 23 February 1993, 1515-1836 UTC. Time (Julian day) is indicated on the abscissa beginning at 1515 UTC. Vertical lines indicate the times 1641 UTC and 1751 UTC respectively.

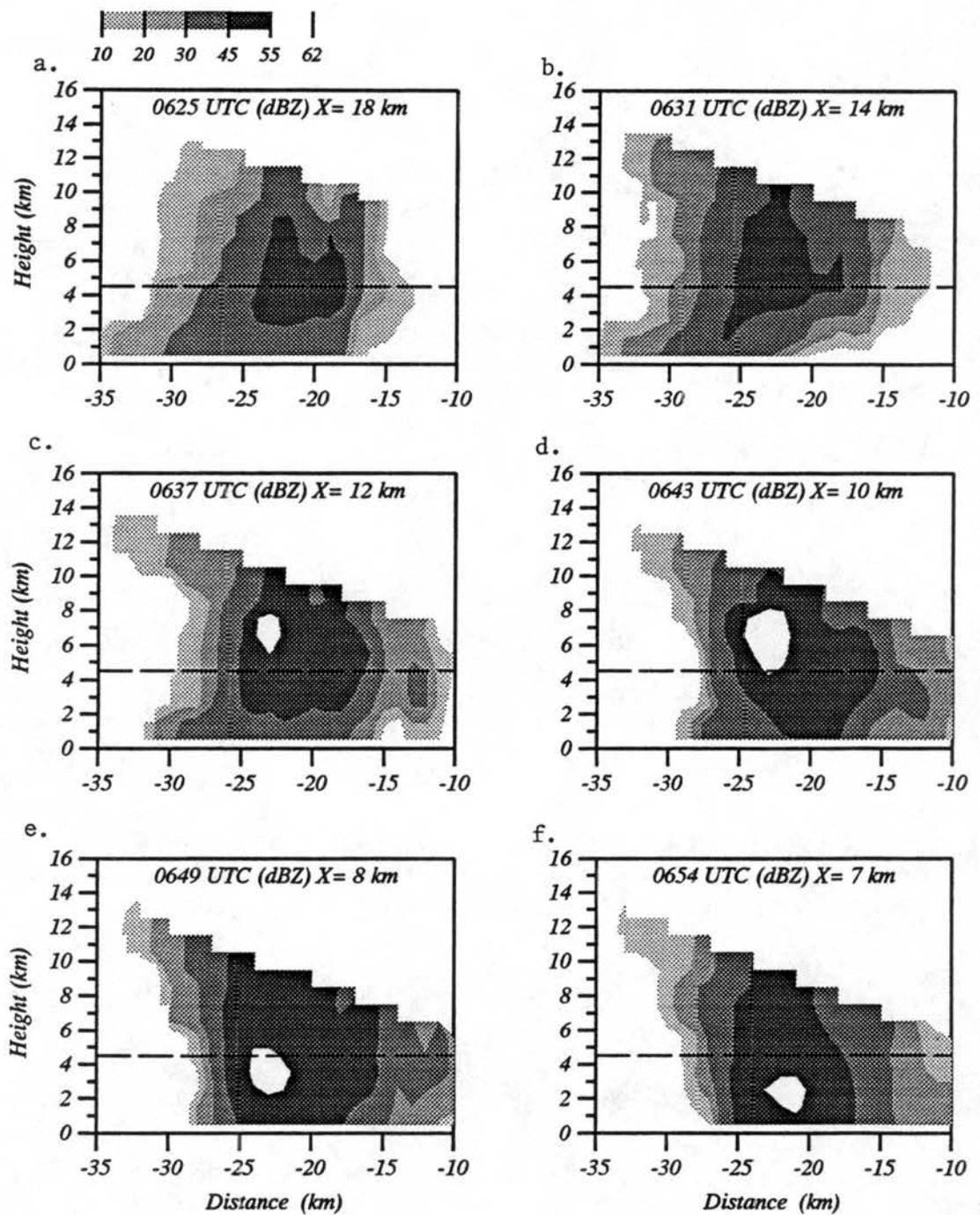


Figure 4.16 Radar reflectivity cross-sections from the PHX NEXRAD radar on 12 August 1994 for: a) 0625 UTC; b) 0631 UTC; c) 0637 UTC; d) 0643 UTC; e) 0649 UTC; f) 0654 UTC.

8/12/94 Case 1: Rain, CG Flash rate, and Delta-V

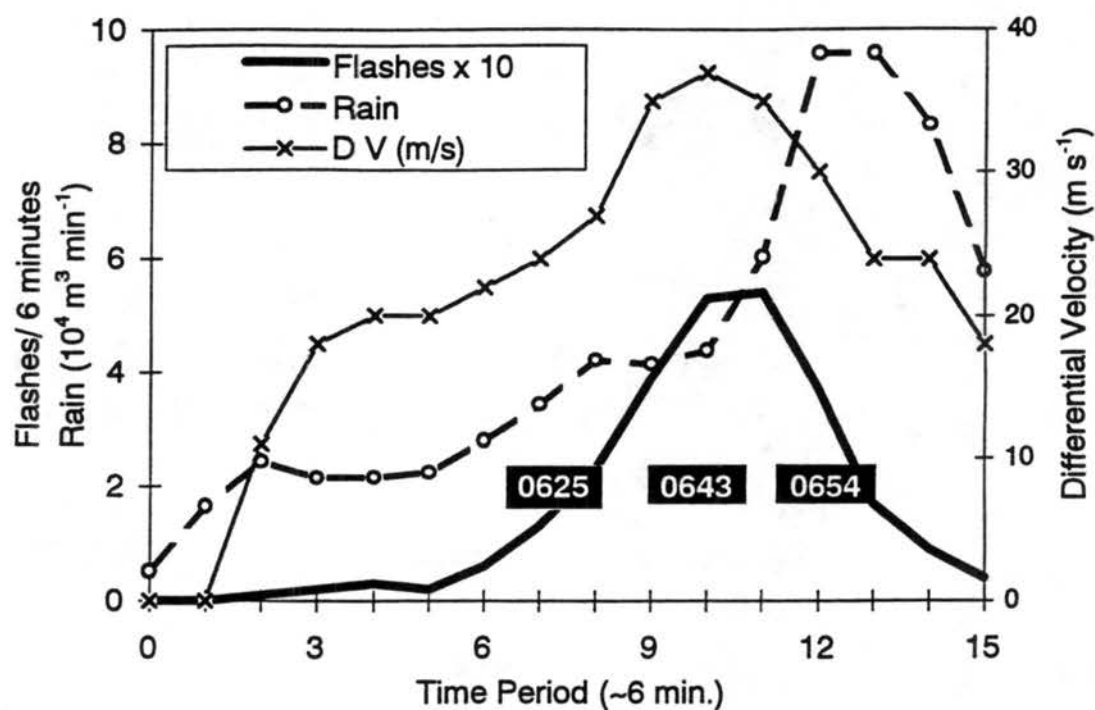


Figure 4.17 Time-series (6-minute intervals) of CG flash count ($\times 10$; left ordinate), rain rate ($\times 10^4 \text{ m}^3 \text{ min}^{-1}$; left ordinate), and differential radial velocity (DV; m s^{-1} , right ordinate) for the thunderstorm observed in Fig. 4.16.

PHX 12 AUG 94 COMPOSITED CELL EVOLUTION (4 CASES)

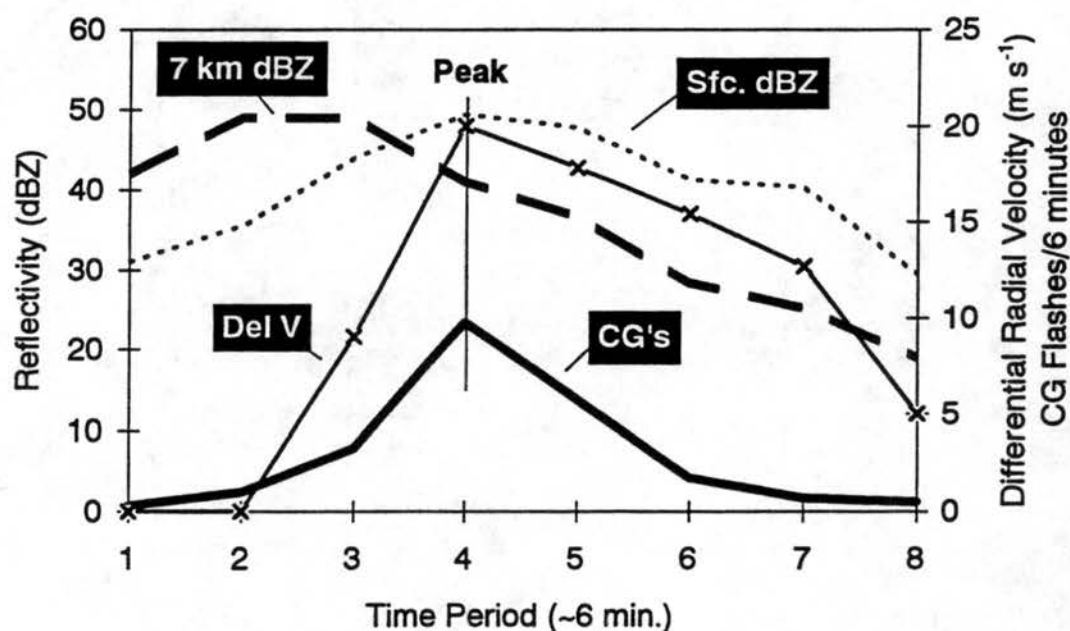


Figure 4.18 Composit ed time-series of radar reflectivity (dBZ; left ordinate) at 7 km and the surface, differential radial velocity (m s^{-1} ; right ordinate), and CG flash counts per 6-minute period (right ordinate), for four cases of electrified convection observed near Phoenix AZ on 12 August 1994.

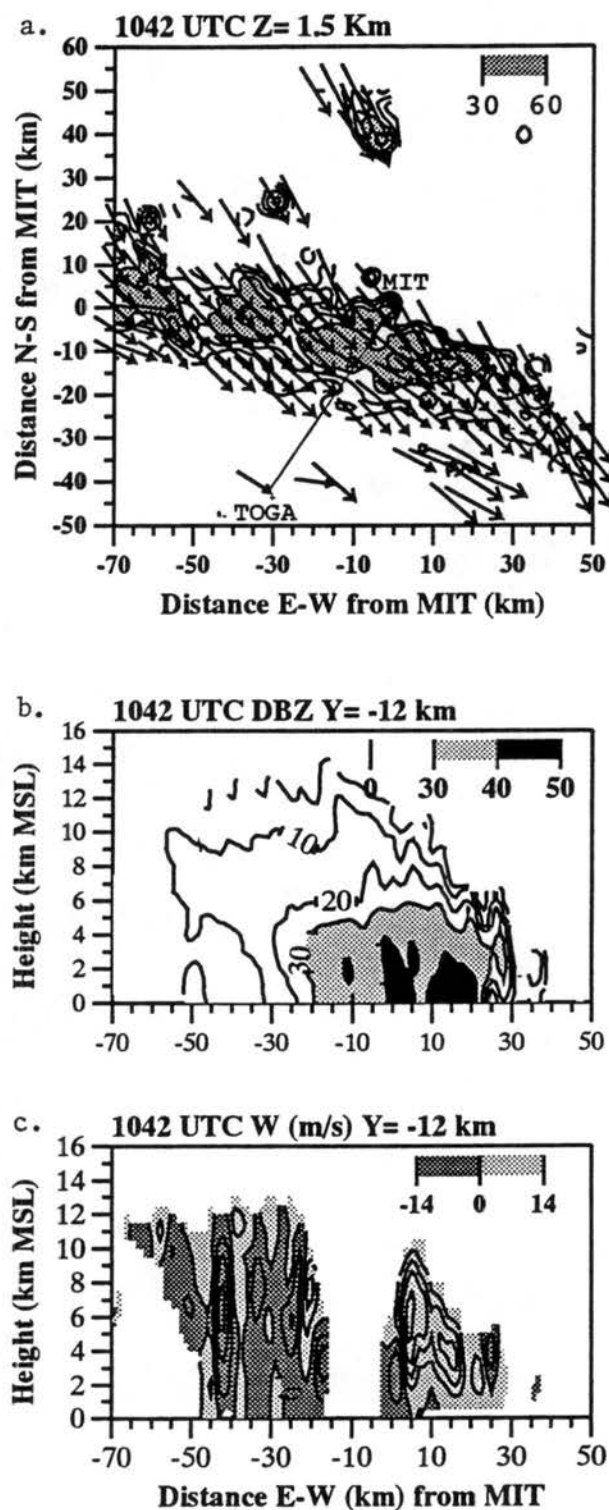


Figure 4.19 Dual-Doppler analysis for the ship case at 1042 UTC, 9 Feb. 1993. a) horizontal cross section of radar reflectivity (dBZ; ≥ 30 shaded) at $Z=1.5$ km and storm relative wind vectors (15 km length = 15 ms^{-1}) centered on the MIT radar. The position of the MIT and TOGA radars is indicated along with the dual-Doppler baseline. b) Vertical cross-section of radar reflectivity made parallel to the X-axis at $Y=-12$ km, reflectivity contoured every 10 dBZ and shaded for values ≥ 30 dBZ. c) same as (b) but vertical velocity, contoured every 1 m s^{-1} .

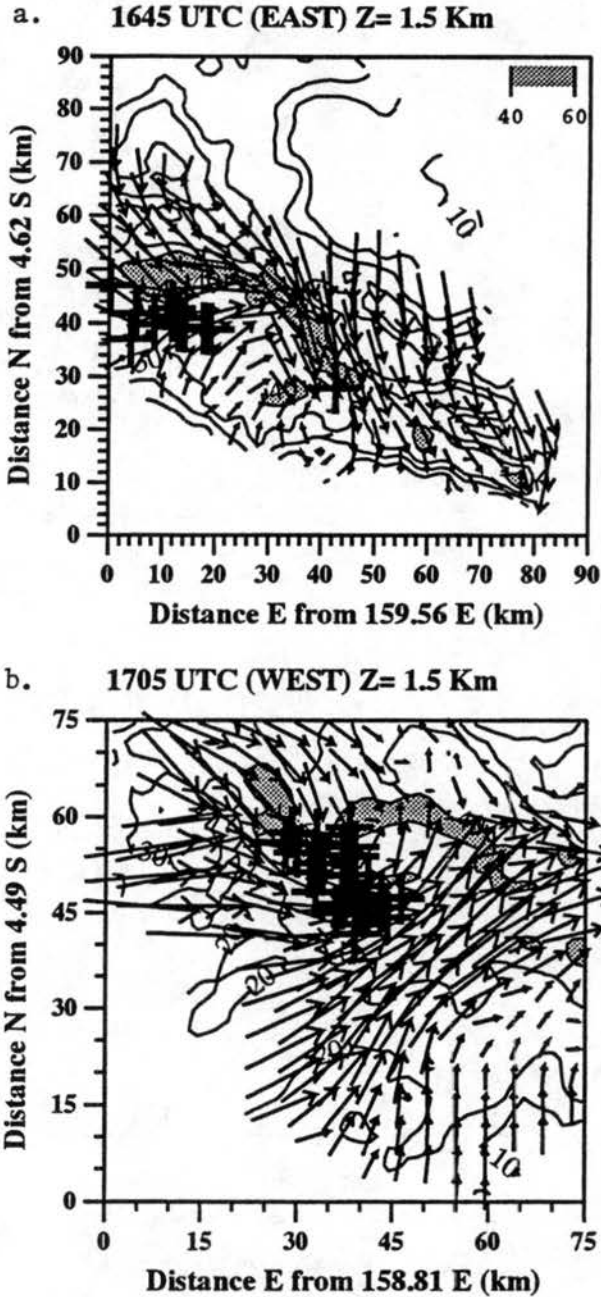


Figure 4.20 As in Fig. 4.19a, but for the aircraft case at a) 1645 UTC eastern side of the system; and b) 1705 UTC western side of the system. Cloud-to-ground lightning flash locations for 20 minutes centered on the time of each cross-section in (a) and (b) are indicated by a (+), flashes were all negative in polarity.

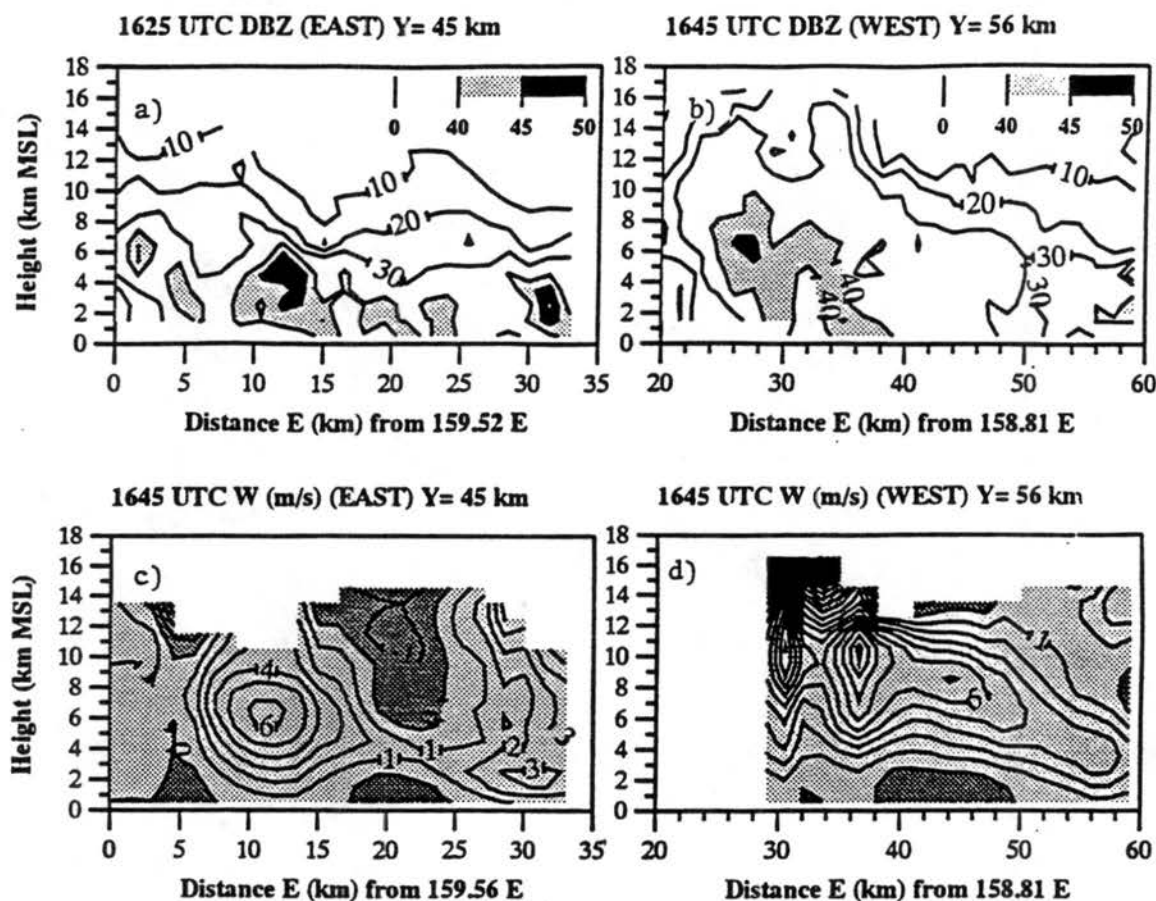


Figure 4.21 a) As in Fig. 4.19b but aircraft case, 1625 UTC, Y=45 km; b) as in (a) but 1645 UTC, Y=56 km; c) vertical velocity 1645 UTC, Y=45 km (eastern side of system), contoured every 1 ms^{-1} ; d) same as (c) but 1645 UTC Y=56 km (western side of the system).

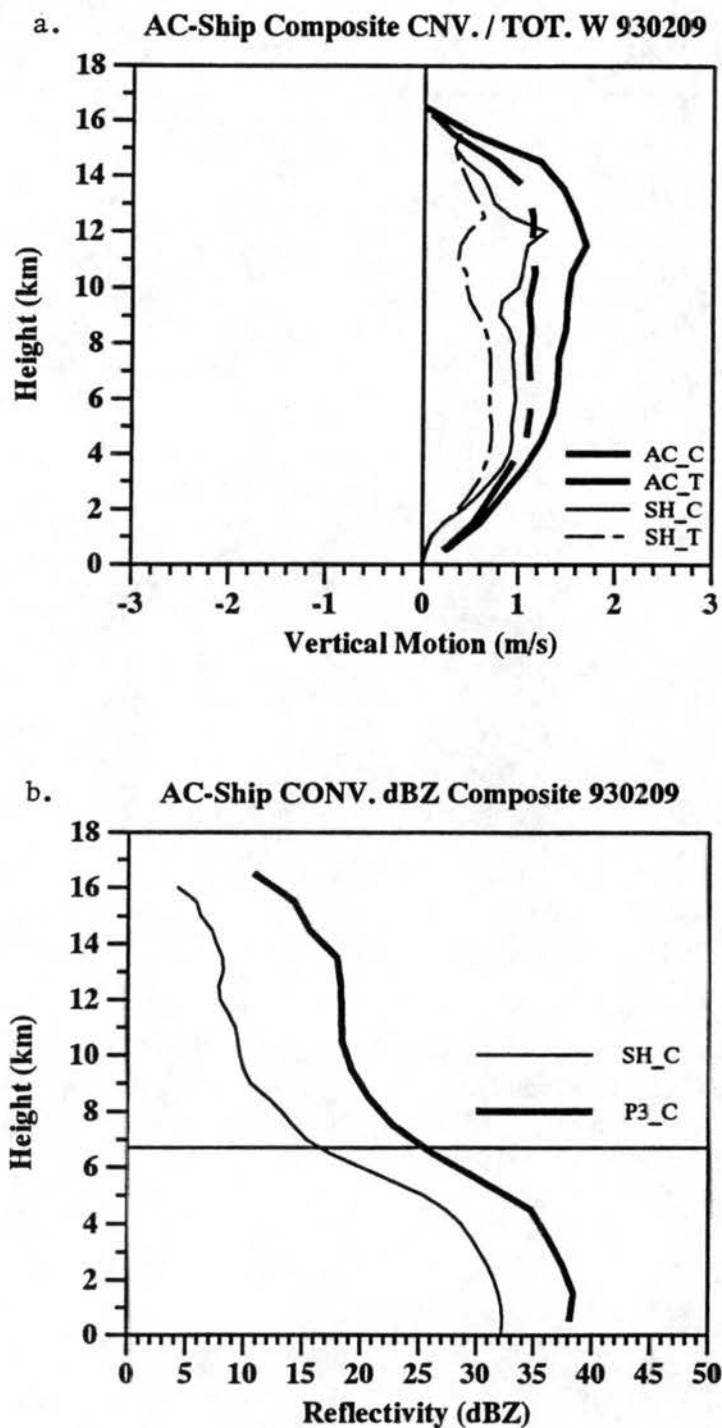


Figure 4.22 a) Composited vertical velocity profiles (updrafts) for the ship and aircraft cases partitioned into convective (bold-solid, aircraft case; light-solid, ship case) and total (bold-dash, aircraft; light dash, ship case) profiles; and b) Composited profiles of partitioned convective radar reflectivities for the ship (light-solid) and aircraft cases (bold-solid). Approximate height of the -10°C level indicated by horizontal line.

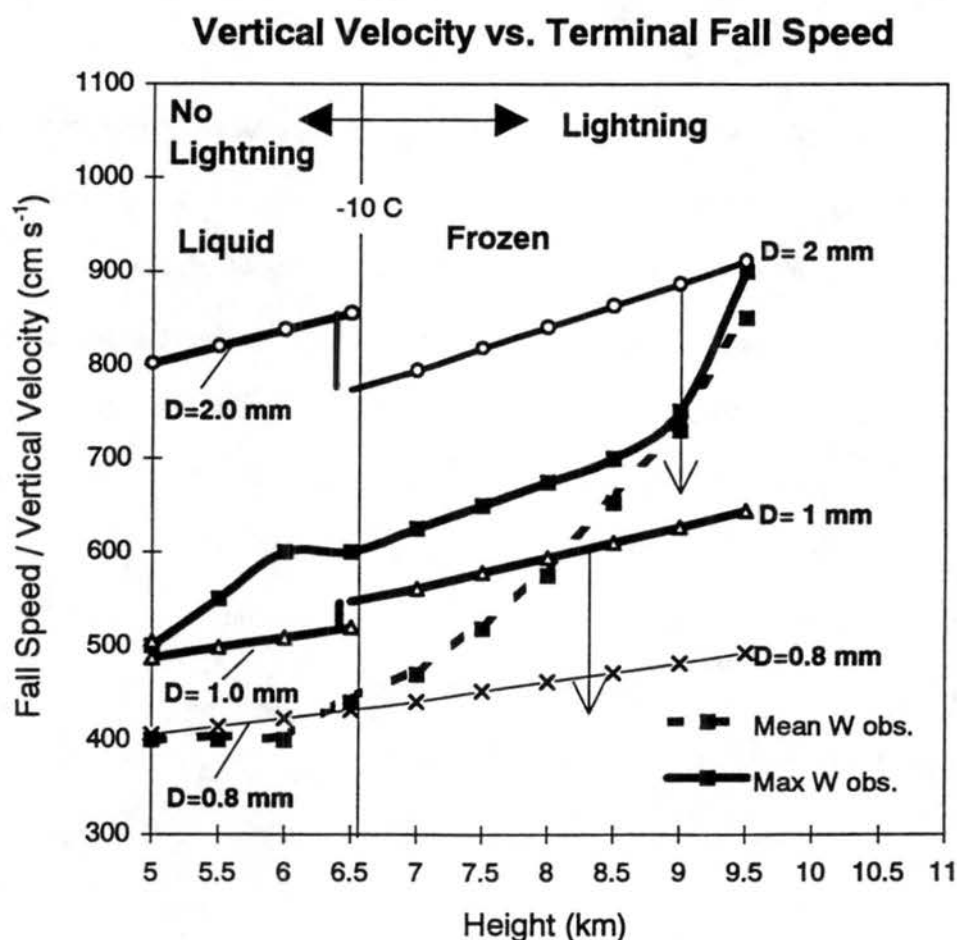


Figure 4.23 Plot of COARE mean (dash-square) and maximum (solid-square) vertical velocities (cm s^{-1} ; ordinate) observed at the top of 30 dBZ reflectivity cores, and terminal fall speed velocities from 0.8 mm, 1 mm and 2 mm liquid and frozen water drops vs. height on the abscissa (cm s^{-1} ; as labeled in the figure). Vertical arrows on fall speed curves for different drop diameters indicate range of error for the fallspeed (based on different fall-speed relationships).

CHAPTER 5

ELECTRICAL, KINEMATIC, AND MICROPHYSICAL EVOLUTION OF DEEP TROPICAL CONVECTION: NUMERICAL MODELING STUDIES

This chapter focuses on cloud modeling simulations performed to explore the *basic* kinematic, microphysical and electrical structures of individual tropical convective clouds as a function of slight changes in their environmental characteristics (e.g., Fig. 3.12). Specifically, the cloud simulations were used as a tool to address questions posed at the end of Chapter 4:

- 1) To what degree does the thermal buoyancy profile control cloud-electrification?
- 2) What role do frozen drops play in tropical cumulonimbus cloud electrification processes (vs. low density graupel)?
- 3) How does the heating/moistening profile change in lightning producing clouds compared to clouds that do not produce lightning discharges?
- 4) Do ice-processes play a larger role in the precipitation production of tropical lightning producing cumulonimbi compared to their non-lightning producing counterparts?

5.1 Model Description

Selection of a specific model was driven by 1) the need for the model to be relatively simple such that the results could be clearly interpreted; 2) the ease with which the model could

be modified; and 3) the computational/time costs in running the model. These requirements, when considered with the scientific goals of the study, dictated the selection of a time-dependent, one-dimensional, bulk-microphysical cloud-model (e.g., Ferrier and Houze, 1989 [FH89]; Simpson et al., 1993; Keenan et al., 1994). The FH89 model was chosen because of its internally consistent, observationally-based representation of updraft dynamics (e.g., an overturning thermal circulation with a variable cloud radius), and its four-ice category (4ICE) representation of bulk cloud-microphysics (e.g., partially described in Simpson et al., 1993 and Keenan et al., 1994). In particular, the 4ICE scheme includes a prognostic equation for frozen raindrops (FD), in addition to prognostic equations for snow, graupel and cloud ice mixing ratio. Observations and modeling studies suggest that addition of FD's as a separate ice category is an important when considering the microphysics of tropical convection (e.g., Takahashi 1978b, 1983, 1990, 1993, 1995; Ferrier, 1994b). This is because FD's play an important role in the development of the ice phase and subsequent cloud electrification in tropical cumulonimbi (e.g., Koenig, 1963; Black and Hallett, 1986; Bringi et al., 1996; Jameson et al., 1996; Ramachandran et al., 1996; French et al., 1996).

The FH89 model has been used previously to simulate tropical oceanic (e.g., Ferrier and Houze, 1989) and tropical island convection (e.g., Simpson et al., 1993; Keenan et al., 1994). However, in order to use the model for examining the microphysical and electrical evolution of tropical convection, modifications to the microphysical scheme were necessary, as was the addition of a parameterization for precipitation based charging. The model "dynamics" (FH89) were not modified. The following section provides an overview of the model.

5.1.1 Model dynamics

Herein we provide a *general description* of the principle equations governing the dynamical processes in the cloud-model. Cloud microphysical and electrical parameterizations are discussed in sections 5.1.2 and 5.1.3 respectively. A more detailed description of the model dynamics can be found in FH89 or Houze (1993). The equations describing the model are formulated in an axisymmetric cylindrical coordinate system (r, λ, z) , where the cloud-radius (r) is permitted to vary as a function of height (e.g., Fig. 5.1). The independent variables λ and z represent azimuthal and height variables respectively. The model prognostic variables (A) include perturbation values ($A = A_e - A_c$; A_e =environment, A_c = cloud) of vertical velocity (w), potential temperature (θ), pressure (p), and mixing ratios of rain (q_r), cloud-water (q_{cw}), water vapor (q_{ve}), cloud ice (q_{ic}), graupel (q_g), snow (q_s) and frozen drops (q_h). In addition to the prognostic variables, diagnostics such as radar reflectivity, precipitation, and vertical profiles of diabatic heating and drying are also computed (e.g., eqs. 3.9-10) at the end of each simulation.

The central prognostic equation for an arbitrary variable averaged over the area of the cloud-cylinder ($\bar{A}$), and the cloud boundary ($\tilde{A}$) is:

$$\frac{\partial \bar{A}}{\partial t} = -\frac{\partial A_e}{\partial t} - \bar{w} \frac{\partial A_e}{\partial t} + \frac{d\bar{A}}{dt} + \quad (5.1)$$

(i)

$$\frac{\tilde{A}}{\rho_e R^2} \frac{\partial \rho_e R^2 \bar{w}}{\partial z} - \frac{1}{\rho_e R^2} \frac{\partial \rho_e R^2 \bar{w} \tilde{A}}{\partial z} - \frac{2}{R} (u'' A'' - w'' A'' \frac{\partial R}{\partial z})$$

(ii) (iii) (iv)

The products of variables in eq. (5.1) labeled as double-primed quantities (e.g., $u'' A''$) represent deviations in the *product* from an average taken along the cloud boundary. Term (i) in eq. (5.1) represents a source-sink term for variable (A_e), the general form of which will be discussed below. Term (ii) represents dynamic entrainment at the cloud-boundary as forced by mass

continuity. The variable in term (ii), $(\tilde{A})$, assumes the mean value of the cloud property $(\bar{A})$ [environment property (A_e)] if the vertical mass flux decreases [increases] with respect to height, representing a detraining [entraining] cloud volume.

Term (iii) represents the vertical transport term (or a flux convergence) and is parameterized by specifying a systematic triangular shaped profile of the variables across the updraft (e.g., similar to the observations of LeMone and Zipser, 1980, and Zipser and LeMone, 1980), together with random turbulent fluctuations for each variable. The turbulent fluctuations are parameterized using diffusive mixing, with separate diffusion coefficients defined for mixing momentum, precipitation, cloud-water, water vapor, and heat (see FH89). Triangular variations in the updraft profile (Fig. 5.1) are specified by parameterizing the value of the updraft as a linear function of radius from the center of the updraft to the edge of the cloud. To create an overturning thermal circulation near the top of the cloud, downward motion is forced along the side of the cloud in the thermal cap region (Fig. 5.1). Term (iv) represents a cloud-boundary lateral eddy mixing term (see FH89). The cloud updraft can be modeled as a “top-hat” profile, though FH89 note that dynamic entrainment prevents the model cloud from developing without unrealistic forcing. We also found this to be the case. The parameterization of an overturning thermal circulation and a triangular shaped updraft profile enables the FH89 model to accurately predict cloud-top height and cloud-water contents simultaneously. The simultaneous accurate prediction of cloud-water and cloud-top height has been problematic in several other 1-D cloud models (e.g., Warner, 1970; Cotton, 1975).

The perturbation pressure in the cloud is determined using mass continuity and the horizontal momentum equation which forces the distribution of the cloud pressure field to be consistent with the specified profile of vertical velocity (FH89; Houze, 1993). The model is integrated numerically using a staggered grid (200 m grid spacing), forward time-differencing

with a variable time-step (minimum time step five seconds), and an upstream differencing technique applied to variables other than the vertical velocity, which employs a second order advection scheme (see discussion in FH89).

Source-sink terms (i) in eq. (5.1) take the following form for vertical momentum, and potential temperature respectively:

$$\frac{d\bar{w}}{dt} = \bar{B} - R_d \theta_{ve} \frac{\partial \bar{P}}{\partial z} \quad (5.2)$$

$$\frac{d\bar{\theta}}{dt} = \frac{L_{f,s,e}}{\pi c_p} (Q_{diabatic}) \quad (5.3)$$

where

$$\bar{B} = \left(\frac{\bar{\theta}}{\theta_e} + 0.61 \bar{q}_v - \bar{q}_L \right) \quad (5.4)$$

represents the cloud buoyancy, $\bar{P}$ the non-dimensional pressure perturbation, and $Q_{diabatic}$ the source/sink of heating associated with phase changes of water (freezing, sublimation, evaporation; $L_{f,s,e}$) in the model domain.

5.1.2 Model microphysics

Continuity equations for each hydrometeor species are presented in eqs. 5.5 a-g (symbols defined in Table 5.1) and include a combination of parameterizations taken from Lin et al., (1983; LFO), Rutledge and Hobbs (1983: RH) and Ferrier (1994a; F94). Terms representing accretional and diffusional growth processes are modeled using continuous collection and the diffusional growth equation respectively, assuming a spherical shape for ice particles (c.f., Young, 1993). For a detailed discussion of the microphysical parameterizations see RH, LFO,

Keenan et al.(1994) and F94. The microphysical continuity equations are written as follows
(note $T \geq 0$ [$T < 0$] implies processes modeled for temperatures greater [less] than 0°C):

$$\frac{d\bar{q}_v}{dt} = -[(P\text{COND} + P\text{REVP}) + (P\text{X}_f\text{MLTEV})_{T \geq 0} + \quad (5.5a)$$

$$(P\text{INT} + P\text{DEPI} + P\text{X}_f\text{DEP} + P(\text{H}, \text{G})\text{WGEV})_{T < 0}]$$

$$\frac{d\bar{q}_{cw}}{dt} = (P\text{COND} - P\text{AUT} - P\text{ACW}) - (P\text{CIMLT} + P\text{MX}_f\text{ACW})_{T \geq 0} + \quad (5.5b)$$

$$(P\text{X}_f\text{ACW} - P\text{CWHFZ}_{T \leq -40^\circ\text{C}} - P\text{IFW}_{T \leq -30^\circ\text{C}})_{T < 0}$$

$$\frac{d\bar{q}_{ci}}{dt} = P\text{CIMLT}_{T \geq 0} + (P\text{INT} + P\text{DEPI} + P\text{CWHFZ}_{T \leq -40^\circ\text{C}} - P\text{CONV} - \quad (5.5c)$$

$$P\text{SAUT} - P\text{IFC} - P\text{XACI} + P\text{HMICE})_{T \leq 0}$$

$$\frac{d\bar{q}_r}{dt} = (P\text{REVP} + P\text{ACW} + P\text{AUT}) + (P\text{MX}_f\text{ACW} + P\text{MRACS} - P\text{X}_f\text{MLT} - \quad (5.5d)$$

$$P\text{X}_f\text{ACRM} - P\text{X}_f\text{ACWM})_{T \geq 0} - (P\text{IFR} + P\text{IACR} + P\text{X}_f\text{ACR})_{T < 0} + V\text{TR}$$

$$\frac{d\bar{q}_s}{dt} = -(P\text{GACS} + P\text{HACS}) + (P\text{SMLT} + P\text{SMLTEV} + P\text{SACWM} + P\text{SACRM} - \quad (5.5e)$$

$$P\text{MRACS})_{T \geq 0} + [(P\text{SDEP} + P\text{CONV} + P\text{SAUT} + P\text{IFW} + P\text{IFC} + P\text{SACI} - P\text{GAUT} +$$

$$P\text{SACW} \cdot \delta_3 - P\text{WACS} + (P\text{IFR} + P\text{RACI} + P\text{IACR} + P\text{SACR}) \cdot \delta_1$$

$$- P\text{RACS} \cdot (1 - \delta_1)]_{T < 0} + V\text{TS}$$

$$\frac{d\bar{q}_g}{dt} = P\text{GACS} + (P\text{GMLT} + P\text{GMLTEV} + P\text{GACWM} + P\text{GACRM})_{T \geq 0} + \quad (5.5f)$$

$$(P\text{GDEP} + P\text{GWGEV} + P\text{GAUT} + P\text{GACI} + P\text{GACS} + P\text{SACW} \cdot (1 - \delta_3) + P\text{WACS} +$$

$$(P\text{GACW} - P\text{IHMG}) \cdot \delta_4 - P\text{WACG} + (P\text{IFR} + P\text{RACI} + P\text{IACR} +$$

$$P\text{RACS} + P\text{SACR}) \cdot (\delta_2 - \delta_1) + P\text{GACR} \cdot \delta_2 - P\text{RACG}(1 - \delta_2))_{T < 0} + V\text{TG}$$

$$\frac{d\bar{q}_h}{dt} = P\text{HACS} + (P\text{HMLT} + P\text{HMLTEV} + P\text{HACWM} + P\text{HACRM})_{T \geq 0} + \quad (5.5g)$$

$$\begin{aligned}
& (\text{PHDEP} + \text{PHWGEV} + \text{PHACW} - \text{PIHMH} + \text{PHACI} + \text{PHACR} + \text{PHACS} + \\
& \text{PGACW} \cdot (1 - \delta_4) + \text{PWACG} + (\text{PIFR} + \text{PRACI} + \text{PIACR} + \text{PRACS} + \\
& \text{PSACR} + \text{PGACR} + \text{PRACG}) \cdot (1 - \delta_2))_{T < 0} + \text{VTH}
\end{aligned}$$

The term VTJ is written:

$$\text{VTJ} = \frac{1}{\rho_e R^2} \frac{\partial(\rho_e R^2 V_J q_J)}{\partial z} \quad (5.5h)$$

and represents a cloud-area averaged terminal fall speed (mass-weighted) convergence for species (J), where V_J is the mass weighted terminal fall speed for precipitation species $j = g, h, r, s$. (Note that the subscript “h” will be used to indicate the higher density ice-category (FD) to maintain consistency with LFO, RH and F94).

Particle size distributions for the precipitation species are parameterized using inverse exponential distributions of the general form $N(D) = N_0 \exp(-\lambda D)$, where N_0 is the intercept of the distribution (Table 5.2a) and λ is the slope (a function of both the intercept and mixing ratio of the species; e.g., LFO, RH). Based on the observations of Houze et al. (1979) and the modeling study of F94, the value of N_0 for the snow category was allowed to increase as a function of temperature to a maximum value of 10^7 m^{-4} to reflect enhanced aggregation efficiencies at warmer temperatures (only minor differences in the results occurred when the exponential increase was turned off). Intercepts for rain, graupel and FD’s were held constant in the simulations (Table 5.2a). Sensitivity tests were conducted to evaluate the response of the model to variations in N_0 (since N_0 is likely to vary in real convection). The overall character of simulations remained the same for *small* (factor of 2-4) deviations in N_0 . However, the results were sensitive to variations *larger* than a factor of five in N_0 . Selection of individual N_0 ’s for each simulation were based on the sensitivity tests, observations and previous modeling studies such as LFO, RH, F94, and Krueger et al. (1995). Cloud-water and cloud ice species were

Table 5.1: Microphysical production terms in equations (5.5 a-g)

Symbol	Description (units = $\text{kg m}^{-3} \text{ s}^{-1}$)
PCOND	Condensation of water vapor
PDEPI	Depositional growth of cloud ice
PX _r DEP	Depositional growth of *X _r : * X_r = snow, graupel, FD
PX _r MLTEV	Evaporation of melting X _r
PINT	Production of cloud ice through primary nucleation
PREVP	Evaporation of rainwater
P(H,G)WGEV	Evaporation of FD and graupel in wet growth
PRAUT	Autoconversion of cloud-water to rain; rate coefficient = .001 s ⁻¹ ; Threshold = .0005 kg kg ⁻¹ (NACT, ACT); .001 kg kg ⁻¹ (IT); .002 kg kg ⁻¹ (MC)
PGAUT	Autoconversion of snow to graupel; rate coefficient = F(temperature). Threshold = .0006 kg kg ⁻¹
PSAUT	Autoconversion of cloud ice to snow; rate coefficient .001 s ⁻¹ Threshold = .001 kg kg ⁻¹
PRACW	Production of rain by collection of cloud-water
PCIMLT	Production of cloud-water by melting cloud ice
PMX _r ACW	Collection of cloud-water by species X _r during melting
PX _r ACW δ ₃ , δ ₄	Collection of cloud-water by species X _r ; δ ₃ , δ ₄ determine fractions of mass distributed to graupel and FD categories in three component freezing processes between cloud-water, snow (δ ₃) and graupel (δ ₄) respectively (c.f., LFO, RH, Farley et al., 1989)
PCWHFZ	Production of cloud ice by homogeneous freezing of cloud-water
PIFW	Bergeron process conversion of cloud-water to snow
PIFC	Bergeron process conversion of cloud ice to snow
PXACI δ ₁ , δ ₂	Collection of cloud ice by species X: *X = rain, snow, graupel, FD For three component freezing (e.g., PRACI), mass placed in snow for $\bar{D} < .5 \text{ mm}$; graupel, $.5 \leq \bar{D} < 1 \text{ mm}$; and FD if $\bar{D} \geq 1 \text{ mm}$; δ ₁ , δ ₂ determine the fraction of rain mass placed in each category (e.g., Farley et al., 1989)
PHMICE	Total rime-splinter production of cloud ice (graupel and FD)
PIHMG	Rime-splinter production of cloud ice by graupel
PIHMH	Rime-splinter production of cloud ice by FD
PMRACS	Production of rain through collection of melting snow
PX _r MLT	Production of rain through melting of species X _r
PX _r ACRM	Production of rain through enhanced melting of species X _r in collisions with rain
PX _r ACWM	Production of rain through enhanced melting of species X _r during collection of cloud-water
PIFR	Probabilistic freezing of rain (mass distributed as in PXACI)
PIACR	Contract freezing of rain by cloud ice (distributed as in PIFR)
PX _r ACR	Collection of rainwater by species X _r (distributed as in PIACR)
P(G,H,R)ACS	Collection of snow by graupel, FD, and rain. Three component freezing in the PRACS process employs δ ₁ , δ ₂ as in PXACI.
PCONV	Conversion of cloud ice to snow
PWAC(G,S)	Collection of graupel and snow by cloud-water
PRACG δ ₂	Collection of graupel by rain; δ ₂ determines fraction of mass that goes to the graupel and FD categories as in PXACI.

Table 5.2a: Hydrometeor distribution intercepts/fall-speed coefficients

Parameter	NACT	ACT	IT	MC
$N_{\alpha}(\text{Rain}) \text{ m}^{-4}$	8×10^6	8×10^6	8×10^6	8×10^6
$N_{og}(\text{Graupel}) \text{ m}^{-4}$	4×10^6	4×10^6	4×10^6	5×10^5
$N_{\infty}(\text{Snow})^* \text{ m}^{-4}$	3×10^6	3×10^6	3×10^6	3×10^6
$N_{oh}(\text{FD}) \text{ m}^{-4}$	1×10^6	1×10^6	1×10^6	1×10^6

*intercepts allowed to increase exponentially to a maximum of $1 \times 10^7 \text{ m}^{-4}$ as a function of decreasing temperature (Ferrier, 1994b)

Table 5.2b: Hydrometeor fall-speed coefficients (mks units)

Coefficient	Rain	Snow	Graupel	FD
a	842	4.836	19.3	140
b	0.8	0.25	0.37	0.5

assumed monodisperse and to possess a negligible fallspeed. Terminal fall speed relationships were formulated as a function of particle diameter in power-law form ($V_f = aD^b$; LFO, RH, F94; Table 5.2b) and integrated over the particle size distributions to provide requisite mass weighted fallspeeds.

The original version of the model (as provided by Dr. Brad Ferrier, NASA GSFC) included parameterizations for neither the probabilistic freezing of raindrops (e.g., Bigg, 1953; LFO) nor the production of ice splinters during riming of frozen drops and graupel (e.g. Hallett and Mossup, 1974; HM74). Due to the hypothesized importance of these processes in tropical clouds (e.g., Koenig, 1963; Gamache, 1990; Young, 1993; F94; Jameson et al., 1996), parameterizations for these processes were added to the model. The parameterization for probabilistic freezing (PIFR) follows LFO, and played only a minor role in the production of hail, graupel and snow.

The addition of a simple rime-splintering parameterization played a much larger role in the model microphysics for temperatures between -3°C and -8°C . The rime-splintering scheme is a

direct representation of the laboratory results reported in HM74, and is formulated similar to that summarized in Gordon and Marwitz (1981) and Cotton et al. (1986). Specifically, the number concentration of small ice splinters (NCIHM) produced per gram of rime collected by FD's and graupel ($PX_{gh}ACW$) at each model time-step (δt), is represented in the following fashion (temperatures $[T]$ between $-3^{\circ}C$ and $-8^{\circ}C$ and cloud water contents $\geq 1 \text{ g m}^{-3}$):

$$NCIHM(m^{-3}) = \delta t (s) \cdot 3.5 \times 10^8 (kg^{-1}) \cdot PX_{gh}ACW (kg m^{-3} s^{-1}) \cdot F_x \quad (5.6a)$$

$$F_x = |a + T(^{\circ}C)| / b \quad (5.6b)$$

$$a = 3, b = 3 \text{ for } -3^{\circ}C > T \geq -5^{\circ}C; \text{ and } a = 8, b = 4 \text{ for } -5^{\circ}C > T \geq -8^{\circ}C \quad (5.6c)$$

It should be noted that rime-splintering is highly dependent on the cloud droplet size distribution (HM74, Mason, 1988), however the cloud-droplets in this bulk model are assumed monodisperse, therefore a threshold in cloud-water content is assumed based on the bulk liquid water measurements made in the HM74 laboratory experiment. In order to account for the dependence of ice particle splintering on graupel/FD residence time in the $-3^{\circ}C$ to $-8^{\circ}C$ region of the cloud (e.g., HM74, Mason, 1996), we do not allow the HM74 parameterization to activate unless the difference between the *mass weighted* terminal fall speed of a particle and the updraft is $\leq 2.0 \text{ m s}^{-1}$ (e.g., so that the residence time in the $\sim 1 \text{ km}$ deep region is at least 5 minutes, similar to Mason, 1996). Note that the relative speed difference of 2 m s^{-1} is also consistent with the particle target velocity used in HM74. In this manner we attempt to couple the dynamics of the cloud to the production of ice splinters, although this coupling is indirect (e.g., Ferrier, 1994b).

The other major source of ice-particles (or mass) is of course primary ice-nucleation. Primary ice-nucleation is parameterized using the Meyers et al., (1992) nucleation scheme, which treats ice-nucleation as a combined function of the ambient temperature and ice-supersaturation. Homogeneous nucleation of ice from cloud-water also occurs at temperatures $\leq -40^{\circ}C$. Due to

the importance of producing representative ice particle number concentrations for input to the electrical parameterization (e.g., Schuur, 1996), a continuity equation for ice-particle number concentration was also added to the 1-D model. In this formulation, a separate continuity equation (e.g., Schuur, 1996) accounts for the sources (primary nucleation, rime-splintering), advection (e.g., F94), mixing, and sinks of ice crystal number concentration. Sinks of ice particle number concentration were computed using the sinks of ice mass (eq. 5.5c; Table 5.1) which were divided by either a mean ice particle mass ($O[10^{-11} \text{ kg}]$), or by a larger ice-mass ($O[10^{-10} \text{ kg}]$) if the loss occurred through aggregation (auto-conversion) or conversion by the Bergeron process. Tests of the ice-nucleation schemes (primary and secondary) in the 1-D model produced a vertical profiles of cloud ice concentration similar to those observed in tropical convection (e.g., Houze and Churchill, 1984; Black and Hallett, 1986; Gamache, 1990).

5.1.3 Electrification parameterization

Numerous laboratory studies (e.g., Reynolds et al., 1957; Takahashi, 1978 [T78]; Jayaratne et al., 1983 [J83]; Baker et al., 1987; Keith and Saunders, 1990 [KS90]; Williams et al., 1991[W91]; Saunders et al., 1991 [SKM91]; Avila and Caranti, 1994; Baker and Dash, 1994; Brooks et al., 1996) have indicated that significant charge transfer occurs when millimeter sized rimed ice targets (i.e., graupel) collide with smaller ice-crystals (diameters of $\sim 100 \mu\text{m}$) in the presence of supercooled water droplets. This charge transfer occurs independent of the electric field strength and is referred to as “non-inductive” charge transfer (as opposed to the “inductive” mechanism which does depend on the electric field vector; see Sec. 1.3.1). The amount and sign of the non-inductive charge transfer (NIC) is a function of temperature and liquid water content (e.g., Fig. 1.9). Observational (e.g. Williams, 1989; 1995) and modeling

evidence (Illingworth and Latham, 1977; Takahashi, 1984; Gardiner et al., 1985; Ziegler et al., 1986; Helsdon and Farley, 1987; Ziegler and MacGorman, 1994; Randell et al., 1994) suggest that the NIC mechanism is the dominant charging process in lightning producing cumulonimbus clouds. Therefore, we based the current model electrification process on the NIC mechanism.

Parameterization of the NIC process in previous studies has been based on laboratory results reported in T78, SKM91, or J83 (e.g., Takahashi, 1984; Gardiner et al., 1985; Ziegler et al., 1986; Norville et al., 1991; Ziegler et al., 1991; Randell et al., 1994; Solomon and Baker, 1994; Scavuzzo and Caranti, 1996; Schuur, 1996). Recent discussion in the literature (e.g., Williams, 1993, 1995b, 1996; Saunders, 1994, 1996; Brooks and Saunders, 1995) has cast serious doubt upon *which* laboratory results (e.g., T78, J83, KS90, SKM91) are suitable for use in a numerical model. The controversy is rooted primarily in the definition of “cloud-water” content (LWC), and “effective” cloud-water content; (EWC; e.g., SKM91). The EWC is a function of the cloud-drop size distribution (in addition to rimer velocity etc.) and represents only that fraction of the LWC that is actually collected by a riming particle (e.g., SKM91).

The laboratory results of SKM91 report charge transfers in temperature-*EWC* space, whereas T78 defines the charge transfer in a temperature and *LWC* space. If the two sets of results are compared after assuming that EWC is approximately equal to one half the LWC (SKM91), Fig. 5.2 results (MacGorman and Rust, 1997¹). Figure 5.2 suggests some overlap in the laboratory results at LWC's and temperatures important to thunderstorm electrification for temperatures colder than -20°C (consistent with observations). Yet there is significant discrepancy between the two sets of results for temperatures $\geq -15^{\circ}\text{C}$ and at low LWC's. This poses a question with regard to *which* set of laboratory results should be used in the 1-D model.

¹ The Electrical Nature of Storms (Oxford University Press; to be published 1997)

Since we had no information *a priori* concerning which results would work best in the model, both sets of results were parameterized and tested.

The advection and mixing of charge in the model cloud is accomplished using prognostic equation (5.1) for the variable Q_x (nC m^{-3}), which represents the electrical charge *residing* on hydrometeor species x . All of the hydrometeors (cloud-water, ice, snow, graupel and frozen drops/hail [FD/H]) are capable of carrying electrical charge. At the end of each model integration, the net charge in the cloud (Q_c) at each grid point is computed by summing over the charge residing on each of the hydrometeor species:

$$Q_c = \sum_x Q_x \quad (5.7)$$

Since charge is assumed to reside *on* the hydrometeor species, Q_x is advected and mixed in precisely the same manner that each of the hydrometeor categories (cloud-water, cloud ice, snow, graupel and FD/H) are advected and mixed. The charge conservation equation (source-sink term in equation 5.1) including components due to non-inductive charge transfer, and production terms associated with conservation of charge during mass transfers between ice/water categories is written as:

$$\frac{dQ_x}{dt} = \left(\frac{dQ_x}{dt} \right)_{\text{NIC}} + \left(\frac{dQ_x}{dt} \right)_m \quad (5.8a)$$

where NIC represents the collisional based charging and “m” refers to microphysical processes that transfer charge between categories (e.g., via the production terms in eqs. 5.5 a-g). NIC charging can occur for collisions between cloud ice and “rimers”, including graupel or FD’s, snow and graupel or FD’s, and in collisions between graupel and FD’s. For NIC charging of species (x) (including *graupel and FD’s*), collisional charge transfers with cloud ice (i) are represented in continuous collection form as,

$$\left(\frac{dQ_x}{dt}\right)_{\text{nic}} = \int_0^{\infty} \frac{\pi}{4} D_x^2 |\bar{V}_x| \epsilon_{xi} N_x N_i \delta q dD_x \quad (5.8b)$$

which when integrated over the distribution of species (x), becomes,

$$\left(\frac{dQ_x}{dt}\right)_{\text{nic}} = \frac{\pi}{4} |\bar{V}_x| \epsilon_{xi} N_i N_{ox} \frac{\Gamma(3)}{\lambda_x^3} \delta q \quad (5.8c)$$

Here, V_x = terminal fallspeed of graupel or FD's; $N_{i,x}$ = number concentration of graupel or FD's; N_{ox} = Intercept of the size distribution for graupel or FD's, ϵ_{xi} = separation efficiency ($\epsilon_{xi} = 1$ if "x" has a dry surface, 0 if any portion of the surface is wet); Γ represents the gamma-distribution, λ_x = slope of the distribution, and δq = charge transferred per collision (nC).

Similar to (5.8c), for collisions between categories x = graupel or FD's and, y = snow, (or graupel if x = FD), we integrate the NIC term in (5.8a) over *both* size distributions to yield,

$$\left(\frac{dQ_x}{dt}\right)_{\text{NIC}} = \frac{\pi}{2} |\bar{V}_x - \bar{V}_y| N_{ox} N_{oy} \epsilon_{xy} \delta q \left(\frac{1}{\lambda_x^3 \lambda_y} + \frac{1}{\lambda_x^2 \lambda_y^2} + \frac{1}{\lambda_x \lambda_y^3} \right) \quad (5.9)$$

In collisions between FD's and graupel, the FD is treated as the "rimer" with regard to the sign of charge transfer (e.g., Fig. 1.9).

The amount of charge transferred to the rimer in each collision (δq) is computed two separate ways depending on which parameterization is used (e.g., T78 or SKM91). If the SKM91 parameterization is used, δq is computed as,

$$\delta q = B D_i^a V^b q' \quad (5.10)$$

where the sign and magnitude of q' (nC), the charge transferred to the rimer, is computed from empirically-based equations presented in SKM91 using an EWC that is $\sim 0.5 \cdot \text{LWC}$. In (5.10), D_i represents the mean diameter of the smaller ice or snow particle, and (V) represents the absolute difference in mass weighted terminal fall speed for the two hydrometeors involved. The coefficient (B), the exponents associated with the mean diameter of the ice/snow (a ; 0.5 to 3.76

as a function of sign and diameter), and the relative collision velocity (b ; values of 2.5 to 2.8) are further described in SKM91.

The computation of δq in the T78 parameterization is accomplished through the use of a look-up table that contains values of charge transfer (q') shown in Fig. 1.9 as a function of temperature and LWC (after Randell et al., 1994; Schuur, 1996). Following Takahashi (1984) and Marshall et al. (1978), a velocity-diameter dependence on the charge transfer is also included in the T78 parameterization:

$$\delta q = 5 \left(\frac{\overline{D}_y}{D_o} \right) \frac{|\overline{V}_x - \overline{V}_y|}{V_o} \cdot q' = \alpha q' \quad (5.11)$$

In (5.11), $(\overline{D}_y)$ = mean diameter of ice, snow or graupel, $D_o = 100 \mu\text{m}$, $V_o = 8 \text{ m s}^{-1}$, $x = \text{graupel}$ or FD , $y = \text{ice, snow, or graupel if } x = \text{FD}$. The coefficient (α) in (5.11) is not permitted to exceed a value of 10 (charge transfer values exceed laboratory measured values; Marshall et al., 1978; Takahashi, 1984). Indeed, due to the *bulk representation* of the model microphysics which use a mass-weighted terminal fall-speed and mean diameter in eqs. (5.10-5.11), the maximum charge transfer allowed in either of the two parameterizations (T78 or SKM91) is constrained to be 250 fC per collision distribution (c.f., Ziegler et al., 1991), which is near the upper-limit of charge transfer values typically observed in laboratory experiments under “mean” conditions (e.g., KS90, SKM91). If the charge transfer is not limited, the velocity-diameter dependence (e.g., eqs. 5.10-5.11) causes excessive charging rates in the model resulting in unrealistic vertical profiles of charge density (e.g., inverted dipole structures which are rarely observed in nature).

The NIC mechanism is activated when temperatures in the cloud are colder than -5°C , and LWC's exceed $.02 \text{ g m}^{-3}$. Once initiated, the conservation (second) term on the right hand side of eq. (5.8a) becomes important. This term is calculated at each time-step by multiplying the mass

transfer terms in equations (5.5 b-g), by the charge per unit mass (nC kg^{-1}) in each respective mass category. Conservation of charge at each time-step and grid point is then accomplished by adding (subtracting) charge to (from) the receiving (transferring) category. Charge that resides on evaporated or sublimated hydrometeor mass is considered lost (e.g., ions are not reattached to other particles). Ions are not included in this version of the parameterization (hence no ion diffusion, attachment, drift currents or screening layers are modeled), and given the nature of the simulations, it is unlikely that their inclusion would alter the *primary* result (e.g., Helsdon and Farley, 1987). We should also note that the charge densities at upper and lower boundaries of the model are set to zero.

Lightning flashes are not parameterized in the model and hence total flash rates cannot be determined. However, the electric field magnitude (vertical component) is computed for grid-points staggered with respect to the modeled charge density. The electric field computation is performed at each grid point along the centerline of the model cloud using the method of images (e.g., Uman, 1987) for a shallow cylindrical distribution of the individual charge volumes (approximated as disks) over the depth of the domain (Fig. 5.3). The affect of the image charge *in the cloud* can arguably be neglected as it's contribution to the vertical electric field along the z-axis is small. However to be complete, the image charge is retained in the formulation of the electric field. Computation of the E-field in this fashion reflects the disk-like geometry of the model grid (Fig. 5.3). This results in the following equation for the *vertical component* of the electric field at grid point (j), for a height H_j above ground, when summing over all disks and images (i) of charge density ρ_{ϕ} (nC m^{-3}), radius R and thickness Δz (200 m), located δz_i ($z_j - z_i$; Fig. 5.3) away from grid point j;

$$\vec{E}_{zj} = \sum_i \frac{\rho_{qi} \Delta z}{2\epsilon_o} \left[\underset{\text{i}}{\text{sgn}(\delta z_i) \left(1 - \frac{|\delta z_i|}{(\delta z_i^2 + R^2)^{\frac{1}{2}}} \right)} + \underset{\text{ii}}{\left(\frac{2H_j - \delta z_i}{[(2H_j - \delta z_i)^2 + R^2]^{\frac{1}{2}}} - 1 \right)} \right] \quad (5.12)$$

Term (i) represents the electric field at point (j) due to a single disk of charge q_i . Term (ii) represents the small contribution to the electric field associated with the image charge ($-q_i$) induced in the earth by the cloud charge layer q_i . In term (i), $\text{sgn}(\delta z_i)=1$ for $\delta z_i > 0$, -1 for $\delta z_i < 0$ (e.g., positive q_i located over point (j) produces a downward pointing electric field; Fig. 5.3). In addition, the fair-weather electric field distribution (exponential decrease in height; e.g., Helsdon, 1980) is assumed for the initial vertical profile of the electric field.

Electrical-charging of the cloud is suspended when the electric field at any grid point in the domain exceeds one and a half times the “break-even” electric field for that grid-point height (personal communication Dr. W.D. Rust, NOAA-NSSL). The “break-even” electric field is defined as the electric field required to maintain the average energy (through acceleration) of an energetic (1 MeV) electron as energy is lost by that electron during ionizing collisions with other molecules in the atmosphere (Marshall et al., 1995b). Since the density of the atmosphere decreases exponentially with height, the break-even electric field also decreases exponentially with height. The break-even field is approximately 10 times smaller than the dielectric strength of clear air $O[10^3 \text{ kV m}^{-1}]$. Previous modeling studies of cloud-electrification have employed *arbitrary* thresholds of electric field strength to determine when break down occurs in the simulated cloud (e.g., 400 kV m^{-1} ; Helsdon and Farley, 1987; 250 kV m^{-1} , Solomon and Baker, 1996) or have allowed electrical processes to continue in the model even after dielectric break down of the medium would have occurred (e.g., Norville et al., 1991; Solomon and Baker, 1994).

Representation of the “bulk” electrical processes in this model, as compared to recent observations of electric fields in lightning-producing clouds (e.g., Marshall and Rust, 1991; Marshall et al., 1995b; Stolzenburg, 1996), indicate that the “break-even” electric field is a reasonable way to determine *where* and *when* the first lightning flash will occur in the modeled cloud. The factor of 1.5 used to increase the value of the break-even field for determining the break down threshold allows for a slight increase in the local electric field strength prior to break down that seem to be present in the observations of Marshall et al. (1995b).

Independent tests of the model using the electrical parameterizations of both T78 and SKM91 suggest that *either parameterization will produce “realistic” results* when the charge transfers are limited to ≤ 250 fC per collision. For the purpose of discussing simulations forced by different large-scale conditions in this study, the T78 parameterization is used. The T78 laboratory results cover a broad region of the temperature-LWC space, agree with large-scale observations (e.g., Williams et al., 1995), and the original T78 laboratory results have recently been replicated by Dr. T. Takahashi (personal communication, Dr. E. Williams, MIT).

5.1.4 Model initialization

The model environment was initialized using composited tropical oceanic, island, and mid-latitude continental sounding profiles discussed in Chapter 3 (e.g., Figs. 5.4 a-d; Fig. 3.12). Initiation of convection in the model was forced by a simulated “gust-front”, represented numerically by a parabolically increasing vertical velocity profile, applied over a specified depth of the sub-cloud layer, for a fixed duration of time. Due to the 1-D nature of the model, the simulations exhibited a strong sensitivity to all three parameters specified in the gust front forcing (depth, duration and magnitude). Therefore, to determine a “control-run” forcing (Table

5.3) for each sounding profile the following 3-step procedure was applied: 1) the depth and duration of the forcing, based on observations and previous experience with the model (personal communication, Dr. Brad Ferrier, NASA-GSFC), were held fixed for several simulations; 2) the vertical velocity forcing was then increased incrementally by 1 m s^{-1} (starting at 1 m s^{-1}) for a series of simulations until a rapid growth phase followed by the development of a deep (e.g., a top of at least 9-10 km) cloud were simulated in the model; and 3) the vertical velocity forcing in step (2) was held constant and the depth and duration of the forcing were varied in a similar fashion. The above procedure, when coupled with the added constraint of available observations enabled the determination of a representative set of forcing parameters for each sounding profile.

Table 5.3: Gust front forcing for control simulations

Run	Duration (s)	W_{force} (m s^{-1})	Depth (m)
NACT	1000	5	400
ACT	1000	4	400
IT	900	4	400
MC	1000	5	1000

5.2 Results

We now summarize the outcome of four separate control simulations (and relevant sensitivity tests; e.g., Tables 5.4a-b), initiated using thermodynamic profiles characteristic of four different large-scale storm environments (Figs. 5.4a-d). The different storm environments are represented by tropical oceanic (Fig. 5.4a, ACT; Fig. 5.4b, NACT), tropical island/continental (Fig. 5.4c, IT), and mid-latitude continental (Fig. 5.4d; MC) sounding profiles

(see Sec. 3.1.5; Fig.3.12). The IT, MC and ACT sounding profiles were all associated with large-scale environments conducive to lightning production. The NACT profile is associated with the presence of below average lightning activity, and is treated as a sounding profile characteristic of deep, but “weakly electrified” tropical oceanic convection. The primary difference between the NACT and ACT profiles occurs in the distribution of moisture, with the NACT profile being dryer than the ACT profile (dewpoints lower by 1-3°C) throughout the depth of the sounding. The temperature distributions of the NACT and ACT soundings were virtually identical. Since the NACT profile represents the “null-case” with regard to lightning activity but represents the majority of oceanic convection, the bulk of the sensitivity studies were run using the NACT profile to test the response of the simulated convection to small changes in the vertical temperature and moisture profiles. Additionally, the control simulations were also run without rime-splintering activated to test the effect of increased ice-concentrations in the -3°C to -8°C temperature range on the electrification of the cloud. In all simulations (except for NACT), removal of the rime-splintering scheme *delayed the electrical break down* of the cloud by approximately 15-20 seconds, but seemed to produce little or no effect on the gross microphysical/kinematic characteristics of the modeled clouds. Because of the small influence of the rime-splintering scheme (as parameterized in *this* model) on the simulated convection in agreement with the finding of Ferrier (1994b), no further discussion of sensitivities run to test the HM74 parameterization will be conducted. To maintain the primary emphasis of this study, the “control” simulations and selected sensitivity tests are discussed *within the context of the four primary questions/topics* posed at the end of Chapter 4. The control simulations are discussed individually with regard to the electrification process, microphysics and dynamics of the modeled convection. Sensitivity studies and heating profiles associated with the control simulations are discussed in a separate section in order to more easily compare and contrast the results.

5.2.1 The ACT simulation

Simulated fields of radar reflectivity, mixing ratios of frozen drops, graupel, and rain along with surface rainfall rate for the ACT simulation (lightning-producing; LP) are shown in Figs. 5.5 a-b. The reflectivities diagnosed in the ACT simulation in Fig. 5.5a resemble observations of LP tropical oceanic convection discussed in Chapter 4. For example, the 36 dBZ reflectivity contour (which we will compare to the 30-40 dBZ *radar volume* mentioned in Chapter 4) extended to an elevation of 8 km (-18°C to -20°C), and remained above the height of the -10°C level (6.8 km) for approximately 12 minutes of the simulation. Local reflectivity maxima of 48-54 dBZ were situated at approximately 2 and 6 km, also similar to the radar observations presented in Chapter 4 (e.g., Fig. 4.5, 4.8a-f, 4.21a-b). The reflectivity maximum situated near 2-3 km (1200-1300s; Fig. 5.5a) was associated with a maximum in rain mixing ratio (3 g kg^{-1}), while that near 6 km (1400 s; -5°C) was associated with FD mixing ratios of 3 g kg^{-1} (Fig. 5.6; mean diameter $O[1 \text{ mm}]$, concentration 1 L^{-1}). Modest graupel mixing ratios were located above the FD maximum (peak graupel mixing ratio 0.9 g kg^{-1}) and also contributed to the depth of the reflectivity core.

The initial precipitation echo in the simulation occurred near 1 km as autoconversion produced rain which subsequently increased through the collection of cloud-water. After cessation of the gust front forcing at $t=1000 \text{ s}$, the maximum in rainwater mixing ratio at 2 km gradually descended to produce a primary “warm-rain coalescence” peak in surface rainfall (46 mm hr^{-1} ; Fig. 5.5b) at $t=1500 \text{ s}$ (15 minutes after formation of the cloud). A pronounced secondary peak of rainfall occurred at $t=2100 \text{ s}$ associated with 1) a collapsing column of rainwater formed during the initiation of a deep downdraft (Fig. 5.7a); and 2) descent and

melting of FD's that originated near the 5-6 km level. Though not as obvious, a weaker peak in the surface rainfall occurred after $t=2400$ s which was associated with the descent and melting of frozen drops originally located above the 7 km level. During this simulation, the graupel and snow categories (Fig. 5.6) contributed little mass to the total rainfall due to their relatively small terminal fall speeds.

The cloud water mixing ratio (Fig. 5.6) peaked at 2.8 g kg^{-1} , with small amounts of cloud-water extending to elevations above 8 km, but only very briefly (≤ 100 s). Initiation of cloud ice occurred at $t=1115$ s at the -7°C level (upper temperature threshold for activation of the ice-nucleation scheme), and after mixing to lower elevations, forced the rapid freezing of raindrops situated between the elevations of 5 and 7 km. Once FD particles were formed, collection of cloud-water rapidly increased with FD's dominating the accretion of cloud-water between the elevations of 5 and 7.5 km. The rime-splintering parameterization activated at 1135 s (riming FD's) and continued for four minutes, providing an integrated splinter concentration to the cloud at temperatures near -5°C of several hundred per liter (peaking at $t=1400$ s). These crystals provided extra contact freezing nuclei for raindrops present in the mixed phase region, which in turn, increased the amount of snow, graupel and FD's (largest mass increase in the FD category). Freezing of raindrops through immersion freezing (probabilistic freezing; Bigg, 1953; LFO, 1983) contributed an order of magnitude less mass to the FD category relative to the other production terms (c.f., LFO).

Ice mixing ratios approached 1.4 g kg^{-1} (Fig. 5.6) between the elevations of 10-15 km. Peak cloud ice number concentrations were on the order of 50-100 ($600\text{-}1000$) L^{-1} and occurred between the -10°C and -25°C (-30°C - cloud-top) temperature levels. The simulated ice-concentrations are similar to those observed in tropical convection by Houze and Churchill (1984) and Gamache (1990), and in hurricane convection by Black and Hallett (1986). The

snow category reached peak mixing ratios of 1.9 g kg^{-1} at an elevation of approximately 9 km due to the freezing of small raindrops, accretion of cloud-water, aggregation of cloud ice (autoconversion), depositional growth of ice and the Bergeron process (no single growth mechanism seemed to dominate the development of snow). Note in Fig. 5.6 that mixed phase conditions existed in the cloud at temperatures colder than -10°C for nearly 1000 s.

Peak updraft speeds (Fig. 5.7a) were approximately 15 m s^{-1} at an elevation of 11.6 km, though updrafts in excess of 12 m s^{-1} extended from 6-14 km as the cloud grew to its maximum depth (15.9 km). A weak downdraft formed at $t=1400 \text{ s}$ in the lowest levels of the domain due to precipitation loading. As the downdraft developed near the surface, vertical velocities in excess of 6 m s^{-1} persisted above the -10°C level for another 10 minutes. During this time, the FD population reached its maximum vertical extent in the mixed phase region of the cloud ($t=1200\text{-}1700\text{s}$). As the updraft weakened in the 5-7 km layer, frozen drops, graupel and snow began to fall, melt, and evaporate into sub-saturated air below 5 km (the majority of the melted ice-mass was associated with melting frozen drops; Fig. 5.6). The resultant contribution to the rain mixing ratio by the melted ice-mass, drove the development of a mid-level downdraft that peaked at 4 m s^{-1} near the 3.5 km level. The peak in downdraft intensity at 3.5 km is evident as a local minimum in the plot of cloud mean mass flux (Fig. 5.7b).

Near the time that FD's (present at the 6.5-7 km level) began to descend, electrical breakdown of the cloud occurred ($t=1560 \text{ s}$, altitude 7 km; cloud-top $\sim 13.5 \text{ km}$). Time series plots of charge development in the simulated cloud are shown in Figs. 5.8 a-e. Very weak charging of the cloud initiated at $t=1150 \text{ s}$ between FD's and cloud ice at the 6 km (-7°C) level, with the most rapid increase in electrification occurring after $t=1400 \text{ s}$ as the updraft extended deeper into the mixed phase region (e.g., Figs. 5.6, 5.7a) of the cloud. The freezing of raindrops, subsequent accretional growth of FD's, graupel and snow, coupled with an increasing cloud ice

concentration provided an environment conducive to charge generation by the NIC mechanism. The maximum mean volume charging rate computed for the 2400 meter deep negative space charge region ($2.4 \times 10^{-11} \text{ C m}^3 \text{ s}^{-1}$; Fig. 5.8a) occurred just prior to electrical break down. If the charging rate is assumed to balance the charge neutralized in a lightning flash (ranging from 1-100 C; Uman, 1987), a crude estimate of the flash rate during this period of electrification would be approximately $0.1 \text{ to } 10 \text{ min}^{-1}$.

Figure 5.8a indicates that the resulting *net* charge density in the cloud was distributed in a tripolar structure with a region of peak negative charge densities of -3 nC m^{-3} (-14°C to -18°C), sandwiched between two layers of positive charge centered at -5°C and -35°C ($+1 \text{ nC m}^{-3}$ and $+2 \text{ nC m}^{-3}$ respectively). The vertical electric field at break down (Fig. 5.9) was approximately 110 kV m^{-1} at an elevation of 7 km and -90 kV m^{-1} at 9 km. It is important to note that more complex charge distributions are often observed in nature (e.g., Marshall and Rust, 1991; Marshall et al., 1995b; Stolzenburg, 1996), however the electrical parameterization used in this one-dimensional model does not include inductive charging between cloud-water and ice (the degree to which this mechanism contributes to overall cloud-electrification is currently in dispute), ion transports, or screening layers (e.g., Brown et al., 1971).

The tripolar structure of the cloud in Figs. 5.8a-e resulted from the superposition of charge densities residing on each category of ice. However, Figs. 5.8a-e also suggest that the primary charge carriers in the cloud were the FD (Fig. 5.8b) and cloud ice categories (Fig. 5.8e). Graupel (Fig. 5.8c) carried net negative charge throughout the depth of the cold region of the cloud, possessing a local maximum in charge density of -2 nC m^{-3} at the 9.4 km level. When superimposed on the FD and ice charge densities, the graupel charge density reinforced the negative region of the tripole while decreasing the magnitude of the net positive charge in the upper region of the tripole. Positive charge transfer between the FD's, snow and ice, and graupel

(graupel acquiring positive charge) at temperatures close to -10°C , coupled with negative charge transfers to graupel aloft in collisions with the ice and snow, resulted in a *net* charge distribution for graupel of a single polarity (negative). Conversions of snow to graupel via various microphysical processes (e.g., PGAUT, SACW, PSACR etc.) also played a large role in the modulation of the charge density on graupel, because snow almost always acquired a sign of charge opposite to that of graupel in the NIC parameterization. When snow mass was shifted to the graupel category, net charge residing on the snow (opposite that from graupel) was also placed in the graupel category.

In summary, the ACT simulation produced a deep convective cell with 36 dBZ reflectivities and significant FD mixing ratios that extended well into the mixed phase region of the cloud (8.2 km). Simulated electric fields were large enough to cause break down, thus this cell should have been capable of producing lightning discharges at a modest rate. Vertical velocities in excess of 6 m s^{-1} suspended moderate concentrations of FD's above the height of the -10°C level, where they actively participated in the electrification of the cloud through the NIC mechanism. As the FD mass began to separate from the snow and ice-mass on the cloud-scale (e.g., Fig. 5.8a-e), the cloud developed a well defined tripolar structure and electrical break down occurred. Over the duration of the model run, approximately 34% of the column integrated rainfall originated from melted ice-mass (primarily FD), with the remaining 66% contributed by warm-rain processes.

5.2.2 The NACT simulation

The NACT sounding profile (Fig. 5.4b) required a 1 ms^{-1} increase in vertical velocity (Table 5.3) forcing for the developing convection to overcome a slightly increased value of CIN,

and dryer air at low-levels (relative to the ACT profile). With the appropriate forcing in place, the NACT simulation produced a convective cloud with a top of approximately 10 km and a peak surface rainfall rate of 40 mm hr^{-1} . The reflectivity structure of the simulated NACT cloud (Fig. 5.10a) appeared consistent with observations of deep, but non-lightning producing (NLP) convection observed during TOGA COARE (e.g., Fig. 4.11; Table 4.1). The 36 dBZ core was confined to elevations of less than 6 km and a primary peak in reflectivity occurred near the 2 km level, associated with a warm-rain coalescence maximum (peak rain-mixing ratio of 2.7 g kg^{-1}). When the gust-front forcing ceased at $t=1000 \text{ s}$, the rain mass situated at 2 km continued to accrete more cloud-water and began to fall to the surface, producing an initial peak in rainfall (Fig. 5.10b) at $t=1500 \text{ s}$. As the updraft continued to grow in the vertical, rain and cloud water were lifted to the 0°C level, ice was initiated (1150 s), and FD, snow and graupel mixing ratios all rapidly increased near the 5 km level (0°C to -5°C ; Fig. 5.11).

It is interesting to note that the peak graupel and snow mixing ratios in the NACT simulation (Fig. 5.11) were larger (peak graupel mixing ratio 1.4 g kg^{-1}), than that of the ACT simulation (peak graupel mixing ratio 0.9 g kg^{-1}). This can be attributed directly to the weaker vertical velocity and mass-flux structure of the NACT cloud (peak upward vertical velocity of 9.6 m s^{-1} at the 4.6 km level; Figs. 5.12a-b) which was unable to loft rain and generate cloud-water via condensation at the same rate, or to the same temperature level, as the updraft in the ACT cloud. Consequently, a large portion of the rainwater fell out of the NACT cloud near $t=1500 \text{ s}$ prior to the updraft reaching the 0°C level, leaving only a limited amount of rainwater available for production of higher density, faster falling FD's. Concentrations of FD's that did form near the freezing level, rapidly fell out of the cloud due to their greater terminal fallspeeds and their location within a weak downdraft at mid-levels of the cloud (Fig. 5.12a) (note the very weak secondary peak in surface rainfall at 2250 s). The removal of the FD's from the weaker

NACT updraft made relatively more cloud-water and small amounts of rainwater available for the accretional/freezing growth of snow and small graupel, thereby increasing the mass of those two categories relative to the FD mass. Hence the snow and graupel mixing ratios increased to a maximum, coincident with the complete removal of cloud and rain water (e.g., Fig. 5.11) from the updraft at temperatures colder than 0°C . Subsequently, the NIC mechanism shut down (Figs. 5.13a-e) due to a lack of cloud-water (associated with lower condensation rates in the weak updraft) which prevented the cloud from reaching break down electric fields.

In marked contrast to the ACT simulation, the electrical development of the NACT cloud was dominated by graupel and snow interactions (Figs. 5.13c-d). Due to the low liquid water contents present in the cloud between $t=1600$ and 2000 s, the graupel charged positively in collisions with snow and cloud ice, and as indicated in Figs. 5.13a and 5.13c, began to fall out of the cloud around 2000 s. Interestingly, once charges of opposite polarity were separated in collisions between graupel and snow (Figs., 5.13c-d), the regions of opposite charge on the graupel and snow centered near the 7.4 km level virtually canceled algebraically, resulting in only a small net charge density of -1 nC m^{-3} . This cancellation was due in part to the relatively small difference ($1\text{-}3 \text{ m s}^{-1}$) in terminal fall speeds between the graupel and snow. The “charge-cancellation” is similar to that noted by Randell et al. (1994) in a two-dimensional simulation of tropical monsoon convection. Since the FD mixing ratios were much smaller above the -10°C level, the FD category did not exert a large effect on the overall electrical development of the cloud. The resultant charge structure (Fig. 5.13a) approached that of a tripole, but was slightly more complex (especially at $t=2000$ s) with charge concentrations of -1 nC m^{-3} and $+1 \text{ nC m}^{-3}$ situated at the 7.4 and $5\text{-}6.5$ km levels respectively. The peak electric field (not shown) during the simulation reached a value of 56 kV m^{-1} at the 7.4 km level at $t=1900$ s.

The relatively weak electrification observed in the NACT simulation is consistent with observations of weakly electrified convection over the warm-pool and of similar convection in the Australian summer-monsoon (e.g., Rutledge et al., 1992). Even though the cloud-top in the NACT simulation was only 10 km, it is incorrect to assert that the lack of lightning in the NACT cloud, or in similar clouds in nature, is due to the low cloud-top. Visual observations from TOGA COARE suggest that clouds with tops as low as 10 km occasionally produced lightning flashes (personal communication, C. DeMott, CSU). Conversely, observations presented in Fig. 4.11, and 4.13 (and elsewhere in the tropics; e.g., Williams et al., 1992) suggest that warm-pool convection can be much deeper than that of the NACT simulation while still not producing much, if any, lightning. One question that does arise with regard to the NACT simulation is; under what circumstances will the 1-D model simulate deep, NLP or marginal lightning-producing convection (MLP; defined at the bottom of Table 5.4a) with tops approaching say, 15 km, as in Fig. 4.11? Secondly, can the NACT sounding parameters be modified in a physically realistic way such that an LP cloud occurs (e.g., can a 10 km deep cloud that produces a lightning flash be simulated)?

5.2.3 Sensitivity studies using modified NACT and ACT sounding profiles

To address questions regarding the sensitivity of modeled convective/electrical development to changes in the thermodynamic structure of the NACT and ACT soundings, the sounding profiles (primarily the NACT) were modified in several different ways and sensitivity tests were conducted using the same gust front forcing as the control cases. Tables 5.4a-b summarize the primary sensitivity tests together with the ACT and NACT control simulations for comparison.

Table 5.4a: Description of sensitivity runs

Run	Description	CAPE J kg ⁻¹	MLMR g kg ⁻¹
ACT	COARE Lightning producing environ. Control run.	2400	18.5
NACT	COARE Non-lightning producing. Control run.	1750	17.8
ACT2	COARE ACT sounding forced with 5 ms ⁻¹ as in NACT vs. 4 ms ⁻¹ in control run.	2400	18.5
ACT3	COARE ACT sounding warmed by 0.5°C from 660 mb to the Tropopause. Same forcing as ACT.	2200	18.5
NACT2 _A	NACT sounding with PBL dewpoint increased 1°C. Forcing same as ACT.	2500	18.8
NACT2 _B	Same as NACT2; forcing same as NACT	2500	18.8
NACT2 _C	NACT sounding with PBL dewpoint increased 0.5°C; forcing same as NACT	2050	18.3
NACT3	NACT sounding with same moisture profile as ACT from 950 - 700 mb (PBL/forcing same as NACT)	1800	17.8
NACT4	NACT sounding cooled 0.5°C uniformly from 660 mb to tropopause. (PBL/forcing same as NACT)	1900	17.8
NACT5	NACT sounding cooled 1°C uniformly from 600 mb to tropopause- graduated cooling from 660-600 mb. (PBL/forcing same as NACT)	2100	17.8
NACT6	NACT sounding warmed uniformly by 0.5°C in PBL. Same forcing as NACT.	1800	17.8
NACT7	NACT sounding run with 6 ms ⁻¹ updraft forcing.	1750	17.8

Table 5.4b summarizes for each sensitivity test simulation shown in Table 5.4a; the cloud top height, the maximum height of the 36 dBZ contour, the maximum vertical velocity and the height at which it occurred, the time of electrical breakdown, the peak charging rate, the primary and secondary peak rain rates, the domain maximum graupel mixing ratio, and the domain maximum FD mixing ratio along with the height at which it occurred.

Examining Table 5.4b, it is clear that shallow (e.g., cloud-top ≤ 10 km) LP clouds could not be simulated by the model. However, deep MLP storms with tops ranging from 13 to 15 km (see definition of “marginal” at the bottom of Table 5.4a) were simulated. In order to simulate MLP convection, the NACT sounding was destabilized by either cooling levels aloft, or by Table

5.4b: Summary of sensitivity test results

Run	Top km	36 dBZ Ht. km / °C	w_{max} / Ht. m s ⁻¹ / km	B.-dwn sec.	dQ/dt _{pk} nC m ⁻³ s ⁻¹ (-)	Rain Pk1/Pk2 mm hr ⁻¹	Pk q _{gr} g kg ⁻¹	Pk q _{ra} /Ht. g kg ⁻¹ /km
ACT	15.9	8.2/-19	15.0/11.6	1560	2.4×10^{-11}	47/29	0.9	3.6/5.7
NACT	9.8	6.0/-6	9.6/4.6	-	9.4×10^{-12}	40/9	1.4	1.6/5.3
ACT2	18.4	11.8/-49	36.2/15.8	1222	5.3×10^{-11}	92/65	0.7	8.0/6.1
ACT3	15.0	7.2/-13	12.3/7.2	1733*	1.1×10^{-11}	46/26	0.9	3.1/5.7
NACT2 _A	15.9	8/-17	14.5/6.6	1552	2.0×10^{-11}	57/29	0.7	3.8/5.7
NACT2 _B	18.7	12/-50	38.7/14.8	1201	5.3×10^{-11}	104/73	0.7	9.0/6.1
NACT2 _C	17.1	9.4/-26	23.7/13.0	1337	3.7×10^{-11}	77/39	0.6	5.5/5.7
NACT3	14.7	7.2/-13	12.7/6.4	1680*	1.6×10^{-11}	56/23	0.9	3.1/5.7
NACT4	13.0	6.5/-9	10.7/4.8	-	4.0×10^{-12}	39/10	1.2	1.9/5.3
NACT5	13.7	7/-12	11.2/5.8	1786*	1.1×10^{-11}	39/10	1.2	1.9/5.5
NACT6	16.1	8.3/-19	15.5/6.6	1470	3.6×10^{-11}	68/25	0.7	4.0/5.7
NACT7	17.6	10/-33	27.1/14.6	1277	4.0×10^{-11}	97/42	0.6	6.2/5.9

* Electrical breakdown occurred in these clouds just prior to the end of active non-inductive charging, the dissipation of cloud water, and the descent of FD mass out of the mixed phase region. For this reason (and the fact that the peak volume charging rates were so low), these simulations are considered to be "marginal" lightning producers.

moistening the sounding at low levels (e.g., Table 5.4a). The amount of cooling or moistening applied to the NACT sounding was determined in part by the standard deviations in the NACT dewpoint and temperature profiles. The standard deviation of the dewpoint (temperature) in the NACT PBL composite was approximately 1°C (0.6°C), while the standard deviations in dewpoint and temperature aloft (e.g., pressure less than 950 mb) were 2°C and 0.6°C respectively. For the ACT sounding, the standard deviations in PBL dewpoint and temperature were 0.5°C and 0.4°C, while the deviations aloft were 2°C and 0.6°C respectively. From this analysis, it was concluded that a 0.5-1°C cooling (or warming) of the column from 660 mb to the tropopause represented a *strong* perturbation to the sounding profile. Similarly, a 0.5-1°C moistening of the boundary layer is considered to be a relatively large perturbation in the composite dewpoint.

The sensitivity tests NACT4 and NACT5 (Table 5.4b) suggest that strong negative temperature perturbations aloft produced a deep model cloud (a 3-4 km increase in cloud-depth), but little if any lightning. Conversely, if the ACT sounding was warmed by 0.5°C at mid-levels and above (approximately one σ), the maximum mean volume charging rate in the main negative charge region of the storm was significantly reduced, and the 15 km-deep cloud was classified as MLP. Note that the mean volume charging rates of the NACT4-5 and ACT3 sensitivity tests were a factor of two lower than that considered necessary for significant lightning production (e.g., Latham and Mason, 1961). When the PBL temperature of the NACT sounding was increased uniformly by one standard deviation ($\sim 0.5^\circ\text{C}$; a relatively large increase), the cloud became a lightning producer with overall characteristics similar to that of the ACT simulation. However even with increased PBL temperatures in NACT6, the simulation still required stronger forcing than that of the ACT simulation.

In contrast, when the PBL dewpoint of the NACT sounding was increased one standard deviation ($\sim 1^\circ\text{C}$; NACT2_{AB}, a marked increase in CAPE and MLMR, Table 5.4a), the cloud electrical intensity and depth increased dramatically. When the slightly stronger NACT forcing was used, peak updrafts of 38.7 m s^{-1} and cloud-tops in excess of 18 km resulted. Indeed, even when relatively weaker surface forcing was applied (NACT2_B; Table 5.4a) the model produced a highly electrified cloud with a structure very similar to that of the ACT simulation. Increasing the NACT PBL dewpoint by only 0.5°C still provided sufficient instability to the NACT profile for the model to produce deep, highly electrified convection. As a final test of the model sensitivity to moisture changes in the NACT sounding, the mid-level dewpoints were increased from $2\text{-}4^\circ\text{C}$ to match that of the ACT sounding. The resultant model-cloud was deep (14.7 km top), but only a marginal lightning-producer.

The sensitivity tests in Tables 5.4a-b also highlight the extreme sensitivity of the model convection to the applied forcing. For example, consider the ACT and ACT2 simulations. The ACT2 simulation was run using precisely the same thermodynamic structure, but with 1 m s^{-1} greater vertical velocity forcing in the gust front (e.g., same forcing as the NACT simulation). The resultant model cloud exhibited a marked response to the increase in forcing as evidenced by the approximate doubling of the ACT2 maximum updraft speed, 3 km increase in cloud top, and doubling of the maximum mean volume charging rate in the negative charge region. Now recall that the NACT forcing was required in the NACT simulation to produce a cloud that was *only* 10 km deep, and which produced no lightning. Clearly, the cloud-model suggests that a strong sensitivity exists between convective/electrical intensity and the thermodynamic structure of the cloud-environment.

Further, the simulations suggest that the modeled clouds responded sensitively to changes in the PBL thermodynamic structure, and in particular, to the PBL moisture content. Similar increases in PBL moisture and temperature (e.g., standard deviation from the mean) produced deep, highly electrified convection (e.g., NACT2_{A,B}, NACT6). Yet with *relatively* similar forcing, a one-standard deviation increase in the PBL moisture content produced a much deeper, more electrically active cloud (e.g., a 50% increase in charging rate; Table 5.4b) in the NACT2_A simulation, relative to the one- σ temperature forcing in the NACT6 simulation. If the forcing was reduced, in the case of NACT6 a deep, electrically active cloud was not produced. Conversely, when the forcing was reduced for the NACT2_B simulation, a deep, lightning producing convective cloud similar to that of the ACT simulation was produced. Finally, when the PBL moisture content was increased by only 0.5σ , and the same forcing used as in NACT6, a deep, lightning producing cloud was simulated by the model. The apparent increased

sensitivity of the model convection to PBL moisture content is consistent with previously noted sensitivities between CAPE, MLMR, θ_w , and CG lightning flash density (e.g., Chapter 3).

In summary of Table 5.4b, the MLP clouds had the following characteristics: 1) a mean maximum frozen drop mixing ratio of 2.3 g kg^{-1} , situated at a mean height of 5.5 km; 2) a mean maximum graupel mixing ratio of 1.1 g kg^{-1} ; 3) a mean maximum 36 dBZ height of 6.8 km (-10.4°C); and 4) a mean primary to secondary rainfall peak ratio of 0.34 with a mean primary rainfall peak of 44 mm hr^{-1} and a secondary peak of 15.6 mm hr^{-1} . The lightning producing clouds exhibited 1) a mean maximum FD mixing ratio of 5.7 g kg^{-1} situated at a mean height of 5.8 km; 2) a mean maximum graupel mixing ratio of 0.7 g kg^{-1} ; 3) a mean maximum 36 dBZ height of 9.6 km (-30°C); and 4) a mean primary to secondary rainfall peak ratio of 0.55 with a mean primary peak in rainfall of 77 mm hr^{-1} , and secondary peak of 43 mm hr^{-1} . In the MLP runs, the decrease in FD production permitted more rainwater and cloud-water to be frozen and/or accreted into the graupel category, thereby increasing the graupel mixing ratio. The reverse occurred in the lightning-producing cases where much more of the rain and cloud-water mass was collected by the higher density FD category. In general, the snow and ice mixing ratios in the deep lightning producing clouds with peak updraft intensities of $15\text{--}25 \text{ m s}^{-1}$ are comparable to those produced in the MLP clouds (peak snow plus ice mixing ratios of $2.6\text{--}2.9 \text{ g kg}^{-1}$). The much deeper LP storms with peak updraft speeds in excess of 30 m s^{-1} (rarely if ever observed over the tropical ocean; e.g., Zipser and LeMone, 1980; LeMone and Zipser, 1980; Lucas et al., 1994) had peak snow plus ice mixing ratios of $3.4\text{--}4.0 \text{ g kg}^{-1}$. The largest difference in the microphysical characteristics of the MLP and LP clouds seemed to be the mixing ratio reversal in the rimed ice-category, with LP (MLP) clouds containing substantially higher mixing ratios of FD's (graupel) in the mixed phase region of the cloud. A larger concentration of high density, faster falling precipitation ice provided a greater charge transfer in collisions between

ice/snow and precipitation ice, and also produced a more efficient cloud-scale separation of the charge based on the marked difference in terminal fall-speed characteristics between the FD species and the other ice categories. Further, the increase in FD mass between the 0°C to -10°C temperature levels provides for a larger secondary peak in rainfall because the FD's remove copious amounts of cloud-water through riming prior to rapidly falling to the surface. Indeed, the FD category dominates the collection of cloud-water relative to the graupel and snow categories due to a relatively larger terminal fall speed. Hence the LP storms produce a larger, broader secondary peak in rainfall that is related to the production of FD's above the 0°C level.

5.2.4 The tropical island simulation

For comparison to the tropical oceanic simulations, we also initialized the cloud-model with a sounding profile characteristic of the tropical island/continental environment of the Maritime Continent region of northern Australia (IT; Fig. 5.4c). The IT sounding was forced using a gust front depth (400 m) and vertical velocity (4 ms^{-1}) similar to the ACT simulation, however the duration of forcing was reduced from 1000 s to 900 seconds. The reduction in the duration of the forcing (relative to the ACT simulation) was necessary to confine the depth of the IT convection to the vertical bounds of the model domain. Wind-profiler observations collected over the islands of Melville and Bathurst (northwest of Darwin) suggest that the depth and magnitude of the forcing used in the model for this type of convection are appropriate (personal communication, Dr. P. May, BMRC).

The radar reflectivity, mixing ratios of FD, graupel, and rain (Fig. 5.14a), and the surface rainfall rate (Fig. 5.14b) all indicate that the simulated IT cell was much more intense (electrically and kinematically) than either the ACT or NACT control simulations. Note that 36

dBZ tops extended to the 11.6 km level, peak FD mixing ratios reached 10 g kg^{-1} and peak graupel mixing ratios approached 1 g kg^{-1} (Fig 5.16). For comparison to the IT model simulation, Fig. 5.15 shows a vertical cross-section of radar reflectivity through a deep island thunderstorm observed on 28 November 1995 during the Maritime Continental Thunderstorm Experiment (MCTEX, Keenan et al., 1996; cross-section courtesy Mr. L. Carey, CSU). Note that MCTEX convection exhibited double peaks in radar reflectivity (48-54 dBZ) situated in the 2-3 km and 6-7 km levels respectively. Reflectivities in excess of 36 dBZ extended to an elevation of 11.5 km, and echo-tops approached 19 km (Fig. 5.15). Peak flash rates in the ensemble of convective clouds containing this cell were approximately 50 min^{-1} (as measured by a flat-plate antenna; Carey et al., 1997). Polarimetric radar retrievals of precipitation ice-mass (assumed to be frozen drops and graupel) suggest that relative maxima in the precipitation ice-mass were typically located between the 6 and 8 km levels (e.g., Carey et al., 1997) in this type of convection. Hence, the IT simulation appears to have captured some of the gross structural aspects of the island convection in a reasonable fashion (c.f., Keenan et al., 1994).

In the simulated island convection, cloud-water mixing ratios in excess of 1 g kg^{-1} extended to the 10 km level, with peak cloud-water mixing ratios of 3.5 g kg^{-1} extending to near the -10°C level (Fig. 5.16). The depth and magnitude of the cloud-water mixing ratios provided a robust mixed phase environment for the riming growth of FD's and graupel. Peak snow mixing ratios developed in the cloud after the initial pulse of the updraft (Fig. 5.17a-b), and were associated primarily with conversions of cloud ice to snow through autoconversion, the Bergeron process, and growth of snow through vapor deposition (Fig. 5.16). Cloud ice developed rapidly, peaking during the most intense phase of growth, with much of the ice-mass being converted to the snow category at later times (Fig. 5.16). The rime-splintering mechanism played a relatively minor role in the production of ice-crystals as it was activated for only 100 s near $t=930$. Rime-

splintering processes produced a maximum ice number concentration of 1 L^{-1} near the -5°C level prior to being deactivated due to the presence of a strong updraft, which shifted the near-balance level for FD and graupel categories to temperatures colder than -8°C . Primary nucleation of ice crystals in the deep water saturated environment provided the primary source for cloud ice concentrations in the convection.

Interestingly, the relative magnitudes of the primary and secondary rainfall peaks (Fig. 5.14b) were juxtaposed when compared to the ACT (Fig. 5.5b) and NACT (Fig. 5.10b) simulations (and also *all* of the NACT/ACT sensitivity tests). In the IT simulation, the lower of the two rain-rate peaks (100 mm hr^{-1}) was associated with the warm-phase of the rainfall process, while the larger secondary peak (125 mm hr^{-1}) was associated with fallout of both rainwater and melted ice-water from the 5-7 km levels (as in the ACT case). The secondary peak in rain rate also had a pronounced “tail” with peak rain-rates that approaching 25 mm hr^{-1} . These rainfall rates were associated with the fallout of precipitation ice originally situated at elevations in excess of 6-7 km in the mixed phase region of the storm. To ensure that the juxtaposition of the rainfall peaks was not just an artifact of the different autoconversion threshold used for conversion of cloud to rainwater in the IT run (e.g., the IT threshold was twice that of the ACT threshold based on speculation that the CCN spectra over the island might resemble that of a mixed marine-continental climate), the IT simulation was re-run with the ACT autoconversion threshold; no change in the results occurred. Approximately 42% of the IT surface precipitation resulted from melted ice-mass compared to 34% and 21% in the ACT and NACT simulations respectively.

The vertical velocity and mass flux structure of the IT simulation (Figs. 5.17a-b) indicate the presence of a deep updraft in the IT cloud. Peak vertical velocities exceeded 30 m s^{-1} in regions of the cloud above 15 km (consistent with observations taken during ITEX; e.g.,

Simpson et al., 1993; Keenan et al., 1994), while vertical velocities exceeded 10 ms^{-1} from a height of approximately 2.5 km to cloud-top. Early in the simulation, the developing updraft lofted substantial quantities of rainwater to temperature regions colder than -10°C (Fig. 5.16), where it was rapidly frozen and accreted, shifting substantial water mass into the FD/H, graupel and snow categories. The presence of rainwater deep in the mixed phase region of IT-type clouds is supported by videosonde observations and multiparameter radar observations collected in deep island convection during the Maritime Continental Thunderstorm Experiment (MCTEX). These observations suggest the presence of millimeter sized raindrops at temperatures as cold as -20°C (personal communication, Dr. T. Takahashi, Kyushu University; Carey et al., 1997). Later in the simulation, a low to mid-level downdraft (peak of $\sim 5 \text{ m s}^{-1}$) developed in the storm at $t=1600$ seconds as a result of substantial precipitation loading. In turn, the downdraft increased the rainwater mass-flux at elevations below 0°C which resulted in melting ice-mass rapidly falling to the surface, forming the primary peak in rainfall at $t=2000$ seconds. The development of the downdraft markedly reduced the mean upward mass flux of the cloud near the 3 km level (Fig. 5.17b). In general, the simulated vertical velocities of the IT run are in good agreement with previous observations (e.g., Keenan et al., 1994; Carey et al., 1997) and other cloud-modeling studies of “Hector” summarized in Simpson et al. (1993) and Keenan et al. (1994).

The magnitude and depth of the vertical velocities in the IT run were ultimately responsible for the large quantities of rainwater injected into the mixed phase region of the model cloud (Fig. 5.16), which subsequently froze to become frozen drops. In turn, the development of substantial precipitation-ice mass coupled with the rapid production of cloud ice near $t=1200$ seconds (Fig. 5.16), enabled the NIC mechanism to rapidly charge the cloud, producing electrical break down in the simulation at $t=1220$ (9.5 km level; Figs. 5.18a-e). The charge and electric field structure (Figs. 5.18a-e; Fig. 5.19) of the cloud resembled that of a classic tripole (e.g., Williams, 1989),

with a region of negative charge (peak -2.5 nC m^{-3}) situated near the -15°C to -20°C level of the storm, and two layers of positive charge located above and below the negative charge region (peaks of $+2.1$ and $+1.6 \text{ nC m}^{-3}$ respectively). The maximum mean volume charging rate ($4.2 \text{ nC m}^{-3} \text{ s}^{-1}$) in the negative charge region occurred just prior to electrical break down. If the flash-rate scaling approximation used in the ACT simulation is applied to the IT charging rate, the IT negative charging rate could sustain a total flash rate of approximately 0.2-20 flashes per minute (similar to total flash rates observed using a flat-plate antenna during MCTEX; personal communication, Mr. L. Carey, CSU).

Similar to the ACT simulation, the primary charge carriers in the IT run were the FD and cloud ice categories (Figs. 5.18a-e). Further (as in the case of the ACT simulation), the IT cloud would not achieve electrical break down if the FD's were not permitted to participate in the NIC process. Interestingly, the entire charge structure of the IT cloud was elevated approximately 500-1000 m with respect to the charge structure of the ACT simulation (Fig. 5.5a), even though the height of the IT cloud-top at the time of break down was the same as the ACT simulation ($\sim 12 \text{ km}$). For comparison to deep mid-latitude convection, the somewhat elevated tripole structure of the IT updraft is consistent with both observations (e.g., Marshall et al., 1995b; Stolzenburgh, 1996) and modeling simulations (e.g., Ziegler and MacGorman, 1994) of mid-latitude supercell convection, which suggest that the main dipole region in strong *updrafts* may be elevated with respect to that observed in "ordinary" thunderstorms. In terms of lightning type, the elevated charge structure may offer a partial explanation for the increased IC:CG ratio previously noted for the deepest, most electrified convection observed over the continental region of the northern Australia (e.g., Rutledge et al., 1992).

5.2.5 The mid-latitude continental simulation

In order to provide an extreme contrast in convective regime type, the 1-D model was used to simulate mid-latitude convection (MC). Herein we provide a brief overview of the MC simulation, recognizing that the convective environment associated with mid-latitude continental convection is often highly sheared, and hence the full dynamical effect of the environment will not be realized in the MC simulation. We must also add that the microphysical category "FD" becomes somewhat ambiguous with regard to the exact particle type being referred to (i.e., the category could represent hail or frozen raindrops in mid-latitude convection), hence we use the term "FD" to emphasize the presence of high density ice. The MC simulation is probably more characteristic of cumulonimbus convection in the southern plains of the U.S. (e.g., Koenig, 1963) given the microphysical parameterization. For example, the freezing of raindrops is *not* a primary mode by which precipitation sized ice particles are formed in deep cumulonimbi characteristic of the northern tier of the U.S.; the development of precipitation ice in those storms is strongly linked to the accretional growth of graupel (e.g., Knight et al., 1974; Dye et al., 1986). The primary goal of the MC simulation is to provide an example of the gross structural, electrical and microphysical differences that exist between MC and tropical convective scale "processing" as a function of the thermodynamic structure of the troposphere (e.g., Fig. 1.1).

To better simulate the mid-latitude precipitation process (and enhanced CCN concentrations in mid-latitude continental air masses), the autoconversion threshold for cloud-water to rain was increased a factor of four from the NACT simulation (Table 5.1). Additionally, to achieve a more realistic simulation of the cloud (and surface rain rates), it was necessary to reduce the intercept (N_0) of the graupel size distribution by a factor of 5 to 10 relative to the tropical simulations. This resulted in a graupel size distribution for graupel

similar to that used in LFO and modeled in F94. The intercept of the FD distribution was kept at the same value as the IC and ACT simulations (F94).

The sounding used to initialize the MC model simulation is shown in Fig. 5.4d. The sounding possessed a CAPE similar to the ACT and IT soundings (e.g., Sec. 3.1.5) with an MLMR of 14.3 g kg^{-1} and a CIN of $\sim 15 \text{ J kg}^{-1}$. The LFC of the MC sounding was approximately 1.5 km above the surface which meant the depth of the model forcing had to be increased to initiate a deep convective cell. This was achieved by applying a gust front forcing of 5 m s^{-1} over a depth of 1000 meters for a duration of 1000 seconds. This gust front forcing is similar to that produced in squall-line simulations such as Fovell and Ogurra (1988, 1989).

A plot of the simulated radar reflectivity, FD, graupel and rain mixing ratios, together with surface rainfall rate is shown in Figs. 5.20a-b. In contrast to the ACT, NACT and IT simulations, the first rainwater associated with the mid-level cloud developed approximately 700 m below the height of the 0°C level (3.9 km) at a time of 850 seconds. This rainwater was immediately lofted into the mixed phase region of the cloud to a height of approximately 5.6 km (-11°C) where it was frozen and/or collected to become either FD's, graupel or snow. A small amount of the rainwater also fell out of the cloud prior to being frozen, however, after time $t=1200$ seconds, the majority of the precipitation formation was associated with an ice process (accretional growth of high density frozen drops, graupel and snow). Peak FD (high density ice) and graupel mixing ratios were approximately 4 g kg^{-1} and 1 g kg^{-1} respectively. The maximum updraft speed in the MC cloud was 27 m s^{-1} at an elevation of 12 km. Updraft speeds exceeding 10 m s^{-1} (associated with the greater low level instability, e.g., the B_1 profile) extended from a height of 4 km ($\sim 0^\circ\text{C}$) to just below the cloud top (14 km).

The primary and secondary rainfall peaks (45 mm hr^{-1} and 12 mm hr^{-1} respectively) in the MC simulation (Fig. 5.19b) were *both* associated with the melting and fallout of ice-mass. The

primary peak in rainfall occurred after the gust front forcing ceased and was associated with melted FD mass originally situated in the -5°C to -10°C levels. The second peak in the rainfall was associated with ice-mass that descended from above the -10°C level between times $t=1400$ -1500 seconds. Of the total column integrated rainwater, approximately 75% of the mass originated from melted ice in the MC simulation. This percentage for the MC simulation can be contrasted to that of the tropical ACT (34%) and IT (42%) simulations. Clearly, the ice-process dominated production of rainfall in the MC simulation (consistent with observations; e.g., Koenig, 1963; Knight et al., 1974; Young, 1993).

The development of electrical charge in the MC simulation (Fig. 5.21a-e) was rapid, with electrical break down of the cloud occurring 400 seconds (1220 s) after the first radar echo appeared. The NIC mechanism operated very efficiently in the cloud due to the dominant mixed-phase character of the precipitation process. For example, the NIC mechanism initiated between the FD and ice categories (Figs. 5.21b and 5.21e) at the -6°C temperature level (~ 5 km) only 100 seconds after the first precipitation echo appeared near the 0°C level in the simulation. That is, a considerably larger percentage of the water condensed in the updraft of the MC simulation was available to the ice-phase, and hence conditions were more conducive to electrifying the cloud via the NIC mechanism. In the tropical simulations, the clouds did not electrify until the updraft lofted a substantial amount of rain and cloud-water from elevations as low as the 1 km level through the height of the -6°C level, which took approximately 450 seconds in the IT run and nearly 700 seconds in the NACT run. Hence, when the results of the NACT, ACT, IT, and simulations are considered as a whole, the importance of a robust precipitation ice-process in the electrification of the model clouds is demonstrated in addition to the gradual shift in precipitation process (water to ice-based) that occurs as a function of convective rainfall regime and

associated degrees of electrification (e.g., weakly electrified tropical oceanic, highly electrified tropical oceanic, island/continental, and mid-latitude continental convection).

5.3 Convective heating

Modeling studies such as Hartman et al. (1984), Puri (1987), Lau and Peng (1987), and Sui and Lau (1989) have demonstrated the sensitivity of large-scale tropical atmospheric circulations to sub-grid scale vertical distributions of diabatic heating associated with deep convection. Houze (1989) found that the mesoscale vertical structure of diabatic heating in *stratiform* precipitation was very similar from regime to regime. However, significant variations in the structure of the *convective* heating profile have been noted in studies of convective regimes such as Johnson (1984), Houze (1989), Cifelli (1995), and (DeMott, 1996). Since the total sub-grid scale heating associated with convective systems consists of both mesoscale (stratiform) and convective heating components, and the mesoscale component appears to exhibit the least variation (e.g., Houze, 1989), deviation of the total diabatic heating profile (ignoring radiative contributions) will be primarily a function of the convective heating profile.

The observations presented in Chapter 4 coupled with the modeling simulations discussed herein, suggest that NLP tropical convection differs both kinematically and microphysically from LP tropical convection. Therefore, it is also likely that the convective heating profiles associated with each type of convection (i.e., LP, NLP etc.) differ substantially. Two of the many diagnostics provided by the 1-D model and used in this study are the accumulated diabatic heating and drying profiles. These profiles can be interpreted for each simulation with the caveats that one-dimensional simulations of convection (an inherently three-dimensional phenomena) using a prescribed forcing are not likely to be representative of the entire ensemble

of convective clouds occurring in a given area. For example the 1-D model can only represent vertically erect convection and as a result downdraft intensities (evaporative cooling) are likely to be underestimated at low levels; the model is sensitive to the geometry chosen to represent the cloud (e.g., FH89); and the modeled convection is very sensitive to the prescribed “gust-front” forcing. Accepting the weaknesses of the model, we now turn to a discussion of the significant differences noted in diabatic heating profiles (with less emphasis placed on discussion of the moistening/drying profiles) between NLP, MLP and LP tropical convection.

Heating and drying profiles (normalized to a one-hour rainfall in each simulation) for the NACT, NACT3, NACT4, ACT and IT simulations are shown in Figs. 5.22a-b. Recall from Table 5.4b that the NACT3 simulation was an NLP convective cell (but deeper than the NACT cloud), and the NACT4 simulation was an MLP convective cloud. Hence Figs. 5.22a-b represent an evolving spectrum of heating/drying profiles that shift from NLP to LP tropical oceanic convective clouds, and terminate with the intense island electrified convection of the IT simulation (e.g., “Hector”; Keenan et al., 1994). Differences in the heating and drying profiles are systematic as one transitions from the NLP to the intense-LP convection. In the case of the NLP convection (e.g., the NACT profile), the vertical structure of the heating profile (Fig. 5.22a) can be described as having a large low-level peak at 2 km and a much smaller secondary peak near the 5 km level. The drying profile (NACT; Fig. 5.22b) exhibits pronounced peaks at 2.6 km and at 800 m (though the 800 m peak is an artifact of the forcing) and a gradual decrease in drying with height above the 2.6 km level. Above 6 km, the heating falls off rapidly with height in the NACT profile. The heating and drying profiles of the NACT simulation were driven primarily by condensational (water) heating in the early growth phase of the cloud and later modulated by eddy flux contributions of heat in the downdraft (Fig. 5.23a). Ice processes

(deposition, freezing) played only a minor role in the structure of the NACT heating profile (Fig. 5.23a).

As the simulated convection became incrementally more electrified (i.e., moving from the NACT to the NACT3, NACT4 and onto the ACT and IT profiles), the primary heating and drying peaks shifted relative to one another (Figs. 5.22a-b), until finally the upper level heating peaks in the ACT and IT profiles (6-7 km) became larger than their respective low level heating peaks (2-3 km). Furthermore, the ACT and IT low-level peaks were both weaker than the low-level NACT peak (i.e., a juxtaposition of the heating peaks with height occurred). Also note that as the simulations shifted from NLP to LP, the slope of the heating profile above 6 km became gradually less negative with respect to height, and even positive in the case of the IT cloud. The drying peaks behaved similarly (Fig. 5.22b) with regard to the simulation-relative reversal in the degree of drying at upper levels, however the upper level peaks in all simulations were still less than the low level peaks. The change in the heating/drying structure over the spectrum of the simulations was *not* a strong function of the low-level forcing. *Differences in the vertical profiles of heating and drying were more reflective of the sounding characteristics and the dynamical/microphysical characteristics of the modeled clouds. The dynamical/microphysical characteristics in turn, drive significant electrical differences; therefore we offer evidence that lightning observations should reveal gross characteristics of the heating and drying profiles on convective scales.*

The juxtaposition in the heating profile peaks from NACT to IT simulations can be further explored by examining the heating budget terms for the ACT and IT profiles (Figs. 5.23b-c). Note in Figs. 5.23b-c that condensational heating of the ACT and IT clouds took place over a deeper layer of the lower and middle troposphere (consistent with the more persistent mixed phase conditions in the ACT/IT clouds relative to the NACT), and pronounced upper level peaks

in deposition occurred between the 8 km and 12 km levels. Above 11 km in both the ACT and IT profiles, eddy flux convergence of heat further modulated the heating profiles. When the contributions from deposition, condensation and eddy fluxes were summed, the magnitude of the heating in the mixed phase region of the ACT and IT clouds increased and the negative slope of the heating profile above the 6 km level decreased (i.e., diabatic heating occurred over a deeper layer of the LP clouds aloft). The relative change in the heating and drying profiles in transition from the NACT to IT simulations is associated primarily with the development of much stronger updrafts and more robust mixed-phase microphysics in the mid to upper levels of the ACT and IT clouds. Enhanced mixed-phase precipitation processes (e.g., condensation of more cloud water and vertical transport of raindrops, followed by freezing/accretional growth of FD and graupel at temperatures colder than 0°C) in the ACT and IT clouds were responsible for the increased ice-mass contributions to the total rainfall, and enabled efficient NIC processes to produce electrical break down of the modeled clouds.

The transition in heating profile above the 5 km level and below the 4 km level, relative to the time of electrical break down for the tropical oceanic simulations, is further illustrated in a time-series format shown in Figs. 5.24a-c. Moving from top (ACT) to bottom (NACT) in Figs. 5.24 a-c, note that the accumulated low-level heating (condensation) for each simulation exhibited a peak of similar magnitude between $t=900$ and 1200 seconds (just prior to the cessation of prescribed gust front forcing). However, the latter peak in heating at upper levels (5-9 km) became progressively smaller as the simulations transitioned from the ACT to NACT clouds. It is also apparent that the break down time of the clouds shifted to larger values as the upper level heating peak decreased in magnitude. The tendency in the heating profiles in Figs. 5.24a-c, especially above 5 km, is consistent with the relative electrical intensity and the degree of mixed phase ice-processes active in each simulated cloud.

For comparison to the tropical simulations, consider the heating profile of the MC simulation shown in Fig. 5.25. This profile should be considered the “extreme” end of the model-simulated convective heating spectrum. In the MC cloud the overwhelming majority of the precipitation process was associated with a robust mixed-phase. Note that greater than one half of the condensational heating took place above the 0°C level (~4 km) and this was coupled with significant peaks in vapor deposition and eddy flux convergence of heat aloft. Interestingly, the sum of the individual heating components reached a relative maximum between the 6 km and 7 km levels (-12°C to -20°C), which is also where a large majority of the electrical charging of the cloud occurred.

If the time-series of heating below 4 km and above 5 km for the MC case is compared to the intense LP cloud of the IT simulation, some interesting differences and similarities surface (Figs. 5.26a-b). For example, the low level (0-4 km) heating peak in the IT simulation is approximately equal to the upper level heating peak (though in an integrated sense, the upper level heating is larger), while in the MC simulation the upper level peak is larger than the low level peak. This is due to the different distributions of condensational heating in the two simulations. In IT case (Fig. 5.23c), a larger fraction of the total condensational heating (leading to the formation of a “warm-rain” peak in surface rain rate) occurred below the 4 km level in the cloud, whereas just the opposite occurred in the MC simulation (consistent with the dominant ice-based precipitation process). To some extent the difference in the condensational heating is forced by the different specifications of the autoconversion threshold (0.5 g kg⁻¹ in the IT simulation, 2 g kg⁻¹ in the MC simulation), however, it is also true that ice-processes generally dominate the production of rainfall in MC clouds (Knight et al., 1974; Young, 1993).

The height-partitioned time-series of heating for the two simulations also exhibits some interesting similarities (Figs. 5.26a-b). Both simulations produced stronger evaporative cooling

peaks below 4 km due to the development of a larger downdraft. The evaporation of rain and cloud-water in the downdrafts served to reduce the low level heating peak near the surface and to add moisture to the lowest 1 km of the modeled-cloud domains. In both the MC and IT simulations, electrical breakdown of the cloud occurred during the primary peak in upper level heating, which was associated with condensation in the updrafts, and a rapid development of the ice-phase between 6 and 8 km. Additionally, the upper level peaks in both simulations exhibited secondary maxima after electrical break down that were associated with the total glaciation of the IT and MC clouds as they both reached their maximum vertical extent. Note that the ACT upper level heating peak (Fig. 5.24a) also exhibited a secondary maximum (absent in the NACT cloud; Fig. 5.24c), albeit substantially smaller than either of the secondary peaks in the intense IT and MC clouds.

5.4 Summary

The 1-D modeling simulations suggest the following answers to questions 1-4 posed at the beginning of the chapter and at the end of Chapter 4:

- 1) The simulations (Table 5.4b) suggest that tropical oceanic cumulonimbi forming in environments with progressively larger CAPE, (e.g., tropical oceanic, tropical island continental etc.), become progressively more electrified. The importance of the vertical distribution of the thermal buoyancy in the simulations becomes clear when comparing the relative intensities of the ACT, IT and MC modeled clouds to the relative distributions of their respective thermal buoyancies shown in Fig. 3.12. The larger the positive slope in thermal buoyancy with height in the low to mid-levels of the troposphere, the more vertically developed the cloud will become, and the more likely

the cloud is to produce lightning. Consistent with the observations in Chapter 3, model sensitivity tests also suggested that relatively small perturbations in the boundary layer moisture content (which produced large changes in CAPE) exerted a slightly stronger control over the intensity of the simulated oceanic convection relative to similar temperature perturbations (when the standard deviations of the two variables in the composite soundings are compared).

- 2) Frozen drops, *as parameterized in the 1-D cloud-model*, were an explicit component of the NIC process and subsequent electrical break down of the model clouds. The height of the 36 dBZ reflectivity contour in the model simulations was highly correlated to both the relative electrical intensity of the cloud (as gauged by the time to electrical break down) and the peak FD mixing ratios (Table 5.4b). The idea of a threshold in the mean vertical velocity as related to the terminal fall speeds of the frozen drops (e.g., Fig. 4.23; Zipser and Lutz, 1994) seems to be verified in the model simulations. This is readily illustrated in Fig. 5.27 by modifying Fig. 4.23 to include the model temporal mean vertical velocity across the upper 1 km (to account for the volume sampling nature of a radar) of the 36 dBZ core as a function of the modeled 36 dBZ core height for the majority of the sensitivity runs shown in Tables 5.4a-b. In Fig. 5.27 the mean vertical velocities near the top of the 36 dBZ core in the model simulations are very similar to the maximum mean vertical velocities observed in the multiple Doppler retrievals of vertical velocity in several cases from TOGA COARE. Note that the time-mean model vertical velocities associated with the 36 dBZ contour also suggest the presence of frozen drops with diameters of 1-2 mm. However, it should be noted that in the case of a bulk microphysical model it is more appropriate to refer to mass-weighted terminal fall speeds. In the case of exponential size

distributions and terminal fallspeed relationships ($V_t=aD^b$) used for the FD category in the 1-D model, the mass weighted terminal fall speed of the distribution represent *individual* particle diameters approximately twice the size of the distribution mean. Therefore the vertical velocities computed for model 36 dBZ heights in Fig. 5.26 represent volume mean drop size diameters on the order of 0.6-1.0 millimeters (O[1 mm]). Inspection of the peak vertical velocities in Table 5.4b suggests that *peak* updraft speeds must exceed a threshold of approximately 11 ms^{-1} near the -5°C level before the cloud will achieve break down electric fields (similar to Solomon and Baker, 1994). When compared to *mean* vertical velocity observations in tropical oceanic convection (e.g., Zipser and LeMone, 1980), the simulation vertical velocity threshold is a factor of two larger than the strongest 10% of the *mean* updraft cores observed near the 4.5 km level ($\sim 0^\circ\text{C}$). This is consistent with the relatively small fraction 1-10% of deep reflectivity cores (e.g., 30 dBZ cores above the -10°C level) and lightning producing convection observed over the tropical ocean (DeMott, 1996, and Orville and Spencer, 1979 respectively).

- 3) The modeled diabatic heating profiles underwent a marked transition in structure as the simulated clouds evolved from NLP, MLP, LP and intense-LP storms. The transition was characterized by a juxtaposition in the magnitude of the low level (2-3 km) and upper level (5-7 km) heating peaks, as the simulated cells incrementally intensified from NLP to LP (e.g., Fig. 5.22a). The juxtaposition of the low and mid-level heating peaks occurred as a result of the presence of enhanced mixed-phase conditions above the height of the 0°C level. The heating profile associated with the LP convection in the upper regions of the modeled clouds also exhibited a less negative slope with respect to height (relative to the NLP or MLP storms) in association with

enhanced vapor deposition and eddy flux convergence of heat. Drying profiles exhibited a vertical structure trend similar to that of the heating profiles, with more convectively/electrically intense simulations exhibiting higher (lower) drying magnitudes aloft (near the surface).

- 4) As the intensity and electrical character (e.g., NLP, LP, intense-LP) of the modeled storms increased by regime (NACT, ACT, IT, MC), the direct contribution by melting ice-mass to the column integrated rainwater in each regime gradually increased to values of 21%, 34%, 42%, and 75% respectively. The increase in the contribution to rainfall by ice-processes as an *implicit* function of electrical intensity is well supported by the model simulations. The term "*implicit*" is necessary because not all of the ice mass that contributed to the rainfall was *explicitly* associated with the electrification process. Indeed, the *implicit* nature of the relationship between an ice-phase precipitation process and cloud electrification is further illustrated when it is noted that observations suggest that less than 10% of precipitation-sized ice-particles in LP clouds possess a measurable electrical charge (e.g., Dye et al., 1986; Weinheimer et al., 1991). The *explicit* relationship between the precipitation ice-process and total rainfall is a function of the ability of the updraft to supply copious amounts of liquid water and water vapor to elevations above the freezing level, which is directly related to the efficiency with which the NIC mechanism will function in the cloud. Given the increasing fraction of rainwater supplied by ice-process in LP convection and the relationship between ice-processes and electrical intensity, one might hypothesize that deep convective cloud regimes associated with highly electrified convection will produce lightning flash densities that are related in some fashion to the total mass of

rainfall over some area and time (e.g., an “output” of the electrified, convective cloud; e.g., Fig. 1.1, Table 1.1). *This hypothesis will be addressed in Chapter 6.*

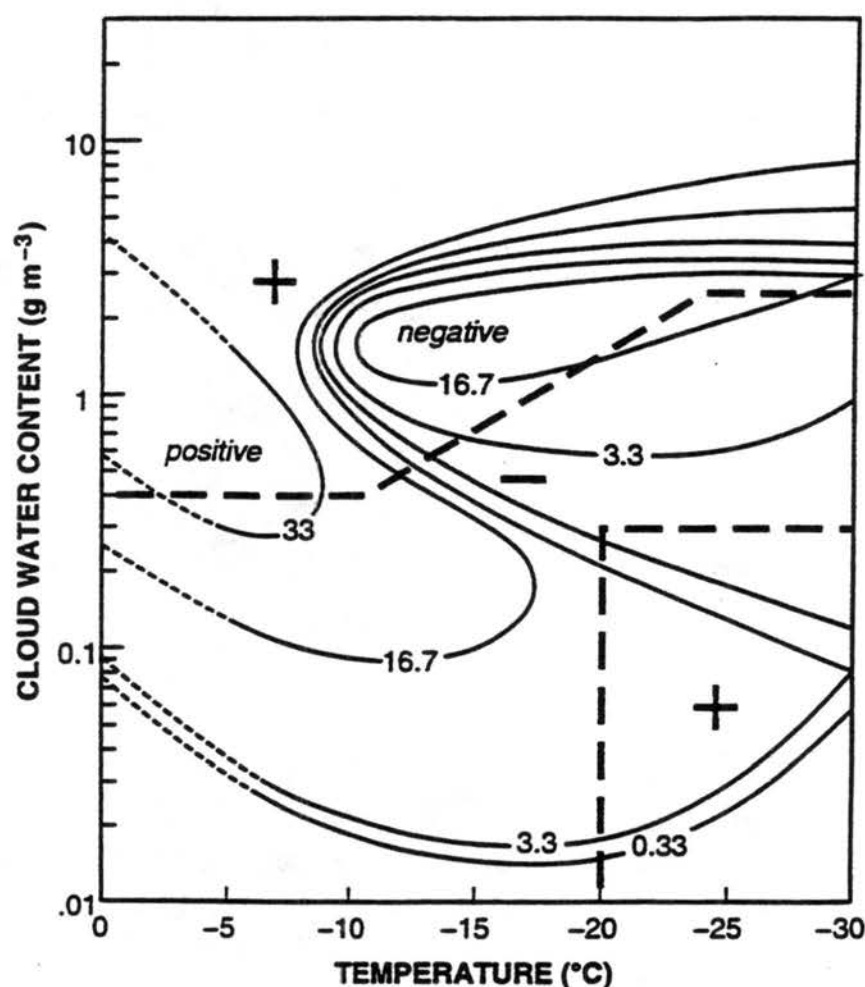


Figure 5.1 Combined laboratory results of Takahashi (1978) [contoured] and Saunders et al. (1991) [bold dashed lines] for non-inductive charge transfer plotted in temperature-liquid water content space. Note that the Saunders et al. values are adjusted to total liquid water content by multiplying the effective water content by a factor of two. The (+) and (-) indicate negative and positive charge regions for the Saunders et al. results. Figure adapted from MacGorman and Rust (1997).

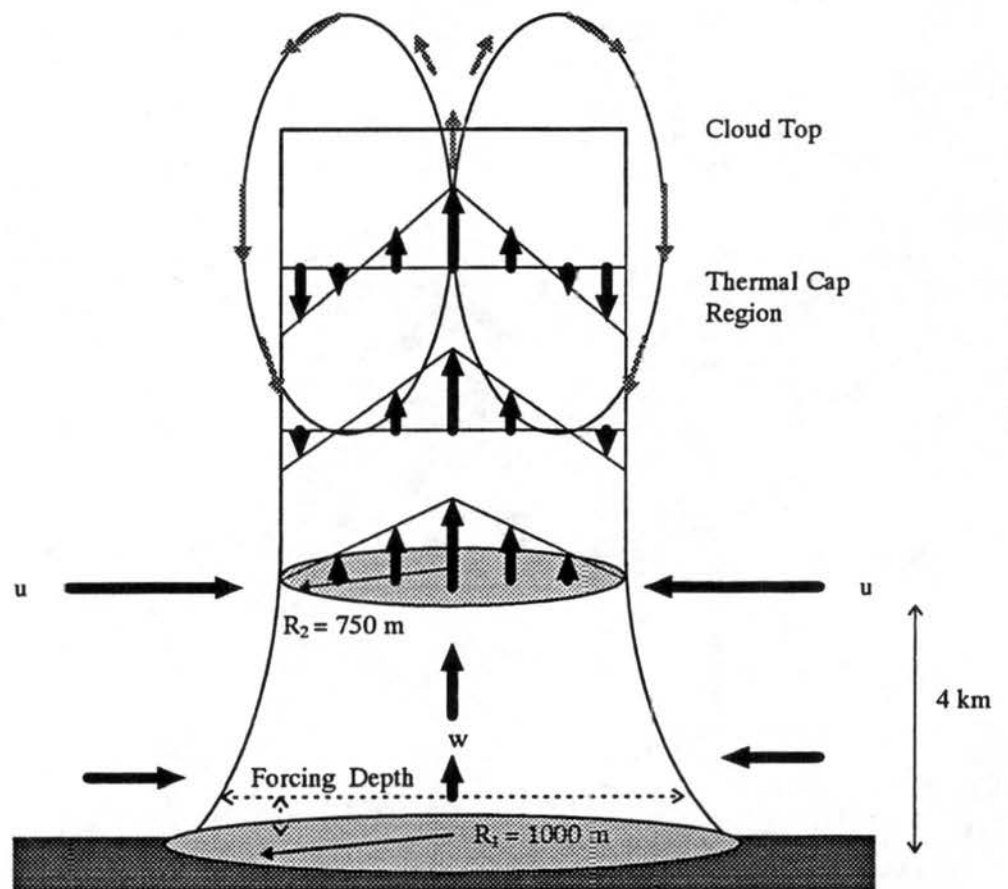


Figure 5.2 Cloud-model geometry illustrating the variable radius with height, the triangular updraft profile, and the thermal cap circulation (light shaded vectors). Bold, vertically (horizontally) oriented arrows represent vertical (radial) velocity.

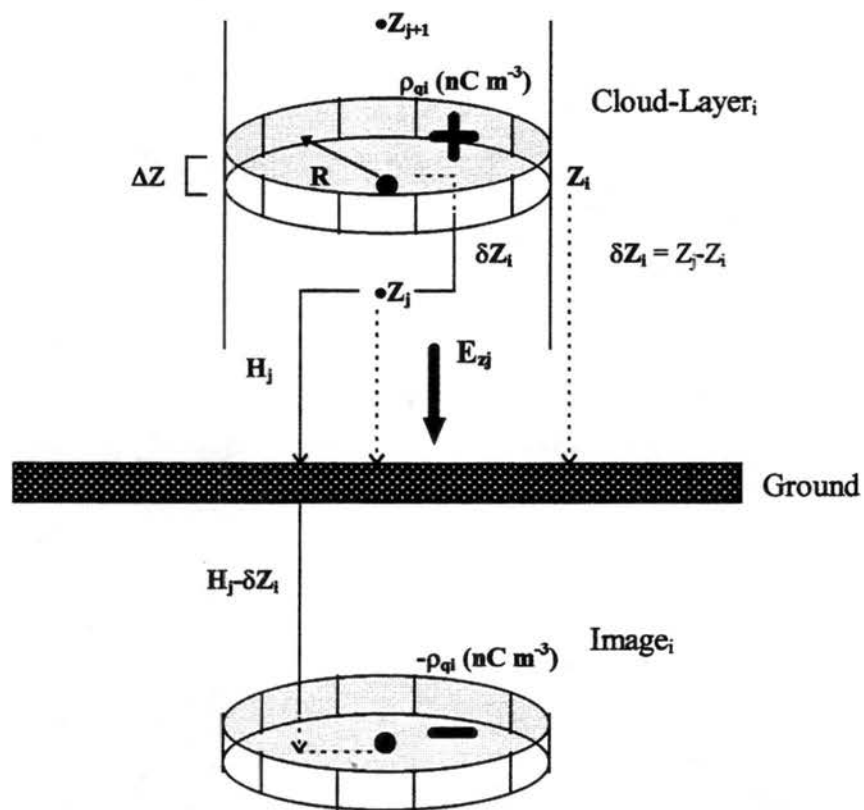


Figure 5.3 Geometry used for computing electric fields at point Z_j in the 1-D model. A disk of positive charge ΔZ meters thick is located in the cloud layer Z_i , of radius R , situated δz_i meters away from point Z_j . An image charge is also placed below the Earth's surface.

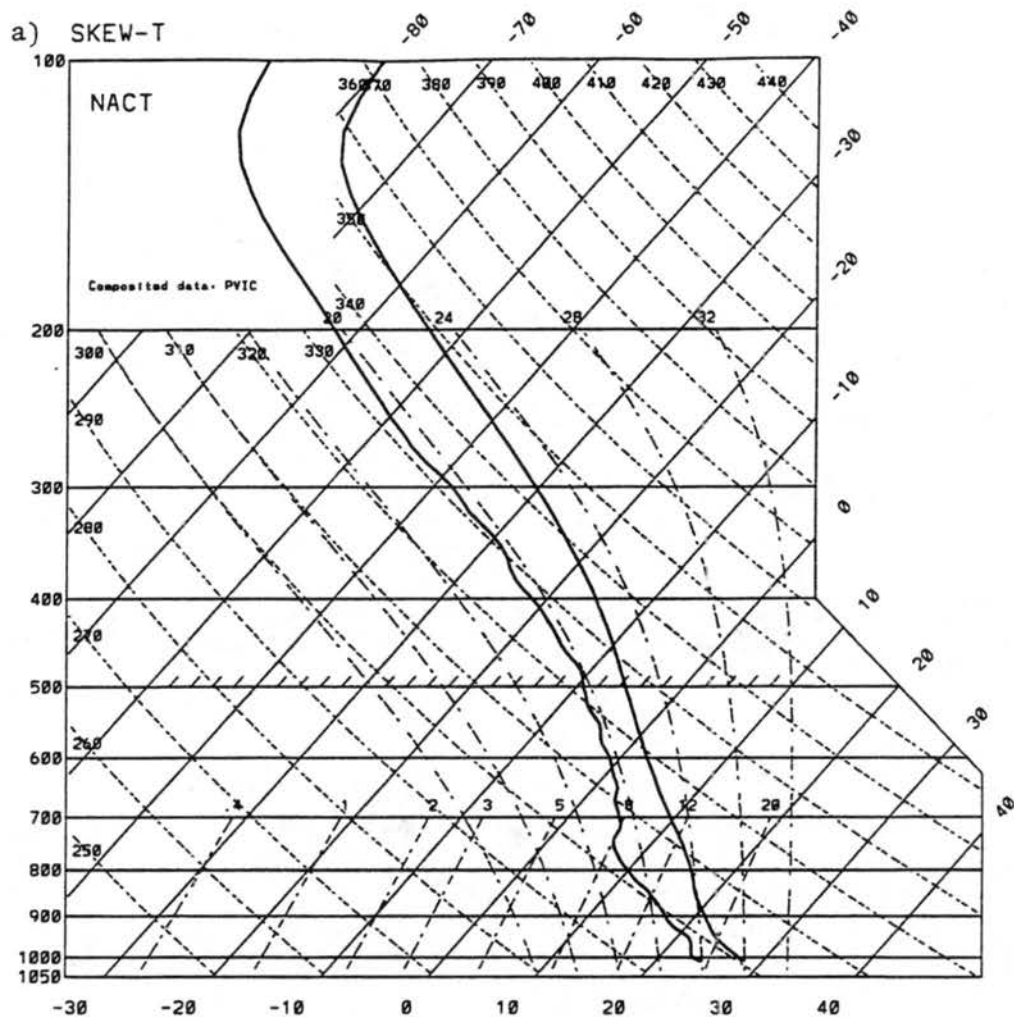


Figure 5.4 a) Skew-T profile of temperature and dewpoint for the NACT sounding;

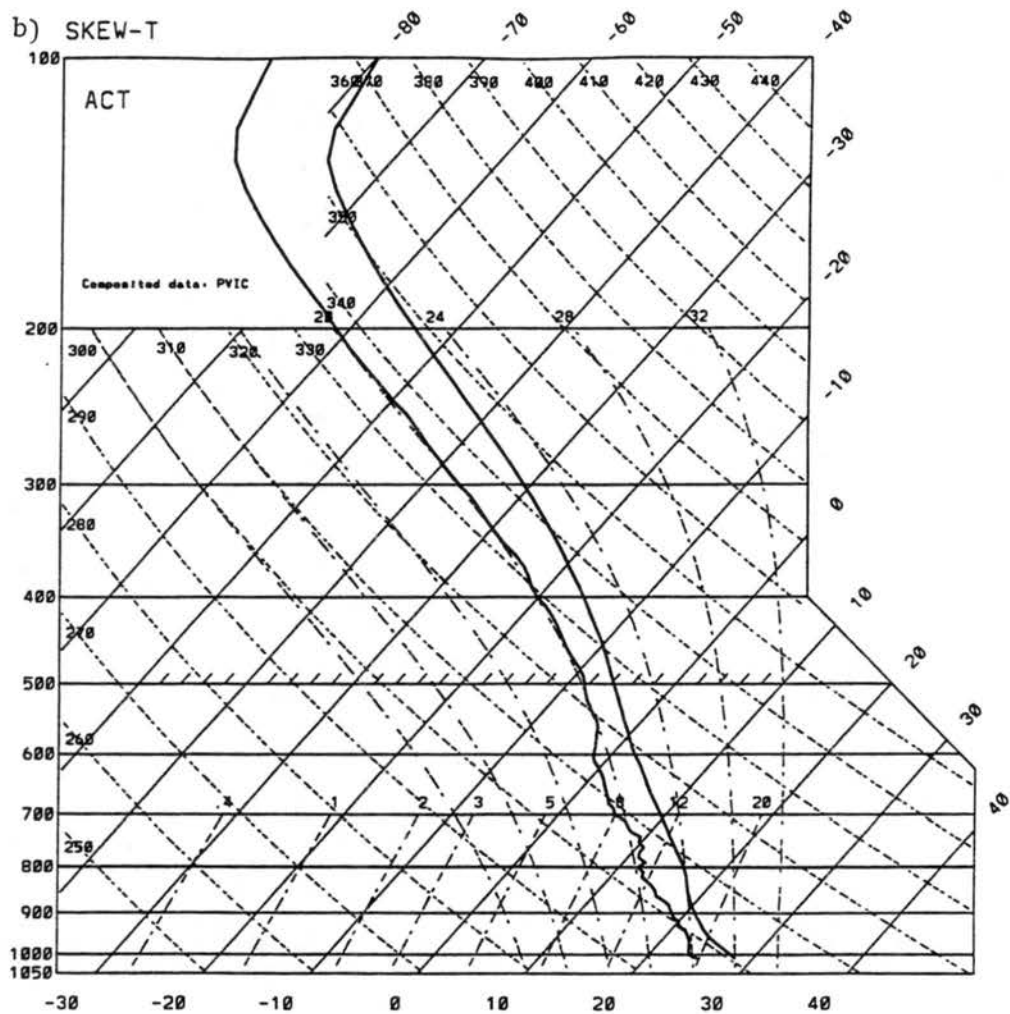


Figure 5.4 b) As in (a) but ACT sounding.

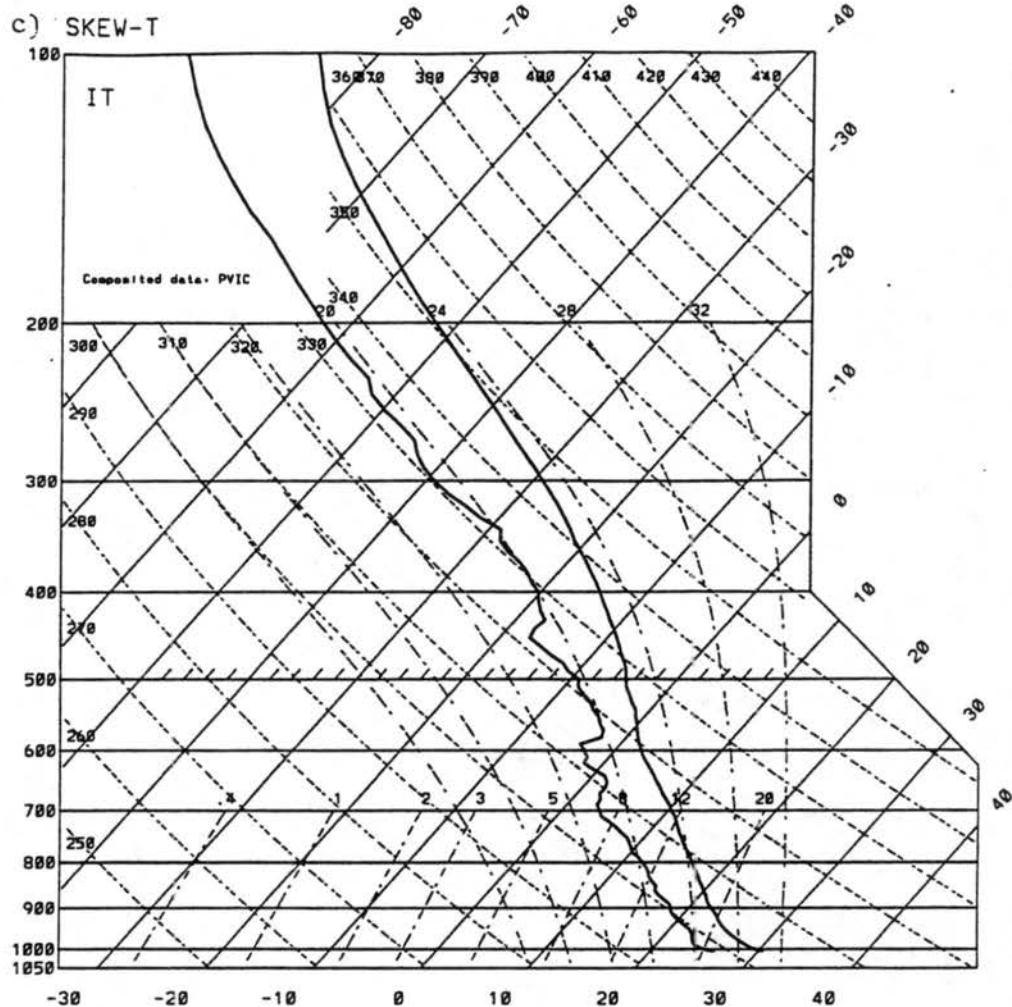


Figure 5.4 c) As in (a) but IT sounding.

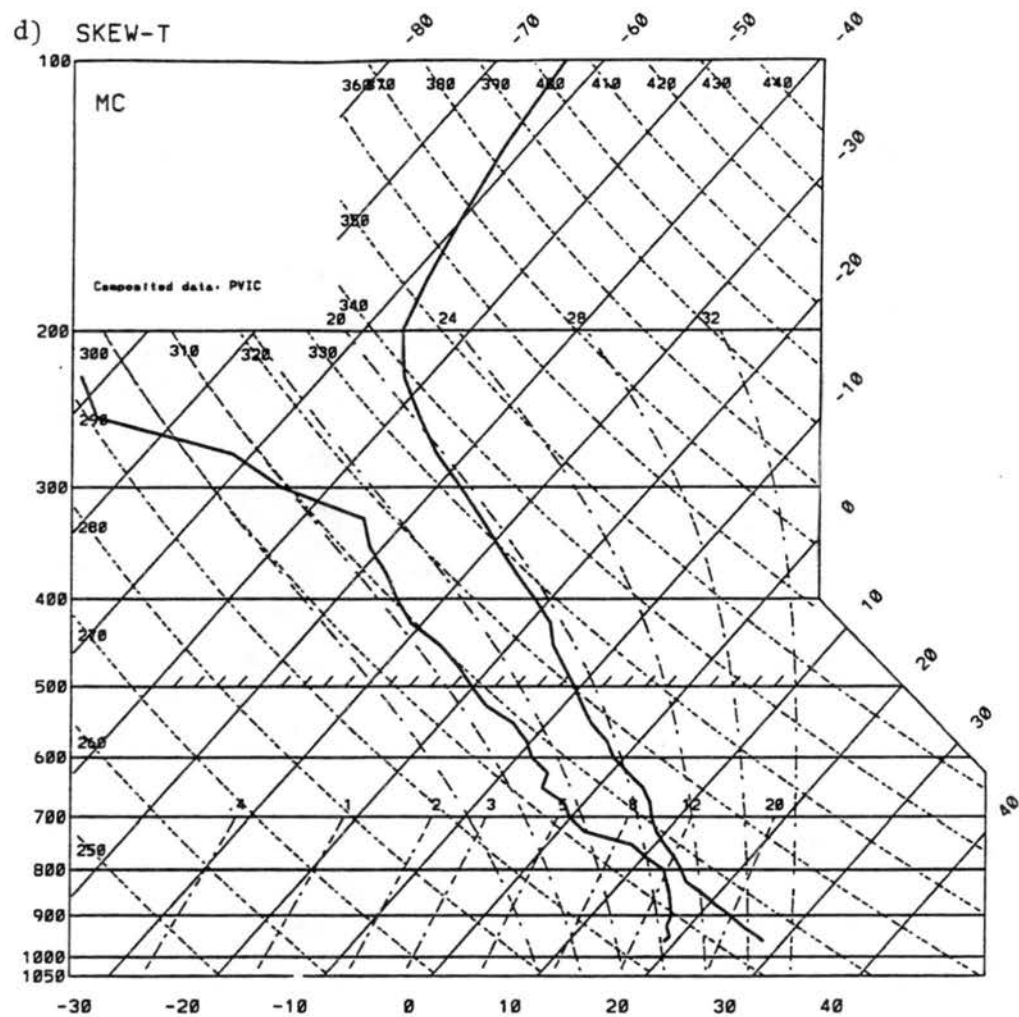


Figure 5.4 d) As in (a) but MC sounding.

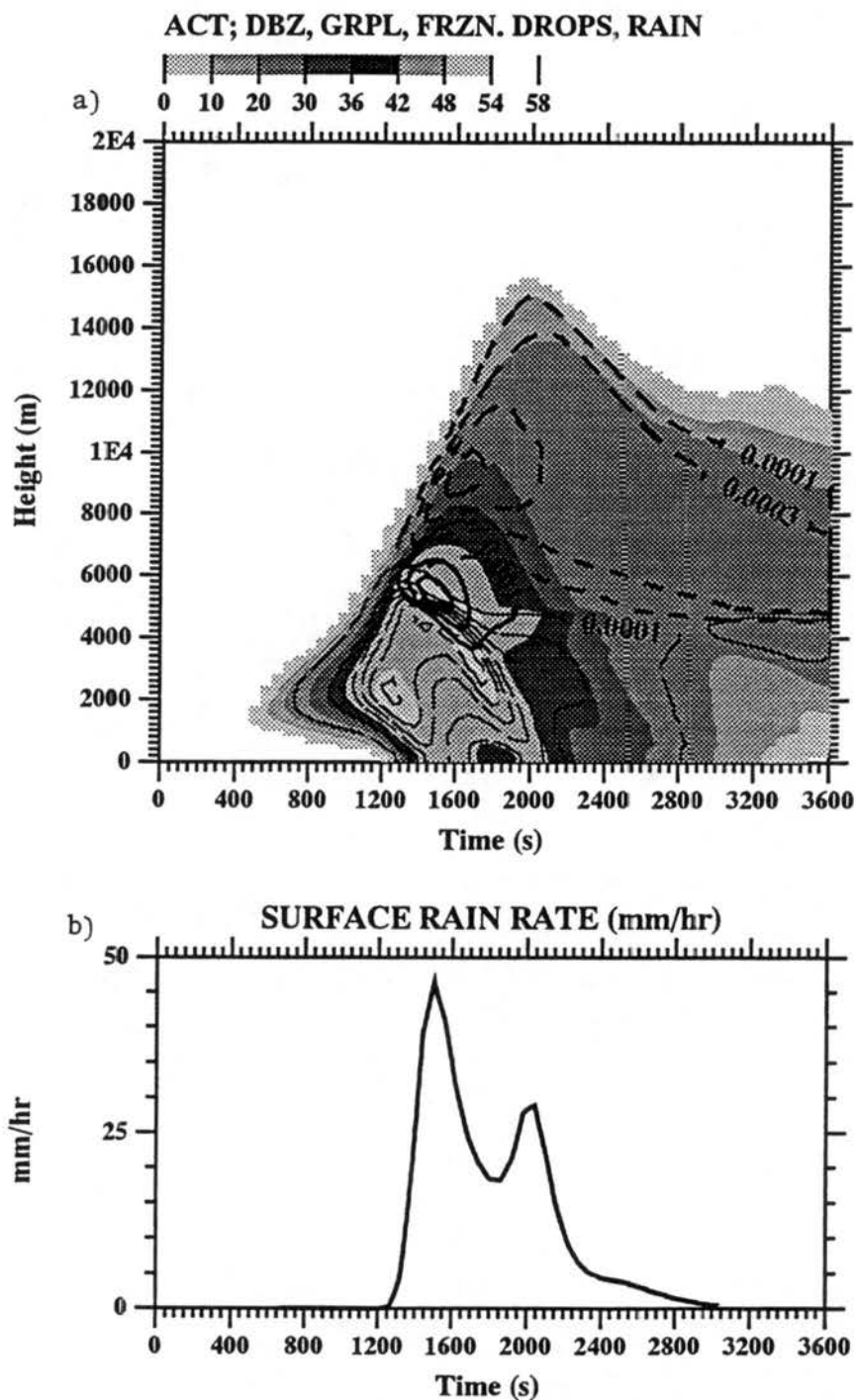


Figure 5.5 ACT simulation time-height representation of a) radar reflectivity, and mixing ratios of graupel (bold dash, contoured every $.0002 \text{ kg kg}^{-1}$), frozen drops (bold solid, contoured every $.001 \text{ kg kg}^{-1}$ beginning at $.001 \text{ kg kg}^{-1}$), and rain (light dash, contoured every $.0005 \text{ kg kg}^{-1}$ beginning at $.0001 \text{ kg kg}^{-1}$). Time (seconds) is plotted on the abscissa, height on the ordinate (km). b) Surface rainfall rate (mm hr^{-1}) with time.

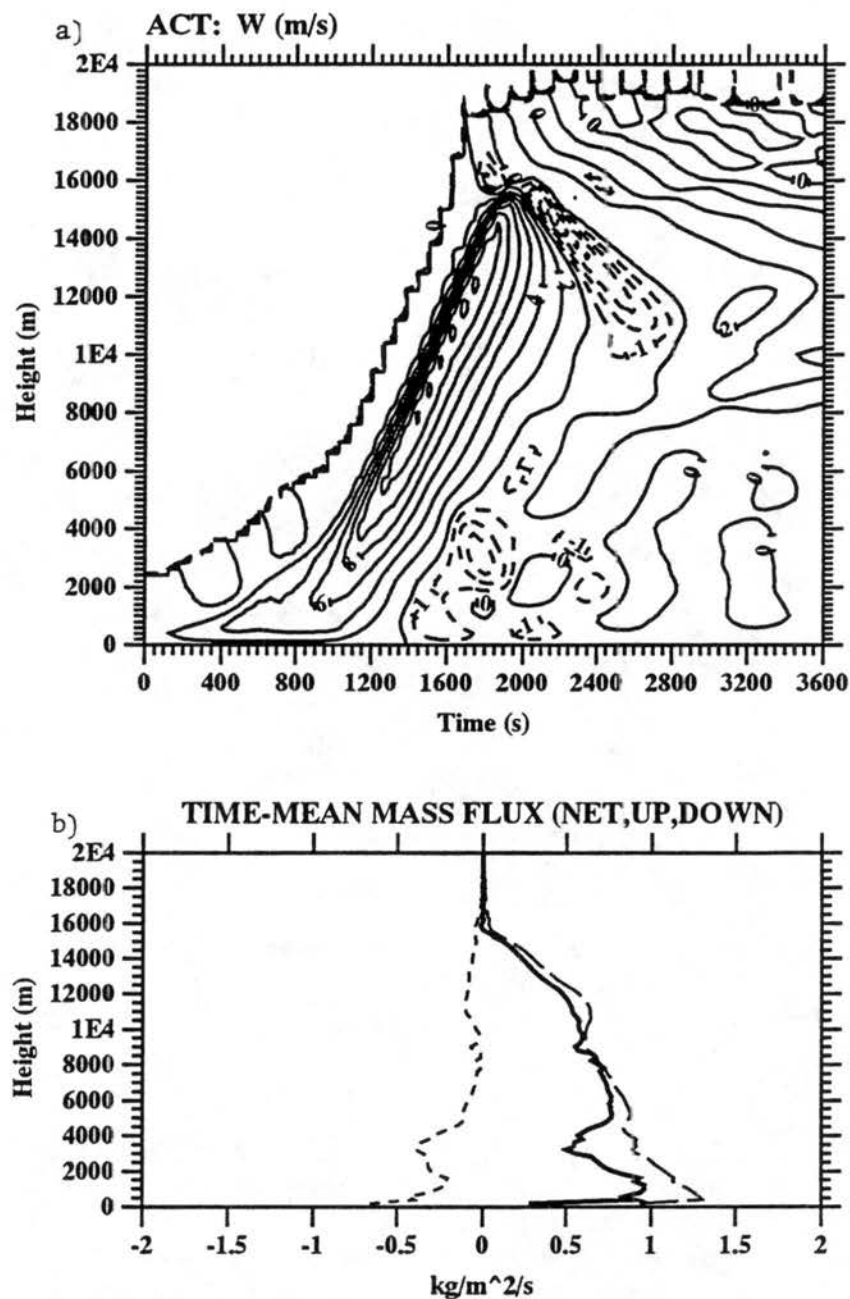


Figure 5.7 a) ACT simulation time-height profile of vertical velocity (m s^{-1}). Updrafts (downdrafts) are contoured every 2 (1) m s^{-1} using solid (dashed) lines. b) ACT-simulation mean mass-flux ($w \cdot \rho$) in $\text{kg m}^{-2} \text{s}^{-1}$ for the net (solid), updraft (long dash) and downdraft (short dash) vertical velocity.

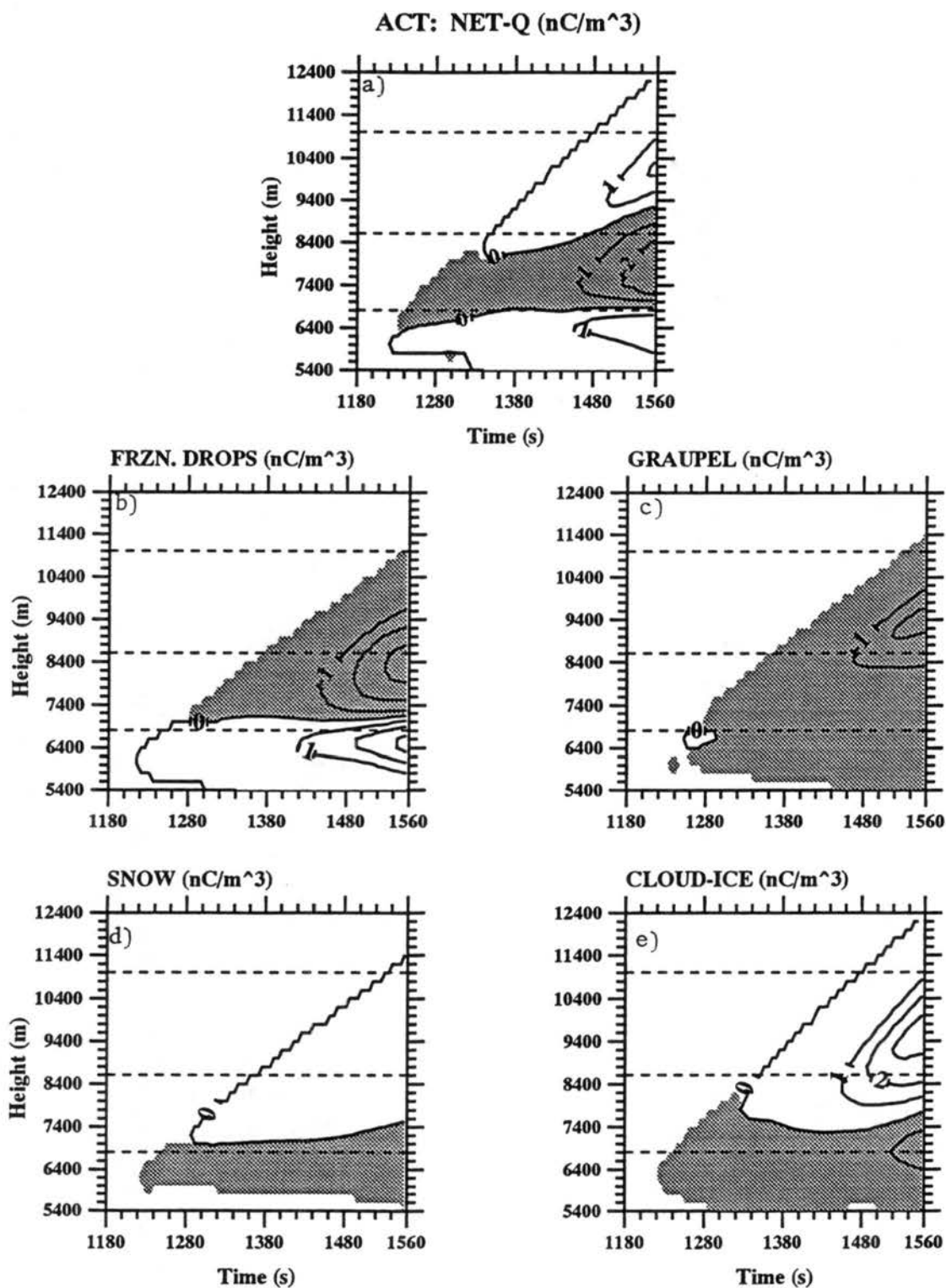


Figure 5.8 ACT simulation time-height profiles of charge density(nC m^{-3}) for a) Net charge; b) charge on frozen drops; c) charge on graupel; d) charge on snow; and e) charge on cloud-ice. Charge densities are contoured every 1 nC m^{-3} , negative charge values are shaded. Charge densities are plotted until the time of electrical break-down.

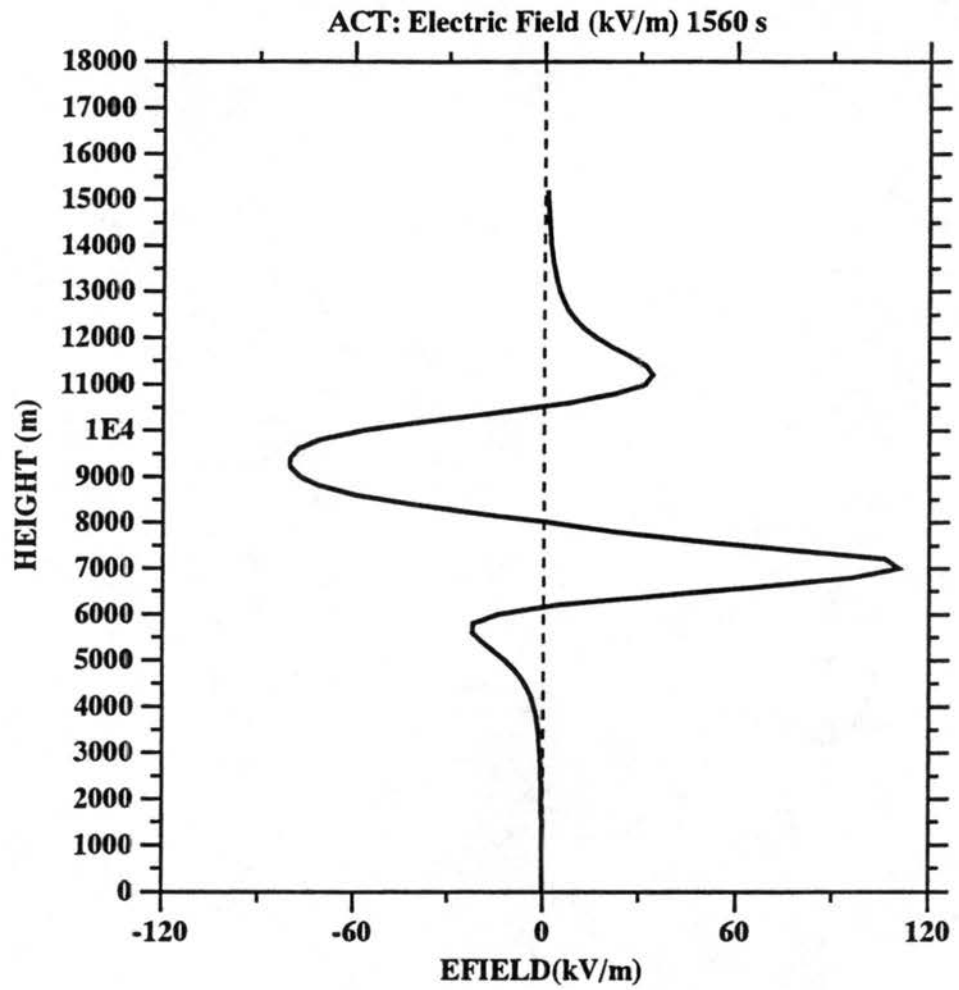


Figure 5.9 Vertical component of the electric field (kV m^{-1}) plotted with respect to height for the ACT simulation at $t = 1560$ seconds (electrical break down).

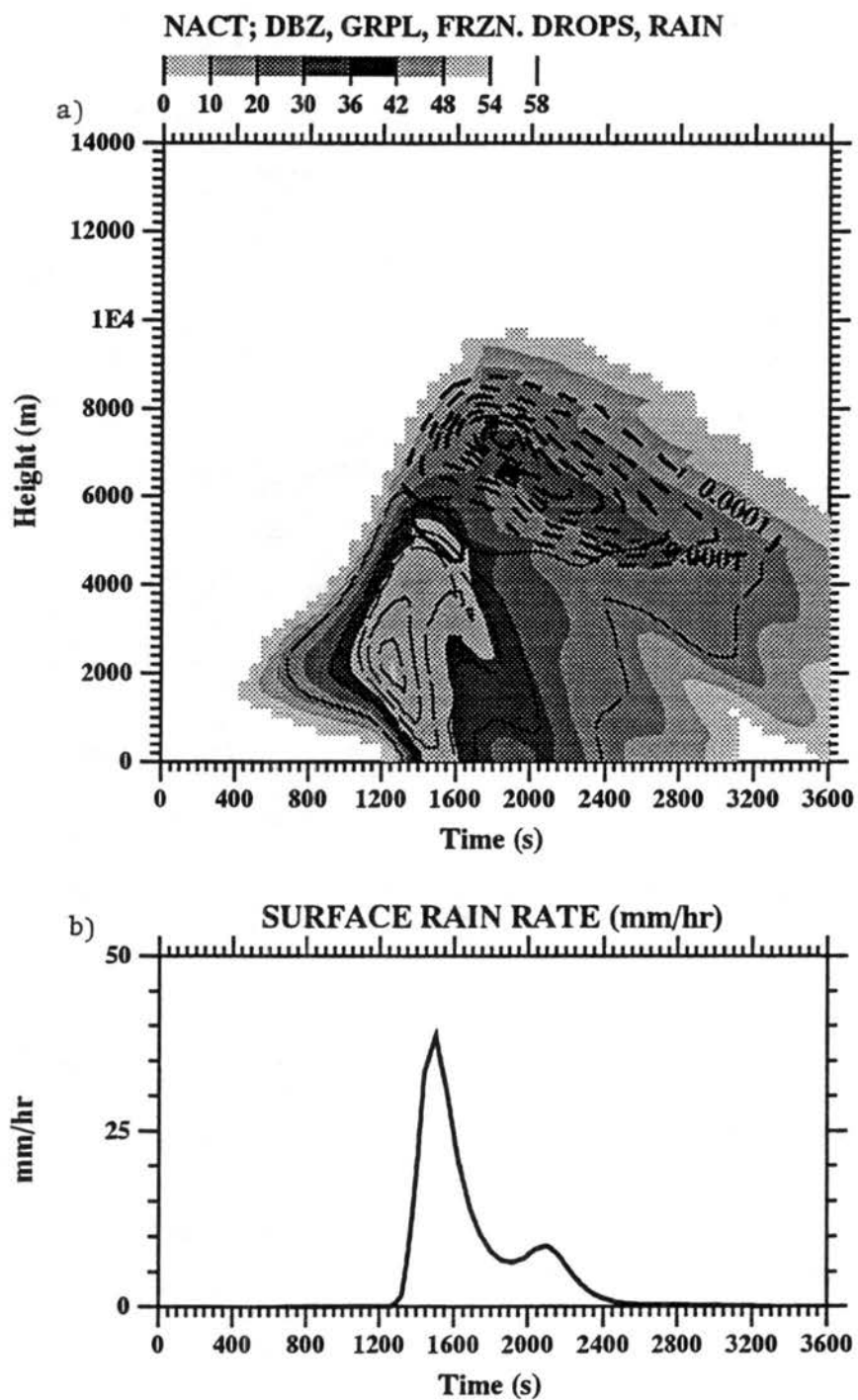


Figure 5.10 As in Fig. 5.5 but NACT simulation of a) radar reflectivity and mixing ratios of graupel, frozen drops and rain; and b) surface rainfall rate.

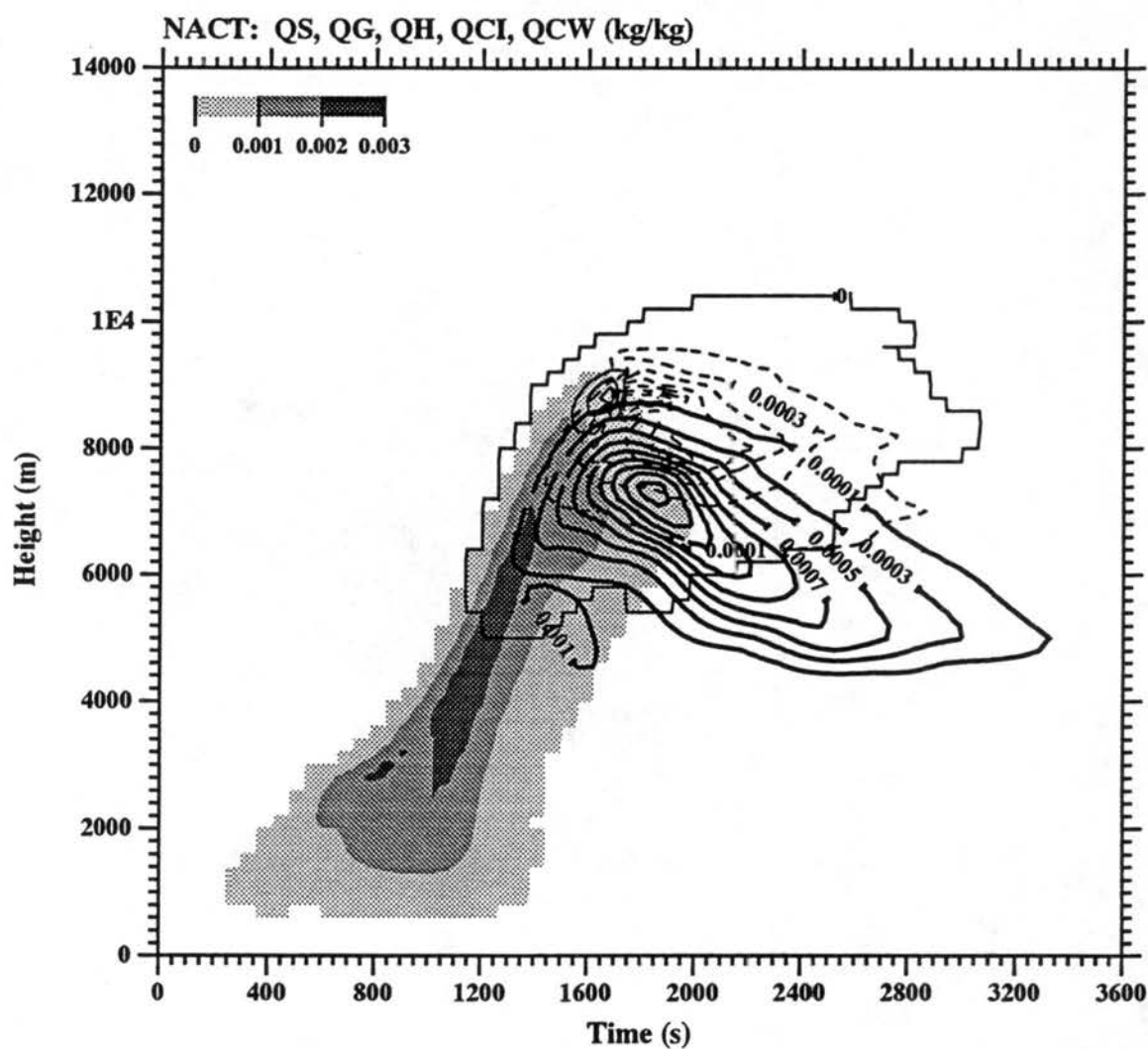


Figure 5.11 As in Fig. 5.6. but NACT simulation of cloud-water (shaded), graupel, frozen drop, snow, and cloud-ice mixing ratios.

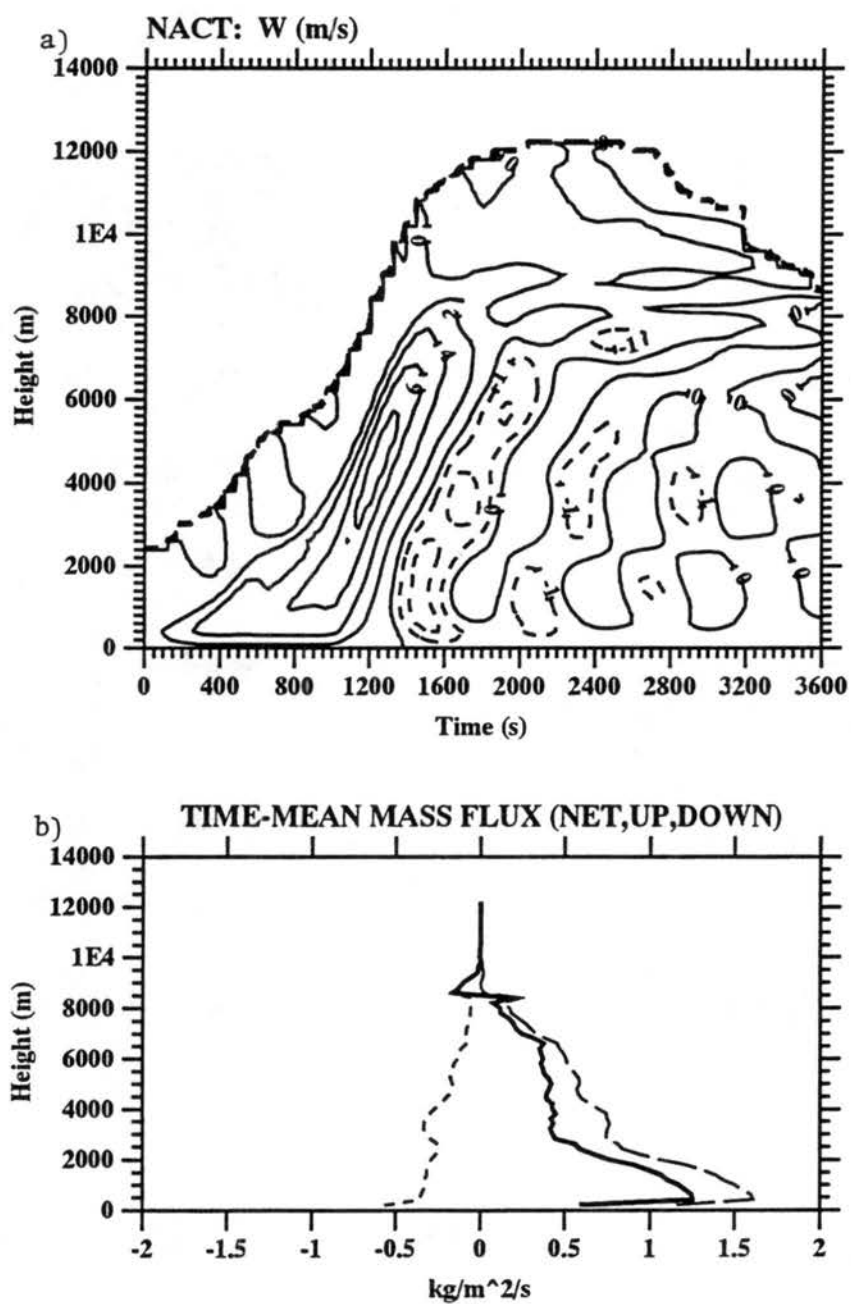


Figure 5.12 As in Fig. 5.7 but NACT simulation: a) vertical velocity and b) net, upward and downward mass-fluxes.

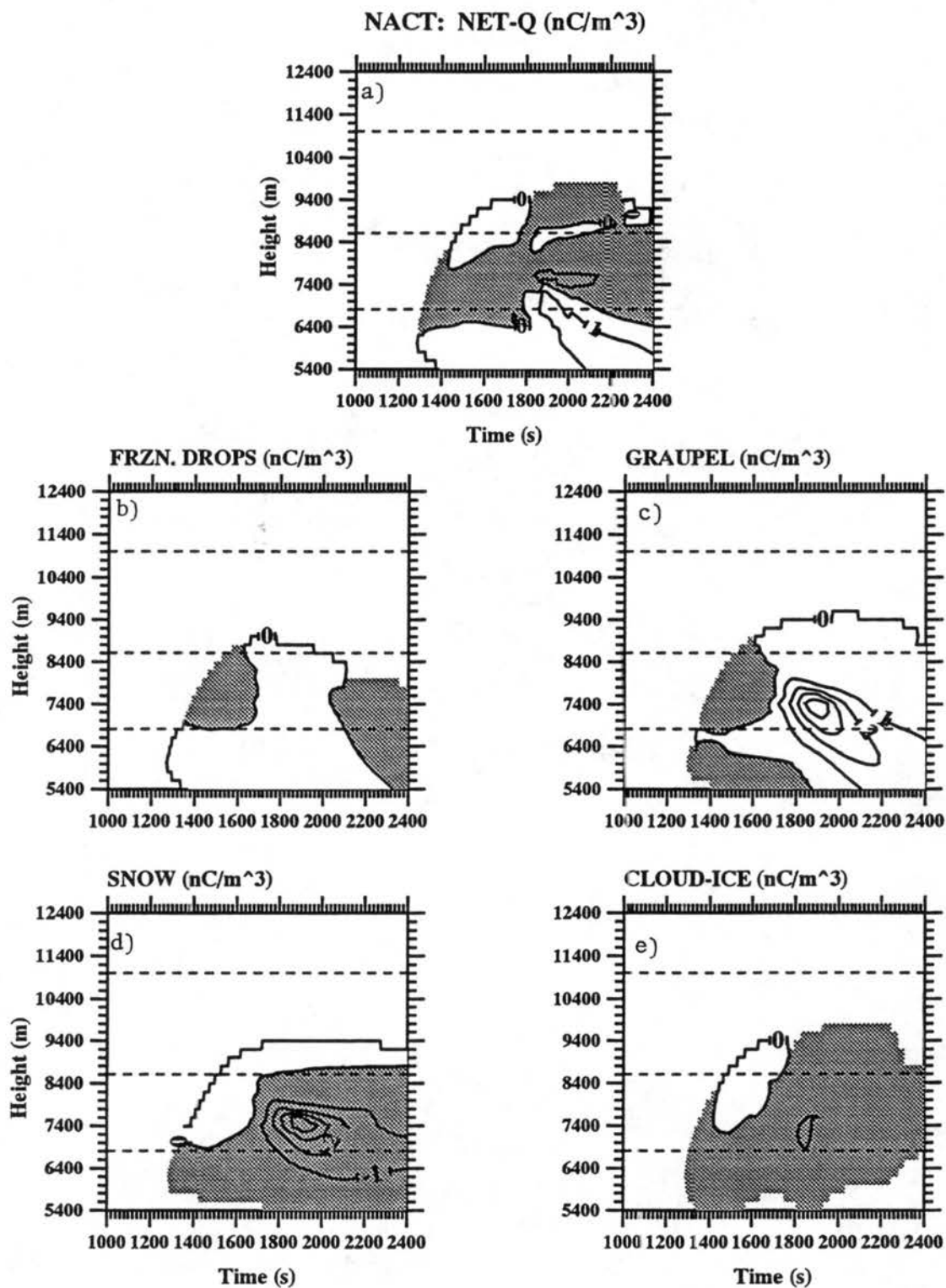


Figure 5.13 As in Fig. 5.8 but NACT simulation. Charge density (nC m^{-3}): a) net; b) frozen drops; c) graupel; d) snow; e) cloud-ice.

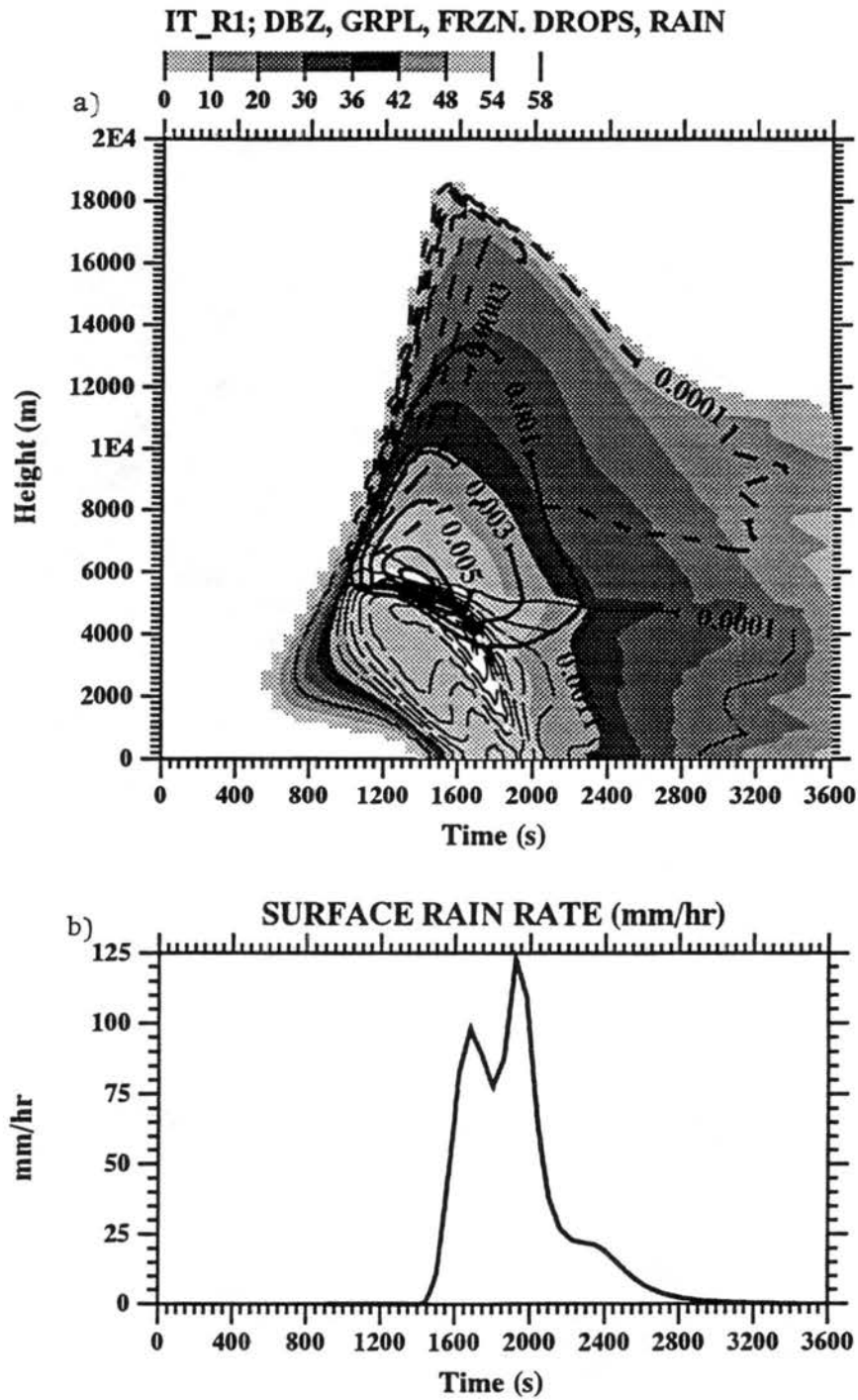


Figure 5.14 As in Fig. 5.5 but IT simulation of a) radar reflectivity, and mixing ratios of graupel, frozen drops and rain; and b) surface rainfall rate.

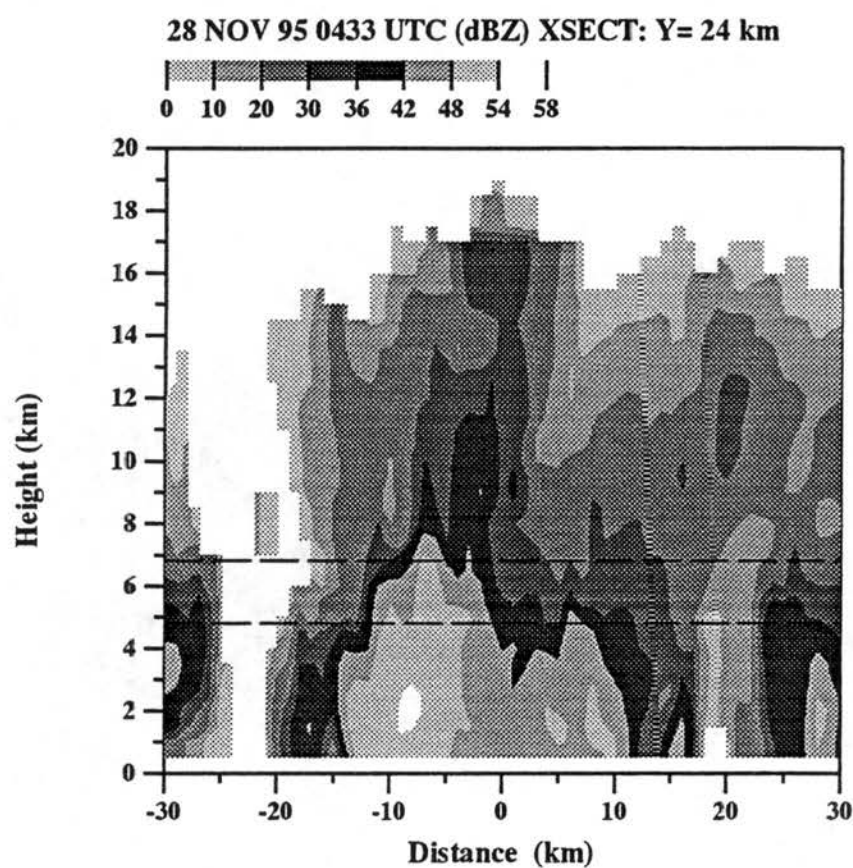


Figure 5.15 A vertical cross-section of radar reflectivity collected by the C-POL radar during MCTEX on 28 Nov. 1995 (0433 UTC).

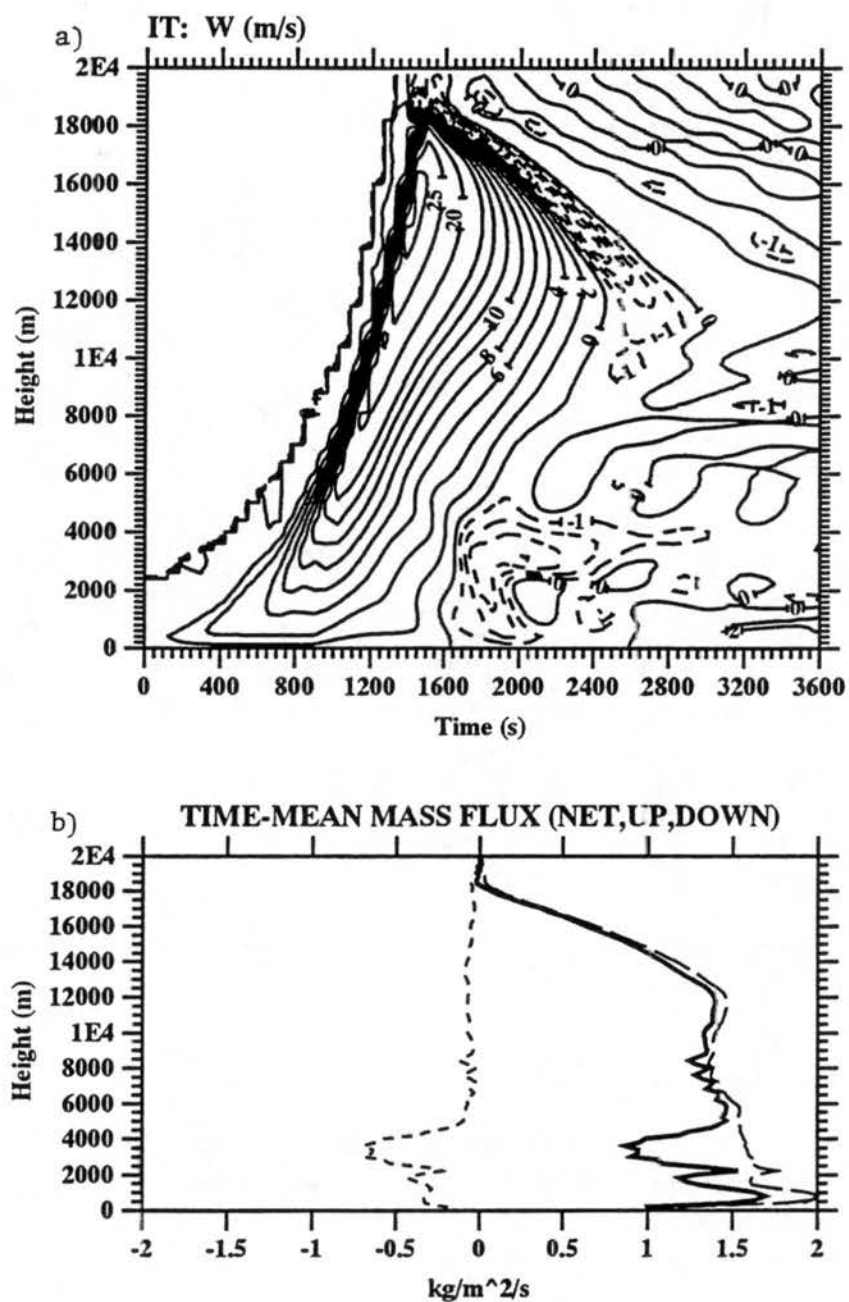


Figure 5.17 As in Fig. 5.7 but IT simulation: a) vertical velocity and b) net, upward and downward mass fluxes.

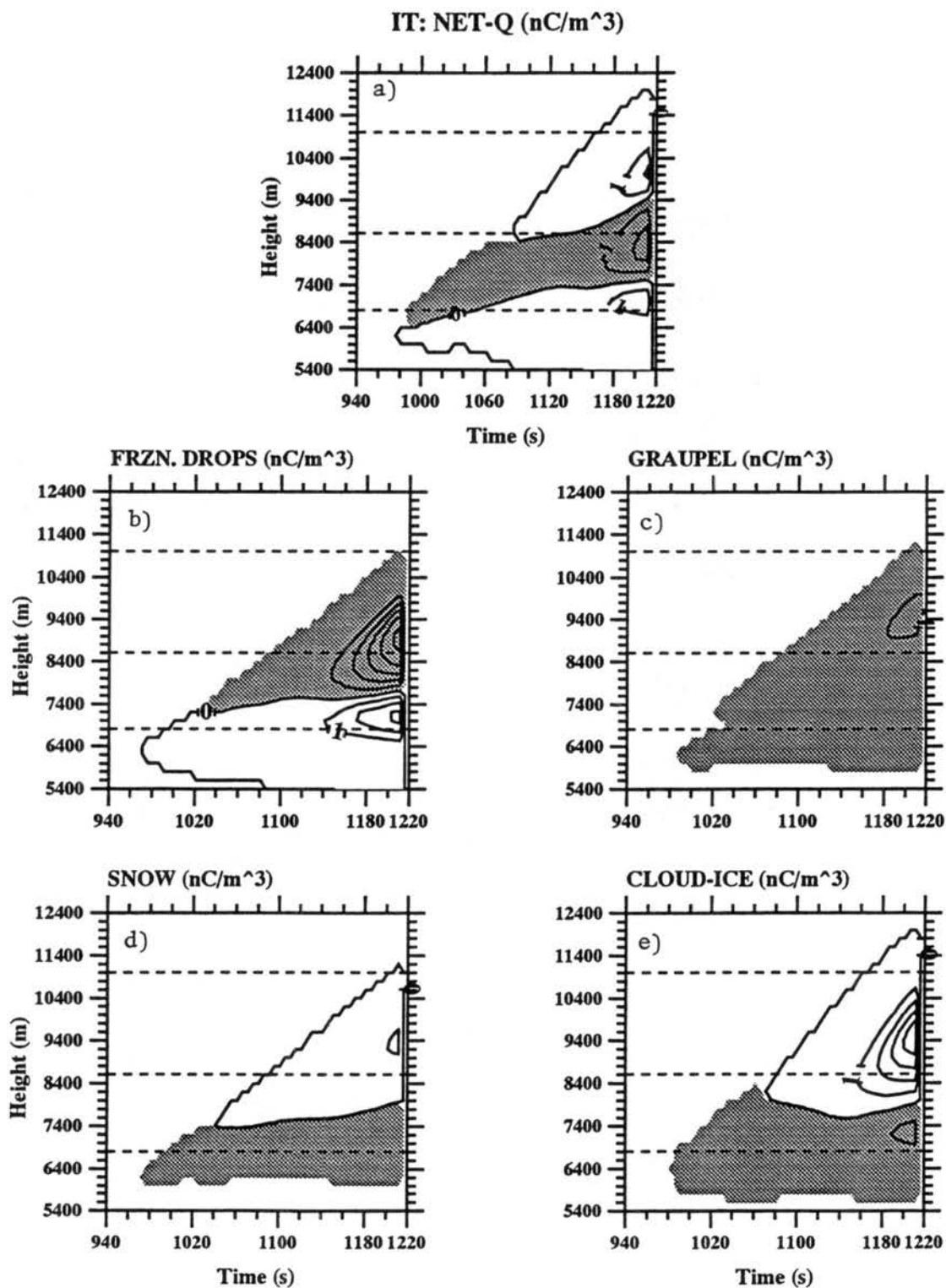


Figure 5.18 As in Fig. 5.8 but IT simulation. Charge density (nC m^{-3}): a) net; b) frozen drops; c) graupel; d) snow; e) cloud-ice.

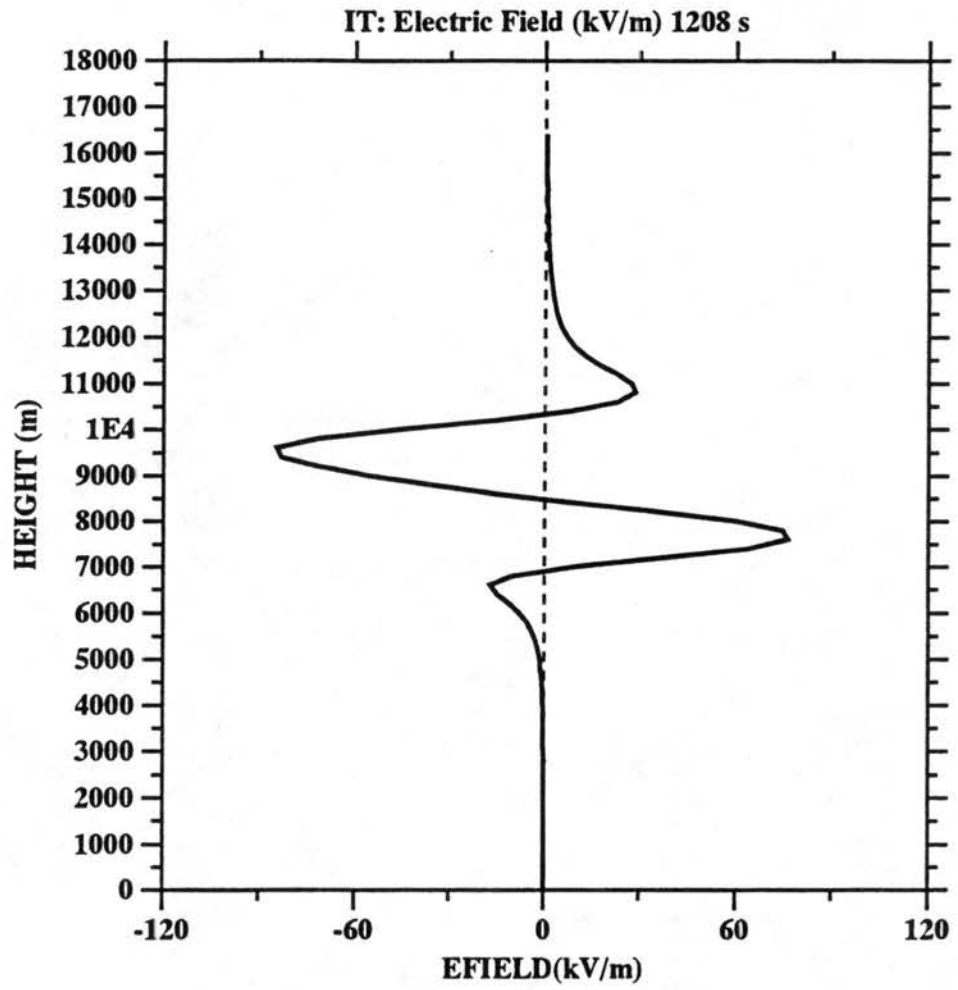


Figure 5.19 As in Fig. 5.9 but IT simulation electric field at $t = 1208$ seconds.

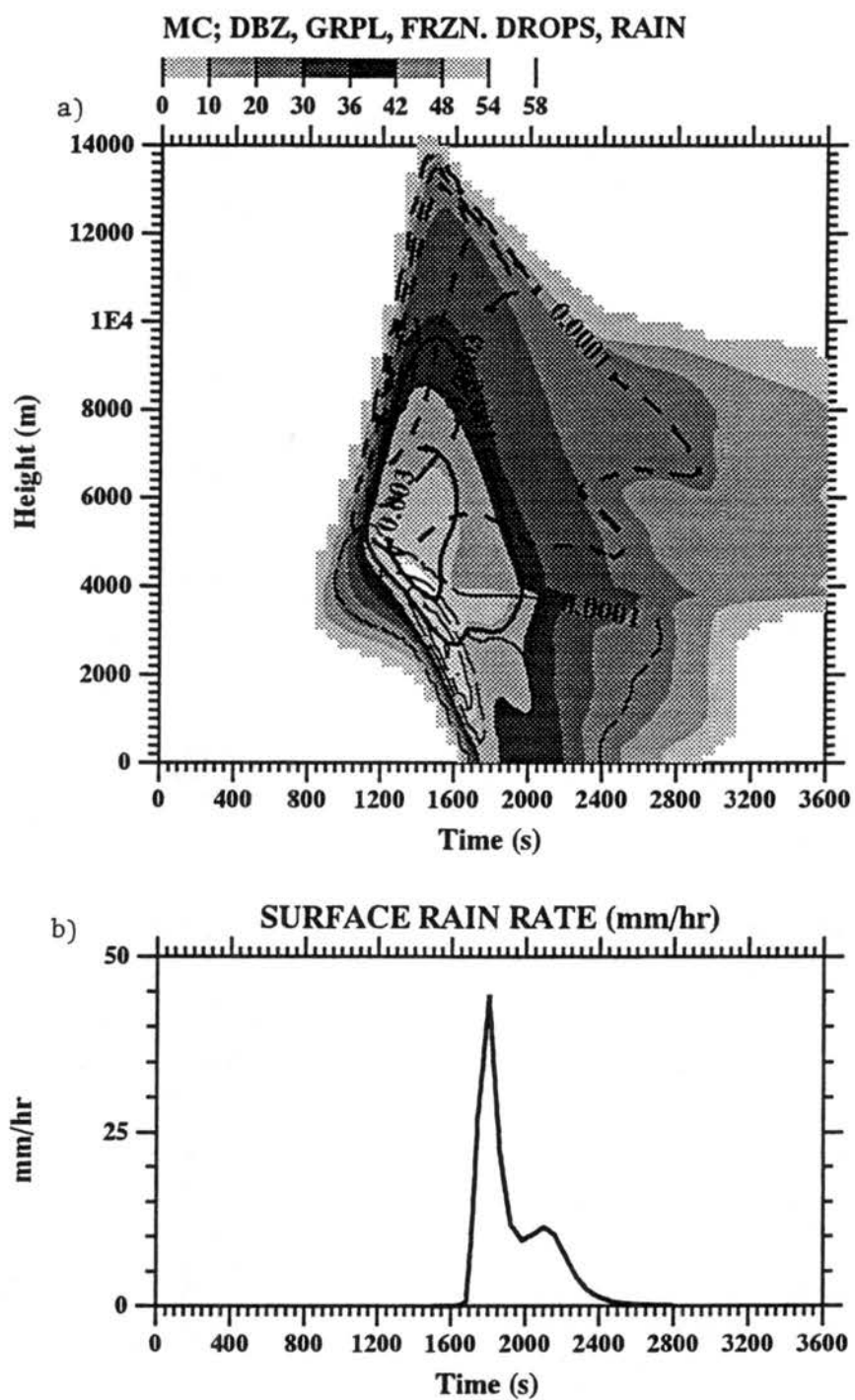


Figure 5.20 As in Fig. 5.5 but MC simulation: a) radar reflectivity, mixing ratios (kg kg^{-1}) of graupel, frozen drops and rain; and b) surface rainfall rate.

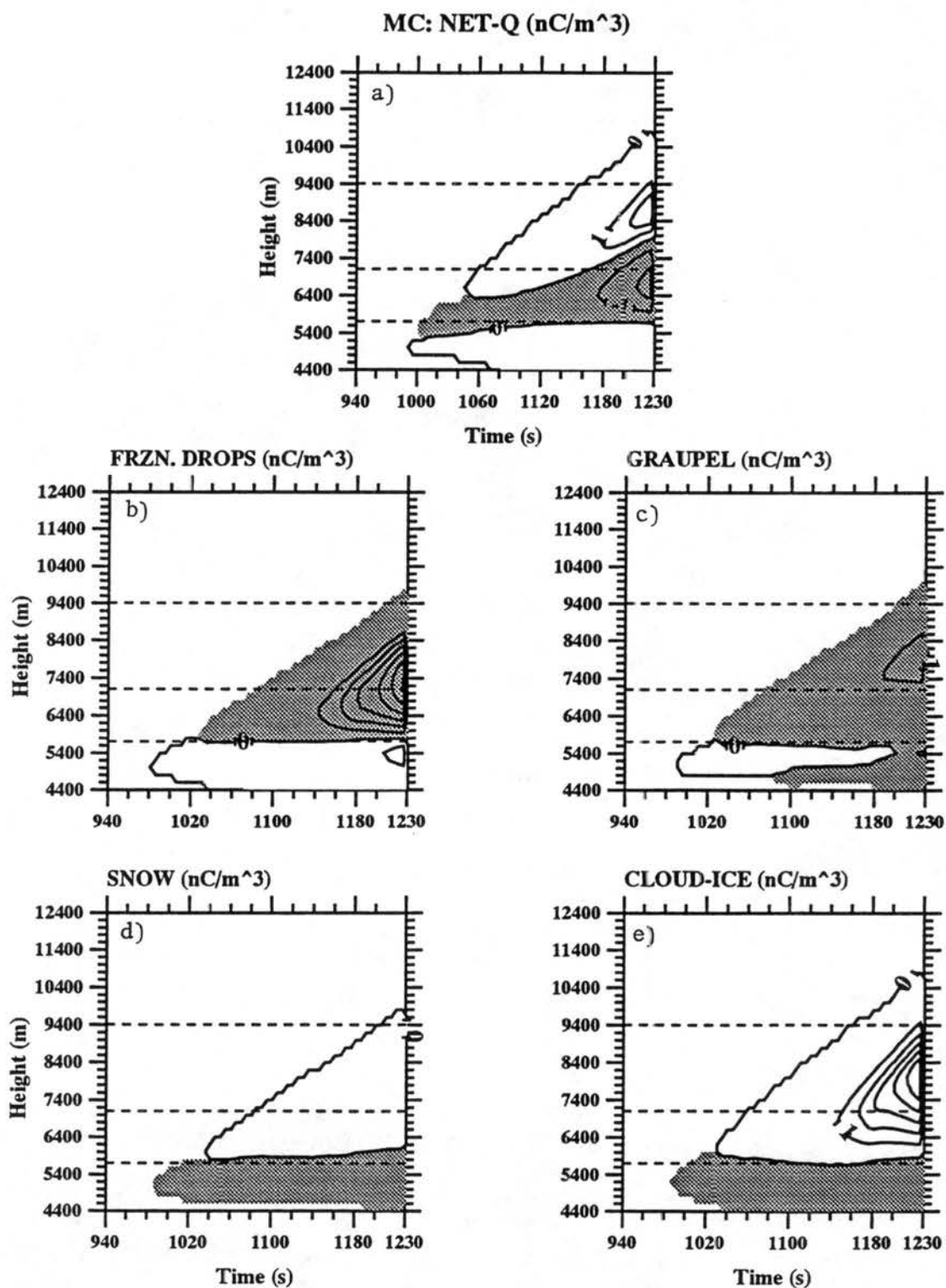


Figure 5.21 As in Fig. 5.9 but the MC simulation. Charge density (nC m^{-3}): a) net; b) frozen drops; c) graupel; d) snow and e) cloud ice.

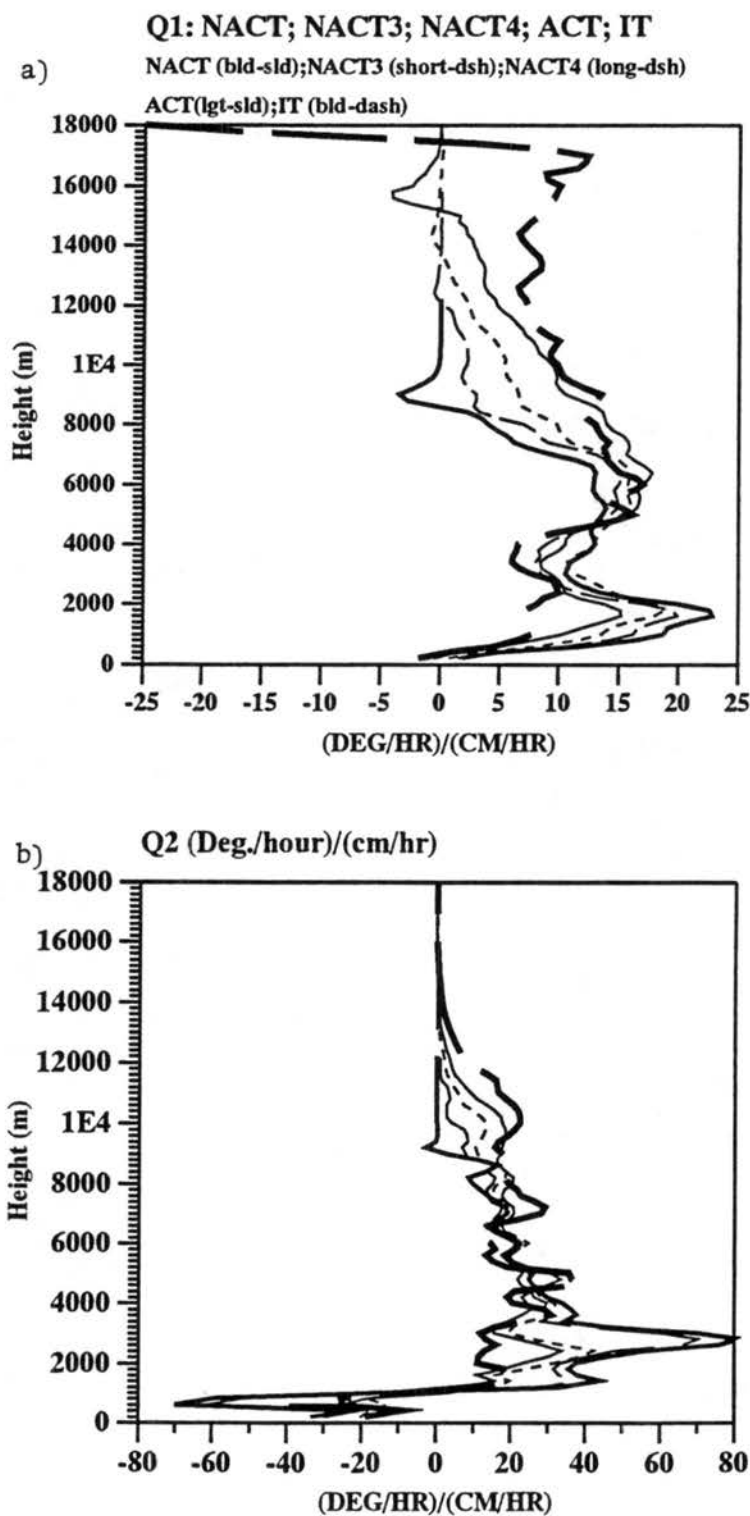


Figure 5.22 NACT (bold-solid), NACT3 (short dash), NACT4 (long-dash), ACT (light-solid) and IT (bold dash) simulation a) diabatic heating rate profiles ($^{\circ}\text{C hr}^{-1} / \text{cm hr}^{-1}$) as a function of height (km); and b) same as (a) but drying rate. The profiles for each simulation are normalized by the mean hourly rainfall rate in cm hr^{-1} .

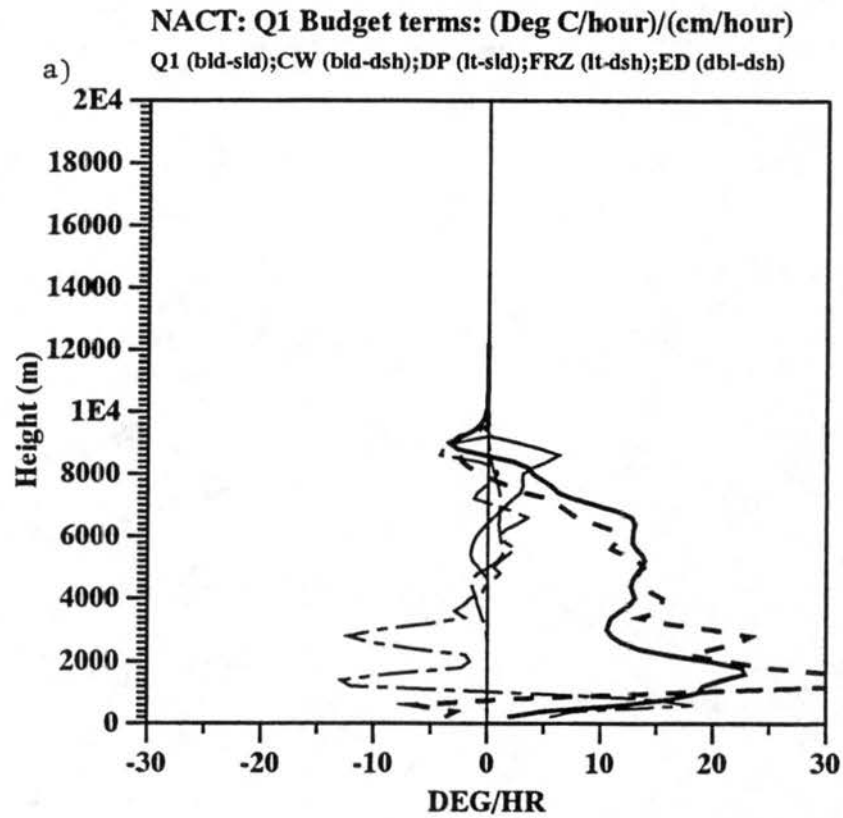


Figure 5.23 a) NACT simulation diabatic heating rate budget ($^{\circ}\text{C hr}^{-1} / \text{cm hr}^{-1}$). Total diabatic heating (bold-solid), condensation of water (bold dash), depositional growth of ice (light-solid), freezing and/or melting of ice (light-dash), and eddy fluxes (alternating long/short dash). The profiles are normalized by the mean hourly rainfall rate in cm hr^{-1} .

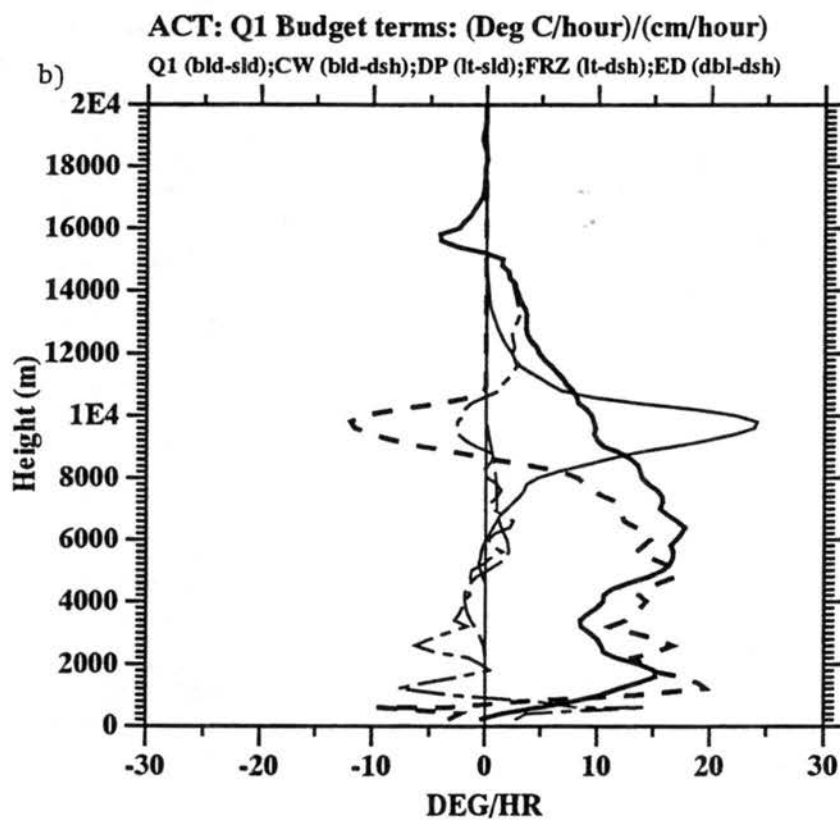


Figure 5.23 b) As in (a) but the ACT simulation.

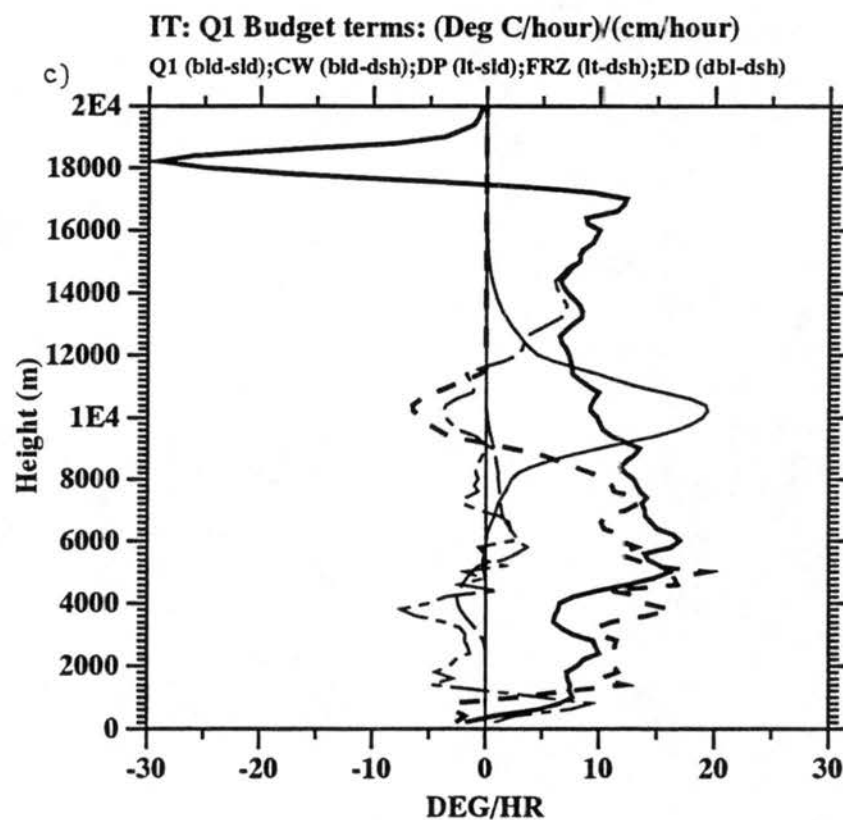


Figure 5.23 c) As in (a) but the IT simulation.

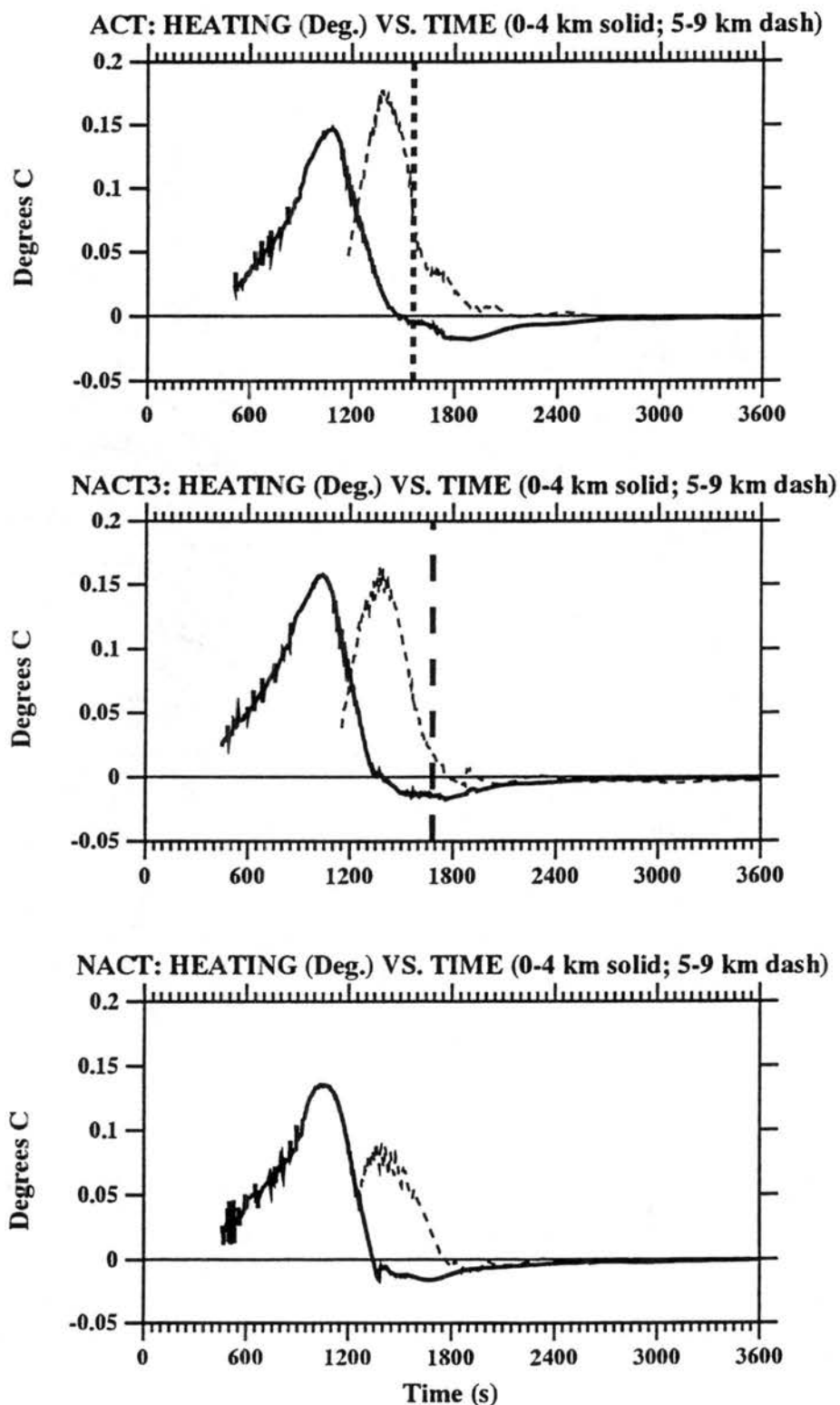


Figure 5.24 Time series of latent heating ($^{\circ}\text{C}$) between the 0-4 km levels (solid curve) and 5-9 km levels (dashed curve) for the a) ACT; b) NACT3 and c) NACT simulations. The vertical bold-dashed line indicates the time of electrical break down.

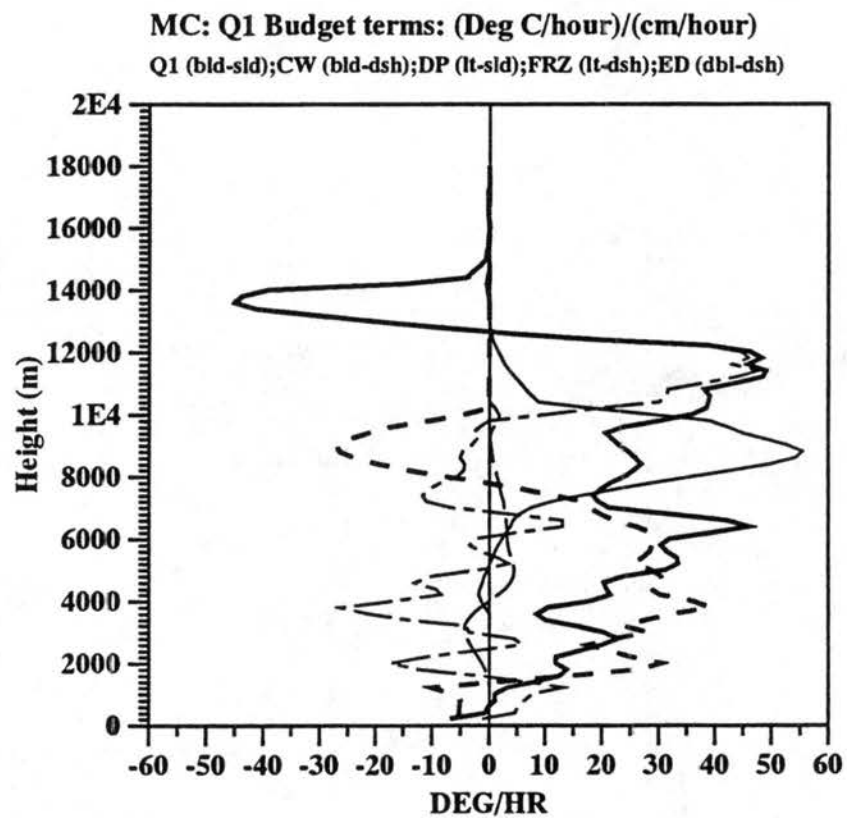


Figure 5.25 As in Fig. 5.23 but for the MC simulation.

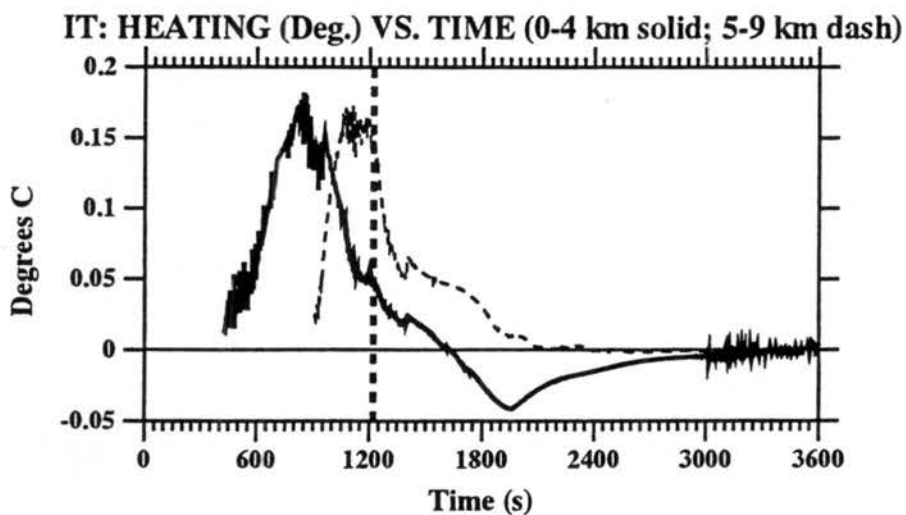
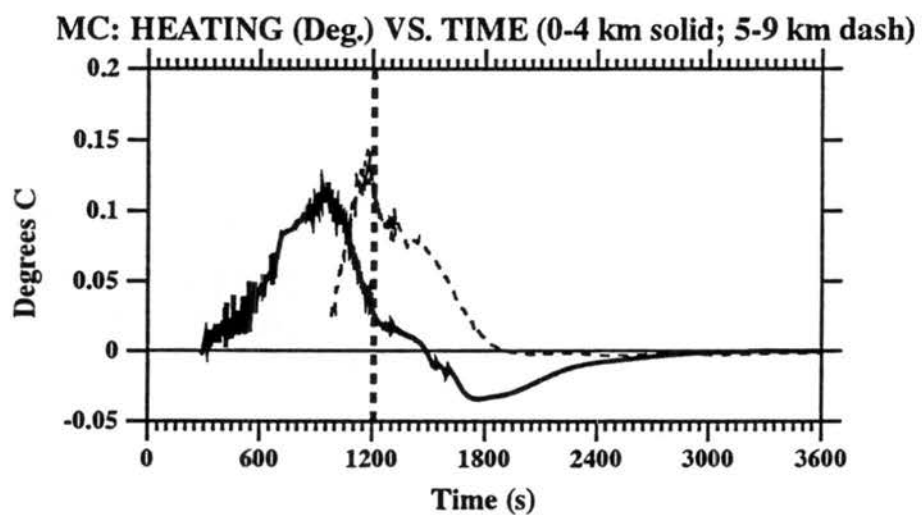


Figure 5.26 As in Fig. 5.24 but for the a) MC and b) IT simulations.

Vertical Velocity vs. Terminal Fall Speed (model results/observations)

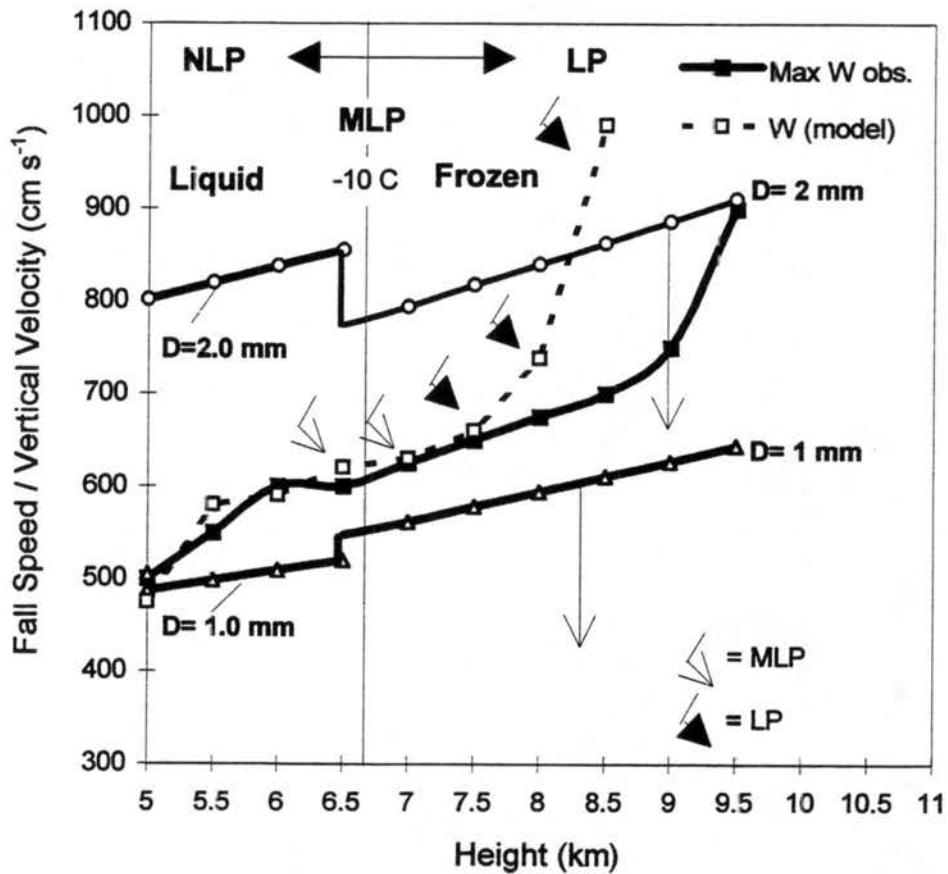


Figure 5.27 As in Fig. 4.23, but mean vertical velocities across the top one kilometer of 36 dBZ reflectivity cores in several model sensitivity tests are included (indicated by dashed line with box markers). Light “crooked” arrows indicate Marginal Lightning Producing simulations (MLP) and dark “crooked” arrows indicate Lightning Producing (LP) simulations. NLP refers to Non Lightning-Producing convection. The observed *mean* vertical velocities from COARE were removed from this plot, observed maximum velocities (Max. W obs.) were retained for comparison.

CHAPTER 6

ON THE RELATIONSHIP BETWEEN CLOUD-TO-GROUND LIGHTNING AND SURFACE RAINFALL

"Outputs" of a convective system such as rainfall and lightning (Fig. 1.1) can be observed, measured and counted. Both of these "outputs" are explicitly related to kinematic and thermodynamic variables associated with the parent convective system that are difficult to accurately measure (e.g., vertical velocity, microphysical characteristics, latent heating; Fig. 1.1, Chaps. 4-5). In turn, "difficult to measure" quantities such as latent heating, are important to the accurate initialization of global numerical weather prediction models (e.g., Puri, 1987). Therefore, it is of great interest to investigate and/or quantify the relationship between two measurable "outputs" associated with deep convective clouds such as rainfall and lightning flash density. The quantitative relationship between the "outputs" can then be used to gain further qualitative, and/or quantitative insight, into parameters associated with the convective rainfall process described in Fig. 1.1. In this chapter we will demonstrate that cloud-to-ground (CG) flash densities (as opposed to the total lightning flash density), can be quantitatively related to the amount of rainfall produced in a convective system, under certain circumstances. Furthermore, since rainfall is directly proportional to the net latent heating in a column of the atmosphere, the CG lightning flash density is related to the column integrated latent heating (e.g., Goodman and

Christian, 1993). As a complimentary effort the modeling simulations of Chapter 5 dealt more specifically with the shape of the heating profile as it relates to the convective/electrical intensity.

Many previous studies have examined the relationship between rainfall and lightning flash density, several of which were described in Chapter 1 (e.g., Table 1.1). To briefly summarize those observations: i.e., studies of *storm-scale* rainfall and lightning (Table 1.1) have reported reasonably good correlations (e.g., 0.7-0.9) between total rain flux, and CG (or total) lightning flash density over lifecycles of individual convective events (or daily time-scales; e.g., Williams et al., 1992). However, rain-yields (defined as the ratio of rain mass to the number of lightning flashes [total or CG's] observed in a specified area) computed over these small temporal and spatial scales exhibited a great degree of variability, sometimes differing by a factor of 10-100. The case-by-case variability of the rain-yield has been clearly illustrated by Cheze and Sauvageot (1997). Cheze and Sauvageot found that when lightning and rainfall data from several individual convective events (each exhibiting high correlations between rainflux and flash density) were processed together, correlations between rainfall and flash density were significantly reduced. However, it was also suggested that the variability observed in "case-by-case" comparisons of the rain-yield might be reduced if the lightning-producing convection occurred under relatively "homogeneous" climatological conditions (i.e., convective "regime"). Indeed, the importance of *rainfall regime* as it relates to the variability of *tropical* rain-yields was noted by Williams et al. (1992) for tropical monsoon and monsoon-break rainfall regimes in the maritime continent region of northern Australia. To place the results of this study in the proper framework with regard to rainfall-regimes, it is beneficial to consider in some detail the relationship between rainfall regime and the rain-yield, especially for regions of the tropics.

6.1 The importance of rainfall regime in the interpretation of rain-yields

Based on observations and modeling sensitivity studies presented in Chapters 4-5, and on previous observations in the tropics, it is clear that an important exception to the “multiscale interaction” analogy discussed in Chap. 1 (Fig. 1.1), is the occurrence of convection in the tropics or sub-tropics that is often associated with heavy rainfall but little, if any lightning (e.g., Maier et al. 1978; Rutledge et al., 1992; Zipser and Lutz, 1994; Zipser, 1994; Chap. 4, Figs. 4.11-4.13; Chap. 5, NACT control and sensitivity simulations). Observations of heavy convective rainfall in weakly electrified tropical convection (particularly oceanic or monsoon convection; e.g., Rutledge et al. 1992; Rickenbach, 1996; DeMott, 1996; Chap. 4 in this study) are examples of the decoupling of rainfall and cloud-electrification processes.

To a first order, observations indicate that the presence of large quantities of precipitation-sized ice are necessary before a cloud is able to produce lightning (e.g., Chaps. 4 and 5; also Williams and Lhermitte, 1983; Dye et al., 1989; Goodman et al., 1988; Ramachandran et al., 1996). In order for convection to initiate an active ice process in which precipitation sized ice particles are produced in a mixed phase environment, a significantly strong updraft must first exist (e.g., Table 5.4b). Hence, the initiation of a vigorous ice phase (and by implication, the electrification phase) depends on the magnitude of the updraft, which from thermodynamical and dynamical considerations, is a function of both the low level forcing, and the magnitude and the shape of the vertical profile of CAPE (Convective Available Potential Energy). The observations presented in Chapters 3 and 4, the model simulations in Chapter 5, and previous observations such as Rutledge et al. (1992), Williams et al. (1992), Williams and Renno (1993), Lucas et al. (1994b), and Williams (1995), all suggest that differences in the magnitude and “shape of the CAPE”, between oceanic and continental storm environments point to a more robust generation of precipitation sized ice mass (which later melts while falling to the surface to produce rainfall)

in continental convection compared to oceanic convection. Indeed, the model simulations discussed in Chapter 5 demonstrated that the contribution of ice-mass to storm integrated rainfall amount as a function of convective regime and the thermal buoyancy profile increased uniformly from a low of 21% in the NLP oceanic convection, to 34% in lightning-producing oceanic convection, and to 42% in highly electrified tropical island convection. In the simulation of mid-latitude continental convection, 75% of the rainfall was associated with melting ice.

The relatively small contribution of ice-mass to total rainfall in tropical oceanic clouds illustrates the fact that rainfall processes do not always require the presence of ice if warm-rain collision/coalescence processes are active, which is often the case in tropical convection (e.g., Young, 1993). A combination of reductions in updraft speed owing to the vertical profile of thermal buoyancy, water loading, and entrainment (e.g., Lemone and Zipser, 1980; Jorgensen and LeMone, 1989; Williams and Renno, 1993; Lucas et al. 1994a,b; Williams, 1995; DeMott, 1996) combined with lower cloud condensation nuclei concentrations, increased depth of the cloud below the freezing level, and higher moisture contents in the lower tropical atmosphere, combine to enable warm-rain microphysics in the tropics to initiate rain in a cloud well prior to the initiation of ice (e.g., first radar echos are observed at temperatures above freezing prior to the clouds developing an ice-phase; see Fig. 5.10a; Young, 1993). The modeling sensitivity studies discussed in Chapter 5 suggest that the formation of rain in the lower region of a non-lightning producing tropical cloud “short circuits” the generation of large quantities of precipitation ice (but not smaller ice-crystals, which are still nucleated efficiently in the water saturated environment and produced by ice-multiplication) by depleting the weaker, water loaded updraft of necessary cloud and rain water contents. Hence, rain can fall from tropical oceanic convection with or without the initiation of an ice-based precipitation process, and subsequently

with or without the generation of lightning in the cloud (e.g., Rutledge et al. 1992; Williams et al. 1992; Zipser and Lutz, 1994).

Though the concept of the rain-yield is not dependent on any particular charging mechanism being operative in a cloud (e.g., convective, non-inductive, inductive etc.), it is worth considering the processes which lead to the electrification of a cloud relative to the previous discussion of rainfall processes. To the extent that a precipitation based charging mechanism (e.g., non-inductive processes; Reynolds et al., 1957; Takahashi, 1978; Jayaratne et al., 1983; Williams and Lhermitte, 1983; Dye et al., 1989; Williams et al., 1989,1992; Saunders et al. 1991; or inductive processes; Gaskell, 1981; Brooks and Saunders, 1994) is responsible for electrifying a cloud, we might expect a weak positive, zero, or even negative correlation to exist between tropical oceanic or monsoon rainfall and the amount of lightning observed. *The correlation would be entirely dependent on the degree to which the ice phase was able to participate in the production of convective precipitation in the cloud* (e.g., Chapter 5). Conversely, a much higher correlation should exist between rainfall and lightning in continental convection (tropical or mid-latitude) where a larger fraction of the precipitation production is driven by ice-based processes (e.g., Knight et al., 1974; Young, 1993; Figs. 5.5, 5.10, 5.14, Chapter 5). Based on this argument we might expect to see a spectrum of rain-yields (correlation coefficients) that would be bounded by high (low) values for oceanic convection and low (high) values for mid-latitude continental convection.

To this point we have based our regime-dependence discussion on the a priori assumption that precipitation based charging is responsible for the electrification, and subsequent lightning production in convective systems. However, even if we consider the convective theory of cloud electrification forwarded by Grenet (1947) and Vonnegut (1953), we can invoke similar, but more dynamical arguments to explain the observed differences in the relationship between

lightning and rainfall for both oceanic and continental convection. If differences in planetary boundary layer ion concentrations are neglected (in order to concentrate on the dynamical effect) in the case of convective electrification, we might expect a stronger coupling between rainfall and lightning production to be found in storms with stronger/larger updraft cores (from a particle number concentration/ion flux perspective). Based on updraft observations and thermal buoyancy profiles ("the shape of the CAPE") in both continental and oceanic/monsoon convection (e.g., Fig. 3.12 herein; LeMone and Zipser, 1980; Lucas et al. 1994a,b; Williams, 1995, DeMott and Rutledge, 1996), this suggests that a higher correlation between lightning and rainfall should exist in continental type convection (in the mean).

Since deep, electrified convection often occurs as a component of mesoscale convective systems (MCS's) we must also consider what proportion of the total rainfall will be associated with MCS stratiform precipitation. MCS's are large [$O(10^4 \text{ km}^2)$], raining, cloud systems that consist of deep convective cells (~90 % of the total CG lightning), and extensive regions of stratiform cloud which produce light precipitation (~10% of the total CG lightning; e.g., Rutledge and MacGorman, 1988). The reduction in CG lightning typically associated with stratiform precipitation in an MCS would have the effect of inflating the postulated "*convective*" rain-yield if the stratiform rainfall component dominated the total rainfall. Water budget studies (e.g., Gamache, 1982; Rutledge and Houze, 1987; Simpson et al. 1993) and observations (Gamache, 1982, Simpson et al., 1993, Cifelli and Rutledge, 1994, Rickenbach and Rutledge, 1996) suggest that this is not the case however. In an MCS, an average of approximately 30% of the total surface rainfall is associated with stratiform rain. Of this 30%, the water budget studies cited above suggest that approximately 20-50% of the water mass originates as detrained ice and liquid condensate produced in the deep convection. Therefore, it seems reasonable to

assume that *at least* 80% of the total surface rainfall (in the mean) associated with an MCS has its origin in deep convection.

In this study we will examine in detail the relationship between warm-season rainfall production and cloud-to-ground lightning (the “rain-yield”) over *large spatial* ($\sim 10^5 \text{ km}^2$) and *temporal scales* (~ 1 month) for *several different regions of the globe*. Cloud-to-ground lightning is used in the calculation of the rain-yield instead of total lightning flash density primarily because of the availability of the data. However, there is a reasonable basis for expecting rainfall and CG lightning to be well correlated, as observations have repeatedly demonstrated that a close coupling exists between the initiation and frequency of CG lightning and the coincident descent of hydrometeors through the lower mixed phase region of convective clouds which ultimately produces heavy rainfall at the surface (e.g., Fig. 4.10; also Workman and Reynolds, 1949; Goodman et al., 1988; Williams et al., 1989; Carey and Rutledge, 1996a). Indeed, more recent studies such as Cheze and Sauvageot (1997) and Carey et al. (1997) have reported (on a case-by-case basis) a *higher correlation* between CG lightning and rainfall, than total lightning and rainfall. In the future, it is anticipated that more comparative rain-yield analyses will be completed using total lightning flash density (which includes the dominant intracloud component) when total flash data become available (e.g., via total flash counting ground based VLF networks, TRMM-LIS or Optical Transient Detector [OTD] data).

6.1.1 A summary of the method used for computing rain-yields

The methods developed and data used for computing the rain-yields for different regions of the globe were discussed in Chapter 2 (Secs., 2.1.5, 2.1.6, 2.2, 2.3, and 2.4.3). We can summarize the technique as follows: i.e., warm season rain-yields (rain-mass divided by CG

flash number) were computed for large temporal (~ 1 month) and spatial ($\sim 10^5 \text{ km}^2$) domains, and for a broad sample of rainfall climatologies (Figs. 2.1 and 2.6a-b). The selection of large areas and long time scales (relative to that of a single convective cloud) reduces standard deviations in the rain-yields for different regions by providing substantial filtering of cloud scale variability. This scale selection differs significantly from many previous studies that have examined rain-yield relationships for individual thunderstorms or convective systems (e.g., Table 1.1).

The two components of the rain-yield, rainfall and CG lightning flash count, were collected using raingauge, radar, and satellite estimates of rainfall, and lightning location network data for CG lightning counts. No attempt was made to correct the flash counts for differences in detection efficiency that may have existed between the networks. We estimate that the error introduced into the rain-yield calculation by potential differences in network detection efficiency is less than a factor of two. Mean monthly rainfall was computed for the same time periods (summer convective seasons) and areas that the lightning data were collected using network raingauge data (U.S.- cooperative gauge network; Australia- TRMM ground truth network) of the highest resolution possible (gauge densities in all networks were $O[10^{-3} - 10^{-4} \text{ km}^{-2}]$). Tropical continental and island rainfall that occurred in the vicinity of northern Australia were further partitioned into monsoon (oceanic, weakly electrified) and continental break-period (tropical continental, highly electrified) convective rainfall regimes.

Rainfall estimates for the W. Pacific Ocean (COARE areas 1, 2, and 3; Fig. 2.6b), were computed using a carefully tested radar reflectivity-rainfall relationship for area COARE-1 (e.g., Short et al. 1995), and a satellite based monthly rainfall estimate over $2.5^\circ \times 2.5^\circ$ grid squares (from the Global Precipitation Climatology Project; e.g., Janowiak and Arkin, 1991) for areas COARE-2 and COARE-3 (Fig. 2.6b). In addition to the raw data used in our computation of the rain-yields, a single rain-yield data point representing a 5-year warm-season mean for Gabarone,

Botswana (24°S 26°E) was computed by Dr. E. R. Jayaratne, (University of Botswana) and is used in our analysis for comparison.

The method we have chosen for relating lightning to rainfall is similar in principal to algorithms developed previously in the satellite and radar communities for estimating precipitation using variables such as infrared (IR) brightness temperature (e.g., Arkin, 1979; Richards and Arkin, 1981), or the area-time integral (ATI) of radar reflectivity (e.g., Doneaud et al. 1984; Lopez et al., 1989). The aforementioned satellite and radar-based precipitation algorithms estimate rainfall by exploiting high positive correlations observed between rainfall (usually measured by raingauges) and variables such as the area of cloud IR brightness temperature colder than 235 K, or areas of radar reflectivity greater than some locally determined reflectivity threshold. The satellite techniques produce the highest correlations when used over large spatial and temporal scales (e.g., Richards and Arkin, 1981). For example, satellite based estimates of precipitation use IR temperature data and are computed over monthly and/or seasonal time-scales and areas of $\sim O(10^5) \text{ km}^2$ (e.g., the GOES Precipitation Index; GPI; Arkin, 1979). The ATI on the other hand can be used on smaller temporal and spatial scales (Lopez et al. used 12 hours and an area of $3.6 \times 10^4 \text{ km}^2$), but relies on a “well-behaved” climatological description of rainfall such as the rain-rate probability density function for different geographical locations (Atlas et al. 1990). Similar to the satellite (GPI) and radar algorithms (ATI) developed for precipitation estimation, we use large temporal and spatial scales in an attempt to force “well-behaved” statistics while retaining important physical information about any underlying “climatological” relationships that exist between rainfall and CG lightning.

It is important to note that the units of the rain-yield (kg/fl) *do not imply that a unique amount of rain mass is associated with an individual lightning flash*. Implicit in our computation of warm season “convective” rain-yields (e.g., summer in the U.S.) is the

assumption that the majority of the rainfall is associated with deep convection. Since our sample of rainfall is undoubtedly contaminated by cloud systems associated with little or no lightning, (e.g., tropical-like rain showers in the southeastern U.S., stratiform region precipitation produced by MCS's), or anomalous deep convective events such as supercell thunderstorms that produce copious amounts of IC lightning and rainfall but few CG flashes during some portion of their lifetime (e.g., Carey and Rutledge, 1996b), *the rain-yield should only be interpreted as an area normalized ratio that indicates a more general relationship between deep convection, lightning, and rainfall.*

6.2 Results

The results of the rain-yield calculations are presented using two approaches (Figs. 6.1, 6.3; Tables 6.1-2). First, we present rain-yield results computed from the summer convective season (June-August) of 1994 in the continental U.S., and then for comparison, rain-yields computed for other regions of the globe (Table 6.1, Fig. 6.1). Next, based on the correlation between lightning flash density and rainfall (Table 6.1), we present "predicted" rainfall for selected areas of the U.S. from the Summer of 1995 (Fig. 6.3; Table 6.2) using the product of CG flash densities from the summer of 1995 and rain-yield values from 1994. Presentation of the results in this fashion allows a demonstration of both the *stability* and *utility* of the rain yield.

6.2.1 Rain-Yields as a function of location on the Globe

Seasonal mean rain-yields for areas in the continental U.S. (Summer 1994), tropical northern Australia (including monsoon and break period data points for the continent, and

Melville and Bathurst Islands), and the tropical western Pacific Ocean (TOGA COARE) are summarized in Table 6.1 and plotted in Fig. 6.1. Seasonal mean rain-yields (Fig. 6.1) for the central and southeastern portions of the continental U.S. converged to a remarkably stable value of $\sim 10^8$ kg/fl (Table 6.1). The stability of the rain-yield in the continental U.S., as indicated in Fig. 6.1 by rain-yield values that closely follow the 10^8 kg/fl rain-yield contour, bears some resemblance to the results of Richard and Lojou (1996) who examined the convective-rainfall and total lightning flash densities associated with approximately 60 thunderstorms that occurred over France (Fig. 6.2). Further, the rain-yield for the southern hemisphere continental interior location of Gaborone, Botswana (subtropical Africa) was for all intensive purposes identical to the yields calculated for the central and southeastern U.S.

Table 6.1: Rain-yields and rainfall-lightning correlation coefficients.

Regime	Rain-yield (kg/flash)	Rain - CG Flash Count Correlation Coefficient (r)
Mid-Latitude Cont.:		
Arid (SW U.S.)	5.7×10^7	0.89
Mid-continent U.S.	1.1×10^8	0.87
Humid SE U.S.	1.3×10^8	0.71
Northern Continent	2.5×10^8	0.45
S. Hemisphere Sub-Tropical:		
Gaborone, Botswana	1×10^8	NA*
Tropical:		
DWN-break	3.4×10^8	NA*
DWN-Monsoon	2.0×10^9	NA*
Island break	8.0×10^8	NA*
Island monsoon	7.7×10^9	NA*
Island-Ocean	2.3×10^9	NA*
Tropical Ocean	$1-2 \times 10^{10}$	NA*

*Not enough data points to form the correlation on a monthly basis.

Rain-yields in the “arid” southwestern U.S. (Fig. 6.1) can be characterized as having the lowest value, with a mean of 6×10^7 kg/fl (Table 6.1) and a range from 5×10^7 kg/fl to 8×10^7 kg/fl. Interestingly, the mean rain-yield for the arid regime compares very favorably to the

ensemble mean rain-yield range of 3×10^7 kg/fl to 6×10^7 kg/fl calculated by Battan (1965) for thunderstorms in Arizona (Table 1.1). For latitudes greater than 40° N (Northern Continent in Fig. 6.1 and Table 6.1), the computed rain-yields were generally higher than in other regions of the continental U.S. with values of $2\text{--}3 \times 10^8$ kg/fl. The higher values of rain-yield computed for the northern latitudes (relative to the rest of the U.S.) reflect an inflation of the assumed “convective” rain-yields by non-convective frontal precipitation.

In the tropics, rain-yields for continental northern Australia (DWN) were similar to the daily values computed by Williams et al. (1992), and ranged from $\sim 3 \times 10^8$ kg/fl for break period (tropical continental) convection, to 2×10^9 kg/fl for monsoon convection (tropical oceanic). Rain-yields for break and monsoon period convection that occurred over Melville and Bathurst Islands approached values of 8×10^8 kg/fl and 8×10^9 kg/fl for break and monsoon periods respectively. The largest rain-yields were observed for regions of the tropical W. Pacific Ocean (similar to the RTR results of Zipser, 1994 for the E. tropical Atlantic), where values approached $\sim 10^{10}$ kg/fl, nearly two orders of magnitude larger than that of the continental yields (mid-latitude or tropical), one order of magnitude larger than the island break-period yield, and a factor of 2–5 larger than the island and continental monsoon yields respectively. Interestingly, the mixed “island-ocean” rain-yield computed for the western Pacific (satellite based rainfall estimate) exhibited a value that was somewhat intermediate to the Melville and Bathurst island “monsoon” and “break” period rain-yields. On a physical basis, this result might be anticipated since deep convection occurring in a region comprised of both islands and open ocean is likely to be comprised of both oceanic (less lightning) and island-like rainfall systems (much more lightning; as observed visually by one of the authors, SAR). However, the possibility that this result is an artifact of computing rain-mass using two different measurement techniques (e.g.,

satellite for the mixed island-ocean rainfall vs. rain-gauges for the island rainfall) cannot be discounted (e.g., Ebert et al., 1997)

The mean rain-yields (Table 6.1) for each separate rainfall regime in Fig. 6.1 vary in a physically consistent fashion. As an illustration, consider the rain-yields for two relative extremes; the arid southwestern U.S. and the humid tropical oceanic environment. The $10^2 - 10^3$ increase in rain-yield associated with a climate-regime transition from arid continental to tropical-oceanic is consistent with the characteristically different cumulonimbus cloud morphologies associated with each regime. Recall that deep, tropical oceanic cumulonimbi are typically associated with weaker updrafts (e.g., LeMone and Zipser, 1980; Lucas et al. 1994a), low flash rates (e.g., Rutledge et al., 1992; Zipser, 1996), and copious amounts of rainfall (e.g., Zipser, 1994; DeMott, 1996; Rickenbach, 1996). The net effect of these more “oceanic” convective characteristics is to substantially increase the rain-yield. Conversely, during the summer in the arid climate of the southwestern U.S. (e.g., Figs. 4.16-4.18), precipitation processes are dominated by ice-based precipitation mechanisms (Fig. 4.16) and the rainfall is accompanied by considerably more CG lightning (Fig. 4.18; c.f., Reap, 1986; Smith and Gall, 1989). Enhanced CG lightning flash counts coupled with potentially significant evaporation of rainfall in the dry sub-cloud layer air of the southwestern U.S. (e.g., recall the strong temporal correlation between CG flash count and macroburst intensity; Fig. 4.18) may explain the reduction in the arid rain-yield relative to *any* of the other regions examined. Therefore, the analysis of rain-yield results for the two rain-type extremes suggests that increases in the rain-yield will occur for rainfall regimes and precipitation processes that are more oceanic in nature (indicated by “Maritime” in Fig. 6.1), and decreases in the rain-yield will occur for regimes that are described as “Arid” (Fig. 6.1). In this respect, the behavior of the rain-yield is analogous to that of the RTR (Zipser, 1994).

Now compare the value of the tropical continental “DWN-break” rain-yield (3.4×10^8 kg/fl) to that of the continental U.S. and Botswana (mean of $\sim 10^8$ kg/fl; Fig. 6.1). The DWN-break rain-yield is a factor of 3.4 higher than the mid-latitude continental values. If the mean daily rainfall for a specific mid-latitude data point such as sub-tropical Florida (“SE humid” data points in Fig. 6.1; ~ 6 mm day⁻¹) is compared to that of the DWN-break rainfall (6.2 mm day⁻¹), we find that differences in rainfall between the two locations do not explain the factor of 3.4 increase in the DWN-break rain-yield relative to Florida (or to the other mid-latitude yields since they are all similar to Florida). However, if the CG flash densities are examined we find that the mean CG flash density for Florida is a factor of 3.5 larger than that of DWN (hence the factor of 3.4 increase in the Darwin rain-yield relative to Florida). If the possibility of measurement error is neglected (in either rainfall measurement or in relative CG flash counts), then the data suggest that for this particular sample, a larger amount of rainfall was produced *per CG lightning flash* over the tropical continent than in Florida.

The reduction in CG flash density between Florida and the DWN-break (associated with an increase in rain-yield for similar rainfall amounts) could be due to differences in detection efficiency between the NLDN and the Darwin-area DF networks, or to a real difference in the total electrical activity observed in the two regions for the time periods sampled (e.g., Fig. 3.11). We cannot quantitatively assess the degree to which differences in detection efficiency existed between the two networks, though we can state that flashes counted for the tropical continental yields occurred in the interior of the Darwin-area network, thus reducing the possibility of detection efficiency errors introduced by range/acceptance criteria. A factor of 3.5 decrease in CG flash density, if due entirely to differences in detection efficiency, would imply that the detection efficiency of the Darwin DF network was less than 30% (if the NLDN detection

efficiency over the Florida Peninsula is assumed to be 70%-90%; Cummins, 1995) which also does not seem likely.

Another possible explanation that could be invoked to explain the observed differences in rain-yield between Florida and the tropical continent, is that Florida may experience persistent, stronger, and more frequent low-level forcing of convection in association with phenomena such as sea-breeze convergence and transient baroclinic weather systems (e.g., stronger synoptic forcing). Additionally, as noted in Chapter 3, the IC-CG ratio may also increase with decreasing latitude (as suggested by Prentice and Mackerras, 1977). This would have the net effect of reducing the CG flash density near Darwin relative to Florida (and hence the rain-yield) for similar total flash densities. Very recently, Hidayat et al. (1997) computed a rain-yield value for the large island of Java (e.g., note that the area of Java is more than an order of magnitude larger than that of Melville and Bathurst Islands combined) that is virtually identical to the mean rain-yield we computed for DWN. Given that Java is located within the Maritime Continent just to the north of DWN, the similarity of the Java and DWN rain-yields further supports the contention of a physically-based difference between the Florida and DWN rain-yields, and serves to further illustrate the basic stability of the mean rain-yield for this region.

Moving toward slightly higher rain-yields in Fig. 6.1, note that the tropical island "break" and "monsoon" regimes had higher rain-yields compared to the nearby tropical continent by factors of 2.5 and 3.8 respectively. Relative to DWN-break, the increase in the island-break period rain-yield was associated primarily with a reduction in the CG flash density (factor of 2.5) relative to that over the tropical continent (Fig. 3.11). The mean rainfall computed for the island-break regime was comparable to that observed for the DWN-break. For the monsoon periods, the CG flash density over the island ($32 \text{ Flashes}/10^4 \text{ km}^2/\text{day}$) was also a factor of 2.5 lower than the continent ($79 \text{ Flashes}/10^4 \text{ km}^2/\text{day}$), but the rainfall over the island (24 mm/day)

was a factor of 1.5 larger than that of the continent (16 mm/day). When the rainfall data were analyzed for the monsoon period over the continent, it was noted that a large gradient in rainfall totals existed from the coastline moving inland. This is consistent with the observed motion and lifecycle of oceanic convective systems during the monsoon which typically move over the continent from the northwest and then dissipate while moving inland (e.g., Keenan and Rutledge, 1992). Hence, the islands situated to the northwest of DWN received a larger proportion of total rainfall from weakly electrified monsoon/oceanic convective systems.

It is difficult to explain the differences observed in the CG flash density between Melville and Bathurst islands and the tropical continental area surrounding Darwin given their close proximity. Some *possible* explanations include:

- 1) The detection efficiency of the Darwin DF network over the islands may have been reduced to 40% of the detection efficiency of the network over the continent. This would reduce the observed differences in rain-yield to an artifact of the DF network characteristics. This seems plausible, but not likely given that the island area was located within 400 km of *all* four of the DF sensors in the network.
- 2) The ocean exerts a stronger influence on cloud microphysical and dynamical properties over the islands than the continent, and hence a larger fraction of the total rainfall over the islands is associated with *non-lightning* producing clouds. This explanation seems reasonable given previous observations of the rainfall and convective lifecycles over both the islands (Figs. 4.2-4.4; also Simpson et al., 1993; Keenan et al., 1994) and the continental area surrounding Darwin (e.g., Rutledge et al., 1992; Williams et al., 1992; Keenan and Rutledge, 1992).
- 3) The morphology of the afternoon convection observed over the islands during the break periods (deep, highly electrified) may be associated with a higher IC to CG ratio than

that of the continent (however, this is speculative). This might lead to similar *total* flash rates but a reduced CG flash density over the islands and a subsequent increase in the rain-yield. This is an interesting possibility. Studies of the convective cloud morphology over the islands suggest the presence of a predominant mode of convection that begins with a gradual progression of shallow, merging convection early in the day leading to the onset of rapid convective development in the afternoon (e.g., Figs. 4.2-4.4). The afternoon development is characterized by very deep (18-20 km), intensely electrified convection (total flash rates exceeding 60 min^{-1} at times; personal communication, Dr. S. Rutledge), with vertical velocities in excess of 30 m s^{-1} in the upper portion of the clouds (e.g., Sec. 5.2.4, Figs. 5.14-18; also Simpson et al., 1993; Keenan et al., 1994). With regard to the character of the rainfall associated with these deep convective systems, water budget studies by Simpson et al. (1993) suggest that approximately 90% of the rainfall is "convective" in character. If the "typical" pattern of highly electrified convection and heavy rainfall over the islands is characterized *primarily* by this type of convection (vs. the *more diverse* sampling of convective clouds over larger continental area) then this explanation is plausible based on similar observations of supercell convection (high vertical velocities in the upper regions of the cloud and elevated regions of charge) over the continental U.S. which produce very high total flash rates, and a large IC-CG ratio (e.g., MacGorman et al., 1989, Marshall et al., 1995; Carey and Rutledge, 1996b).

To summarize Fig. 6.1, the climatological trend of increasing rain-yields moves from lower values over the mid-latitude continental U.S., to values 3.4 times higher over the tropical continent, during monsoon break conditions. As the rainfall regime becomes progressively more maritime in character, the rain-yields continue to increase as the regime transitions from tropical

island-break, to mixed island-oceanic (COARE-3), to tropical monsoon (over the continent and islands), and achieves a maximum in tropical oceanic conditions where the rain-yield has a value of $\sim 10^{10}$ kg/fl. The rain-yields appear to increase in magnitude in a manner consistent with expected trends in the gross characteristics of the convective clouds and rainfall associated with each climate and rainfall regime.

6.2.2 Rain-yields in the continental U.S.

If rain-yields are examined for the continental U.S. (Fig. 6.1), deviations (on the order of 2.5×10^7 kg/fl) from the mean ratio of $\sim 10^8$ kg/fl ($s=0.5 \times 10^8$ kg/fl) can be accounted for by further partitioning the rainfall and CG flash data into continental rainfall regimes. Partitioning the data results in a range of rain-yields and well correlated rain-CG flash relationships as a function of regime (Table 6.1). Correlation's between CG flash density and rainfall (area normalized; Table 6.1) were largest for the "Arid" climate regime where ice-based precipitation processes dominate, and lowest for the "Northern" continental regime where the rainfall and lightning amounts were affected by transient synoptic scale disturbances that produced frontal precipitation without much lightning. As the humidity of the climate regime increases (e.g., the southeastern U.S.), the correlation between CG flash density and rainfall decreases slightly (Table 6.1). One possible explanation for this behavior may be that a mixture of rainfall processes (warm rain- e.g., Koenig, 1963, no lightning; deep convection- copious amounts of lightning; overrunning stratiform precipitation with very little lightning, etc.) is operative in this region and is responsible for a lower, but still reasonable, correlation.

Table 6.1 indicates that the poorest correlation between rainfall and CG flash density occurred for areas situated north of $\sim 40^\circ$ N. The low correlation for the northern latitudes is a

result of episodic, frontal precipitation that produced a significant proportion of the total monthly rainfall but little or no cloud-to-ground lightning. This was indicated when surface synoptic weather maps were examined for the state of Michigan for the months of June and August 1994 (July of 1994 exhibited a rain-yield of 10^8 kg/fl). When several days of precipitation and lightning data associated with frontal activity were removed from the June and August data-sets (leaving precipitation primarily associated with thunderstorms), the rain-yields were recalculated and found to approach the 10^8 kg/fl rain-yield noted for other continental locations in the mid-western U.S. A similar process was conducted for eastern Colorado for the month of June, 1995. Approximately seven days of precipitation and lightning data associated with two frontal-upslope precipitation events that occurred in early and late June 1995 were removed from the rainfall and CG lightning sample. The original rain-yield, computed with the upslope precipitation data left in the sample, was 3.5×10^8 kg/fl. When the upslope precipitation data were removed the yield decreased to 1.8×10^8 kg/fl, much closer to the 1994 mid-continent regime values (Fig. 6.1). Hence the rain-yields are meaningful only when the precipitation for a given location and time period is dominated by convectively-based rainfall.

6.2.3 An inverse approach to evaluating rain-yields

Examining Table 6.1 and Fig. 6.1, it is clear that reasonably stable rain-yields and high positive correlations exist between rainfall amounts and the CG lightning flash density for regions of the southern U.S. (south of $\sim 40^\circ$ N) for the summer of 1994 data. To independently test both the stability (calculated for a single year) and the correlation's associated with the rain-yield results from the summer 1994 season, we have; 1) computed a "predicted" convective rainfall for select areas and months by multiplying the summer 1995 CG flash-density by the

appropriate 1994 rain-yield; and 2) correlated the rain-yield predicted amounts of rainfall with the observed values found in the summer of 1995 NCDC cooperative station monthly rainfall data set. The areas selected were determined by the availability of the data, and the presence or absence of frontal precipitation (note for this particular application, hurricane data points were retained in the sample). Areas significantly affected by non lightning-producing frontal precipitation were identified using daily synoptic weather maps archived at Colorado State University and the National Climatic Data Center. The results of this computation (excluding data points associated with frontal precipitation such as June 1995, eastern Colorado) are summarized in Table 6.2 and shown in Fig. 6.3.

If the "hurricane" data points are neglected, the predicted 1995 rainfall totals shown in Table 6.2 fall within a factor of two of the actual rainfall amounts observed for each of the areas (all $\sim 10^5$ km² or larger) and months indicated. The relationship between the "predicted" and "actual" values is more clearly illustrated in Fig. 6.3. Computing a linear correlation coefficient *for the non-hurricane* data points yields a correlation of 0.95 between the predicted and actual values. This correlation is quite good given that the 1995 predicted rainfall amounts were based on only a single summer of rainfall/lightning data from 1994. Excluding the outlier points associated with hurricanes, the results of Fig. 6.3 suggest that CG lightning flash counts may be a useful tool for diagnosing convectively-based rainfall over large areas and long time-scales during summer convective seasons in the southern continental U.S.

Now, consider the outlier "hurricane" points in Fig. 6.3 associated with months and locations that experienced a significant proportion of the total monthly rainfall from hurricane or tropical storm (TS) rains. For both years in the analysis (1994 and 1995) it was noted that hurricane/TS events caused step-like increases (decreases) in the rainfall (lightning) amounts for

the month, causing marked increases in the rain-yield relative to other months and locations (sometimes by a factor of 5-10; Table 6.2).

Table 6.2: Actual vs. rain-yield predicted rainfall (continental U.S.; 1995).

Location and Month 1995	Actual Monthly Rain mm/day	Rain-Yield Predicted Rain mm/day
Non-hurricane effected		
Florida, July	6.4	6.5
Oklahoma, June	5.7	4.2
Oklahoma, July	2.6	2.9
Oklahoma ⁽²⁾ , August	1.3	0.8
Arizona, June	0.1	0.1
Arizona, July	0.6	0.32
Arizona, August	2.4	1.7
New Mexico, August	1.7	1.5
Nevada, August	0.2	0.1
Ohio, August	3.8	3.0
Texas ⁽²⁾ , August	2.7	3.4
Missouri, August	3.0	2.5
Arkansas, August	0.9	1.1
E. Colorado, July	1.4	0.9
E. Colorado, August	1.0	1.7
Hurricane/ T.S. locations		
Texas ⁽¹⁾ , August	2.5	0.2
Oklahoma ⁽¹⁾ , August	2.7	0.6
Florida, August	8.7	3.3
North Carolina, August	3.8	0.6

Hurricanes can and do produce lightning when making landfall (e.g., Samsury and Orville, 1994) or, for example, when undergoing eyewall intensification (e.g., Molinari et al., 1994), but the increase in rainfall associated with these tropical like systems is apparently totally disproportionate to any increase in lightning activity as reflected in the rain-yield (Samsury and Orville, 1994).

One other way that hurricanes and tropical storms can affect the rain-yield is to make landfall, become quasi-stationary over a region, and simply "rain-out" prior to being entrained into the main synoptic flow (with the rainfall being both stratiform and convective in character).

This was the case for Tropical Storm Dean in early August of 1995. TS Dean moved from the Gulf of Mexico into west-central Texas and Oklahoma and then dissipated after producing heavy rain in western Oklahoma and central Texas. As an example, consider the two data points in Fig. 6.3 labeled OK1 and OK2. The data point OK1 represents the western side of Oklahoma and OK2 represents the eastern side of Oklahoma for the time period of August 1995. It is clear in Fig 6.3 which side of Oklahoma experienced the large enhancement in rainfall and relative reduction in CG lightning associated with moisture from TS Dean (OK1). Interestingly, while western Oklahoma experienced a monthly precipitation that was several inches above normal, eastern Oklahoma experienced drier than normal monthly rainfall¹. Note in Fig. 6.3 that the OK2 1995 data point is located much closer to the regression line (i.e., the rain-yield value is close to the mean "mid-continent" value calculated for 1994), more consistent with the mean rain-yield values calculated for Oklahoma in 1994, and for the months of June and July 1995 (included as points in the regression line and in Table 6.2).

Combining the results of Figs. 6.1 and 6.2 suggests that the rain-yield may be of use in estimating rainfall amounts over regions of the southern U.S. where warm season rainfall is dominated by convectively-based precipitation. Additionally, relative changes in the rain-yield for a given location may be used as a qualitative indicator for major changes in the rainfall regime (similar in approach to that proposed by Zipser, 1994 for the RTR; Table 1.1). The analysis in this section has focused primarily on the potential for application of the rain-yield concept as a rainfall estimator over a mid-latitude continent. An important question arises as to the applicability of rain-yields (and lightning data in general) for estimating rainfall or other related variables such as latent heating in the tropical latitudes. These concepts are developed in the following sections.

¹ According to the Oklahoma Monthly Climate Summary, August 1995 issued by the Oklahoma Climatological Survey.

6.3 Is the rain-yield concept useful in the tropics?

A prerequisite for using rain-yields to estimate rainfall (or any other parameter of interest), is that spatial and temporal tendencies in the lightning must be correlated with similar tendencies associated with the variable of interest (in this case rainfall). It is also worth recalling that the rain-yield, as developed in this study, has been computed *only for large areas and long time-scales*. Therefore, use of the rain-yield for rainfall estimation should only be attempted over *similar scales*. For example, based on observations presented in Chapter 4 of this study, and previous studies such as Rutledge et al. (1992), Williams et al. (1992), Zipser (1994), and Restivo (1996), it is clear that one cannot accurately compute rainfall using the rain-yield and CG lightning flash density for *individual* case studies of tropical convection. However, this is also true of satellite-based precipitation algorithms which are generally more successful at estimating precipitation when applied over large areas and monthly time-scales (e.g., Kidder and Vonder Haar, 1995; Ebert, 1997). The possibility of using a “rain-yield” type of technique for calculating convective rainfall over large spatial and temporal scales in the tropics cannot be investigated thoroughly due to a lack of data (e.g., lightning data). In this section we present a limited amount of observational evidence to further support the idea that rainfall and lightning are fundamentally related. Based on these relationships, we suggest that rain-yields may be useful under certain circumstances for quantifying monthly or seasonal rainfall in the tropics.

6.3.1 Annual cycles of lightning and rainfall over French Guyana, 1989

Figure 6.4 shows the 1989 annual rainfall and monthly total lightning flash counts detected by the SAFIR lightning detection system (Richard, 1990) located in Kourou, French Guyana

(5°S, 52°W). Examining the annual cycle, Fig. 6.4 suggests that there is little relationship between rainfall and lightning for this particular location in the tropics. However, as in many other locations in the tropics, the rainfall regimes in French Guyana vary as a function of the location of the Inter-Tropical Convergence Zone (ITCZ) (e.g., Molinie and Pontikis, 1995). When the ITCZ moves to the south of French Guyana during the months of November to May, the region experiences heavy rainfall accompanied by less lightning, characteristic of a more monsoon/oceanic-like regime (under a moist northerly flow into the ITCZ). During the northern hemisphere Summer and into early Fall, the ITCZ is situated just to the north of the area and the low level meridional winds take on a more southerly component, placing French Guyana in a moist, but more continental-like rainfall regime accompanied by slight increases in the surface wet-bulb temperature (Williams, 1996) and significantly more lightning (e.g., Molinie and Pontikis, 1995). This characteristic shift in lightning/rainfall regime as a function of ITCZ location is very similar to that experienced over tropical northern Australia (e.g., break vs. monsoon periods) and other near equatorial regions of the tropics (e.g., Rutledge et al., 1992; Zipser, 1994).

If we now sub-divide the annual cycle in Fig. 6.4 based on the regional transitions in rainfall regime discussed above, and use the months of June and November 1989 as the “transition” months, we arrive at the plots shown in Figs. 6.5 a-b which reflect the presence of two rainfall regimes (Regimes I and II; Fig. 6.4) in the annual cycle. Note that the co-variation between lightning and rainfall in both regimes becomes more apparent. This regime dependence is further illustrated when the partitioned monthly data points are presented in the form of a scatter plot (Fig. 6.6). The difference in the relationship between the total lightning flash density and the rainfall are striking when considered as a *function of regime*. Indeed, the slopes of the “best-fit” lines through the data points are equivalent to a “rain-yield” value calculated using

total lightning flash density (exact rain-yields cannot be computed because they are a strong function of the coverage area which is not precisely known at this point). It is interesting to note that the slopes of the best-fit lines for Regimes I and II differ by a factor of 10. The Regime I slope (rain-yield) for the months of Dec./Jan.-May is the largest and characterized by low flash densities and high rainfall. The Regime II slope (rain-yield) on the other hand is characterized by much larger flash densities and slightly less rainfall (July-October). The order of magnitude difference in slope (rain-yield) between the two regimes is very similar in character to the order of magnitude change in rain-yield that occurs between the break and monsoon periods over the maritime continent region of northern Australia (Fig. 6.1).

Figures 6.4-6.6 (and earlier discussion regarding the Maritime Continent region of northern Australia) clearly illustrate the need to consider changes in rainfall regime. If the regime changes are considered, it appears that using rain-yields to estimate rainfall over tropical continents has promise (because of the paucity of lightning and tropical rainfall data, our data sample is small). There are too few data points in Fig. 6.6 to say that the correlation between total lightning and rainfall over French Guyana is strong enough to use the rain-yield as an estimator of rainfall for this area. Yet, the systematic changes in rain-yield (identified by regime) coupled with the temporally correlated trends shown in Figures 6.5 and 6.6, along with similar findings demonstrated in other locations are encouraging and suggest further research is warranted.

6.3.2 Diurnal cycles of rainfall and lightning in the Maritime Continent region

If it can be demonstrated that rainfall and lightning flash frequency (in this case, CG lightning) exhibit temporal coherence and similar amplitude variations as characteristics of their

respective diurnal cycles, then this significantly strengthens the hypothesis that in the mean, rainfall and lightning can be quantitatively related. The phrase “in the mean” refers to the time scale over which data points for the diurnal cycle were collected. The monthly diurnal cycles of rainfall and CG lightning flash density are shown in Figs. 6.7 a-d for the tropical continent of northern Australia and the tropical islands of Melville and Bathurst for both break and monsoon period rainfall regimes. Figure 6.8 shows a similar diurnal cycle adapted from Fig. 3.13 for the tropical western Pacific Ocean. With the exception of the monsoon diurnal cycles (which only represent a single *10 day period*; e.g., days 48-57 in Fig. 3.1), the diurnal cycles were computed over time periods of 1-2 months.

The break period diurnal cycles of lightning and rainfall (Fig. 6.7 a-b) exhibit similar structure from phase, amplitude variation, and correlation standpoints. Linear correlation coefficients computed between lightning and rainfall for the island-break and continental-break period diurnal cycles were $R=0.84$ and $R=0.73$ respectively. Both lightning and rainfall peaked at approximately the same time of day in these two locales, but there is considerably more structure in the continental cycle (similar to a recent study by Li et al., 1996). The ratio of the amplitude variations of the lightning and rainfall cycles (here defined as the cycles amplitude ratio; CAR) for both sets of diurnal signals are quite similar (~ 1). This suggests that the two variables, CG lightning flash density and rainfall, vary together in a consistent fashion in both the island-break and continental-break regimes.

When the continental monsoon and island monsoon diurnal cycles of rainfall and lightning are examined (Figs. 6.7 c-d), the effect of heavy rain falling from only weakly electrified convection becomes obvious. In Fig. 6.7c, it is clear that the islands received a great deal of rainfall from convective systems that did not produce much CG lightning, evident as large peaks in rainfall at 1000L, 1400L, and 2200L that were accompanied by only very weak peaks in the

CG lightning counts. Conversely, at 1600L there is a very pronounced peak in CG lightning activity (eg., Fig. 4.2) consistent with the afternoon forcing of convection over the island (even during the monsoon; e.g., Simpson et al., 1993) and a primary peak in the diurnal cycle of rainfall. Over the continent during the monsoon (Fig. 6.7d), the relationship between the lightning and rainfall (for this 10 day sample) appears to be significantly stronger ($R=0.73$) than that of the islands ($R=0.46$), but approximately 50% of the integrated rainfall is associated with little or no lightning activity (e.g., 0000L-1200L).

In contrast to the CAR's computed for the break period at both locations, the CAR's computed for the island and continental locations during monsoonal periods were 0.22 and 0.67, respectively. The decrease in the island CAR during the monsoon period (relative to the break period CAR which was ~ 1) suggests that the relationship between lightning and rainfall over the islands during the monsoon regime is weak. Additionally, while the absolute peak and trough in the island-monsoon diurnal cycles of lightning and rainfall occur in phase, the overall temporal coherence of the cycles is rather poor (which is reflected in a lower correlation coefficient of $R=0.46$).

The monsoon and break period diurnal cycles for the maritime continent can be further compared to similar diurnal cycles observed over the western Pacific warm-pool (Fig. 6.8). Similar to the continental diurnal cycles, the diurnal cycle of total rainfall and CG lightning flash density over the warm pool exhibit some temporal coherence, which is encouraging. However, there are *marked* differences in amplitude variation between the oceanic rainfall and CG lightning diurnal cycles. This results in a CAR of only 0.22 for the oceanic diurnal cycle, similar to that observed for the monsoon regime over the tropical islands of Melville and Bathurst.

The results shown in Figs. 6.5-6.8 suggest that lightning, at least from a CG flash perspective, *may* (with further data acquisition and analysis) be useful for quantifying monthly or

seasonal rainfall amounts over tropical continents and islands provided the rainfall regime is known (e.g., monsoon or break; ITCZ north or south, etc.). The results shown in Fig. 6.8 suggest that using lightning to quantify rainfall over the oceans (or in more oceanic environments such as the monsoon over the islands of Melville and Bathurst) is more problematic. This is due to the large proportion of the total rain that falls from only weakly electrified oceanic convection (e.g., Williams et al., 1992; Zipser, 1994).

Even though the possibility of using lightning data to estimate rainfall over the oceans is problematic, it may not be impossible. The reasoning follows the simplified flow diagram shown in Fig. 6.9 and to some extent, Fig. 1.1. Large scale moisture convergence over the tropical ocean is explicitly related to surface rainfall through the "P-E" term (precipitation-evaporation) in the water budget equation (neglecting storage terms; Piexoto and Oort, 1992; Wu and Liu, 1992). Observations presented in Chapter 3 *strongly suggest* that moisture convergence increases CAPE and θ_w (which are positively correlated; Williams and Renno, 1993; Fig. 3.4) by increasing the boundary layer mean mixing ratio (q ; Figs. 3.2-3.3). As the CAPE increases, the amount and/or intensity of the convection increases, the amount of precipitation formed in the ice-phase increases relative to that produced by collision-coalescence processes (e.g., 21% vs. 34% as suggested in the NACT and ACT model simulations, Secs. 5.2.1-2; see also Figs., 4.5-4.8), and the CG lightning flash rate increases (Figs. 3.1, 3.5-3.6, 3.8-3.10; also Rutledge et al., 1992; Williams et al., 1992; Zipser and Lutz, 1994). Of course, this argument also applies to continental convection as increased moisture convergence also serves to increase the moist entropy of a tropical continental convective boundary layer, and hence the CAPE..

Over regions of the tropical ocean where both deep convection and lightning occur, the above arguments suggest that variations in lightning activity should to some extent, reflect variations in the surface rainfall (Fig. 6.9), thus providing a physical framework for the rain-

yield concept. If the lightning and rainfall vary together in any reasonable fashion (e.g., Fig. 6.8), it may be possible to relate lightning and rainfall quantitatively under certain conditions over the tropical ocean (e.g., perhaps seasonally, using total flash density).

6.4 Summary

We have presented a new approach to quantifying the relationship between lightning (cloud-to-ground lightning in particular) and surface rainfall using a parameter called the “rain-yield”. The “rain-yield” is defined as the ratio of rain-flux to CG lightning flash rate density, and is computed over areas of $O(10^5 \text{ km}^2)$ and time-scales of approximately one month. The advantage of using large spatial and temporal scales to compute the rain-yield is that reasonably high positive correlations between rainfall and lightning can be retained (as in many previous studies), along with a relatively stable value of the rain-yield for given rainfall regimes. Furthermore, cloud and system-scale variability between rainfall and CG lightning are filtered out. This approach is similar to remotely estimating surface rainfall from satellite or radar using algorithms such as GPI and ATI respectively.

The rain-yield exhibited remarkable stability over the mid-latitude continental U.S. with values centered around $\sim 10^8 \text{ kg/fl}$. Further, a physically consistent transition from lower to higher yields occurs as the large-scale environment in which deep convection forms, transitions from a mid-latitude continental regime to a tropical oceanic regime. Exploiting the reasonably high correlations observed between lightning and rainfall (Table 6.1) for mid-latitude continental locations south of approximately 40°N , and the relative stability of the rain-yields (Fig. 6.1) in the 1994 data set, it was subsequently demonstrated that the product of the appropriate rain-yield and monthly CG flash density could produce a reasonably accurate rainfall estimate (within a

factor of two of the gauge measurements) for many different locations south of 40°N (Fig. 6.3; Table 6.2).

Questions arise when trying to follow a similar procedure in the tropics. This is because the precipitation process in the tropics, especially over the oceans, *does not* require a robust ice-phase to initiate the production of heavy surface rainfall (since collision-coalescence processes are efficient). This is in sharp contrast to the electrification process which apparently *does* require a vigorous ice phase to produce electrical breakdown in a convective cloud. Hence, the nature of the relationship between rainfall and lightning in the tropics is even less certain than that hypothesized for the mid-latitudes. To examine this problem, we presented analyses using available data to describe both the annual cycle (1989) of total lightning and rainfall (satellite estimate) for French Guyana, 1989, and the diurnal cycles of rainfall and CG lightning for island and continental regions of tropical northern Australia, and the tropical western Pacific Ocean.

The analysis of the annual and diurnal cycles of both rainfall and lightning suggested that the two parameters are indeed related in the tropics, but the relationship displays a *strong regime dependence* (similar to the results of Zipser, 1994). For example, the positive correlation between lightning and rainfall was more pronounced for continental rainfall regimes than for continental monsoon or oceanic rainfall regimes. Over French Guyana, the regime dependence was a function of the location of the ITCZ. Marked changes in the slope (which was equivalent to the rain-yield) of the best fit line between rainfall and total lightning occurred as the ITCZ moved from north to south of French Guyana. The transition in ITCZ location is a climatological characteristic of the rainfall regime over French Guyana (e.g., Piexoto and Oort, 1992), and can therefore be accounted for when considering the quantitative relationship between lightning and rainfall for that area. Hence, a rain-yield approach to estimating monthly rainfall

similar to that attempted for the continental U.S. in Table 6.2, might be feasible for continental locations of the tropics, provided the appropriate rain-yield for a given rainfall regime is selected.

Over the tropical oceans the quantitative relationship between lightning and rainfall appears less "robust". This was reflected in the CAR value of 0.22 computed for the oceanic diurnal cycle (and to some extent in Fig. 3.16). While the relative peaks in the diurnal cycle of lightning and rainfall (Fig. 6.8) appear to be approximately in phase over the ocean, the diurnal cycle of the rainfall is of much lower amplitude than that of the lightning. This suggests that a considerable portion of the total rainfall over the ocean originates from clouds that do not produce lightning (e.g., Figs. 4.11, 4.13; also Zipser, 1994). However, based on observed, theoretical, and modeled relationships between tropical large-scale moisture convergence, rainfall, CAPE (or boundary layer wet-bulb temperatures), lightning frequency, and convective intensity discussed in Chapters 3-5, we argued that it might still be possible, perhaps over seasonal time-scales, to quantitatively relate lightning activity to rainfall over the tropical oceans.

Warm-Season Normalized Rain-Yields

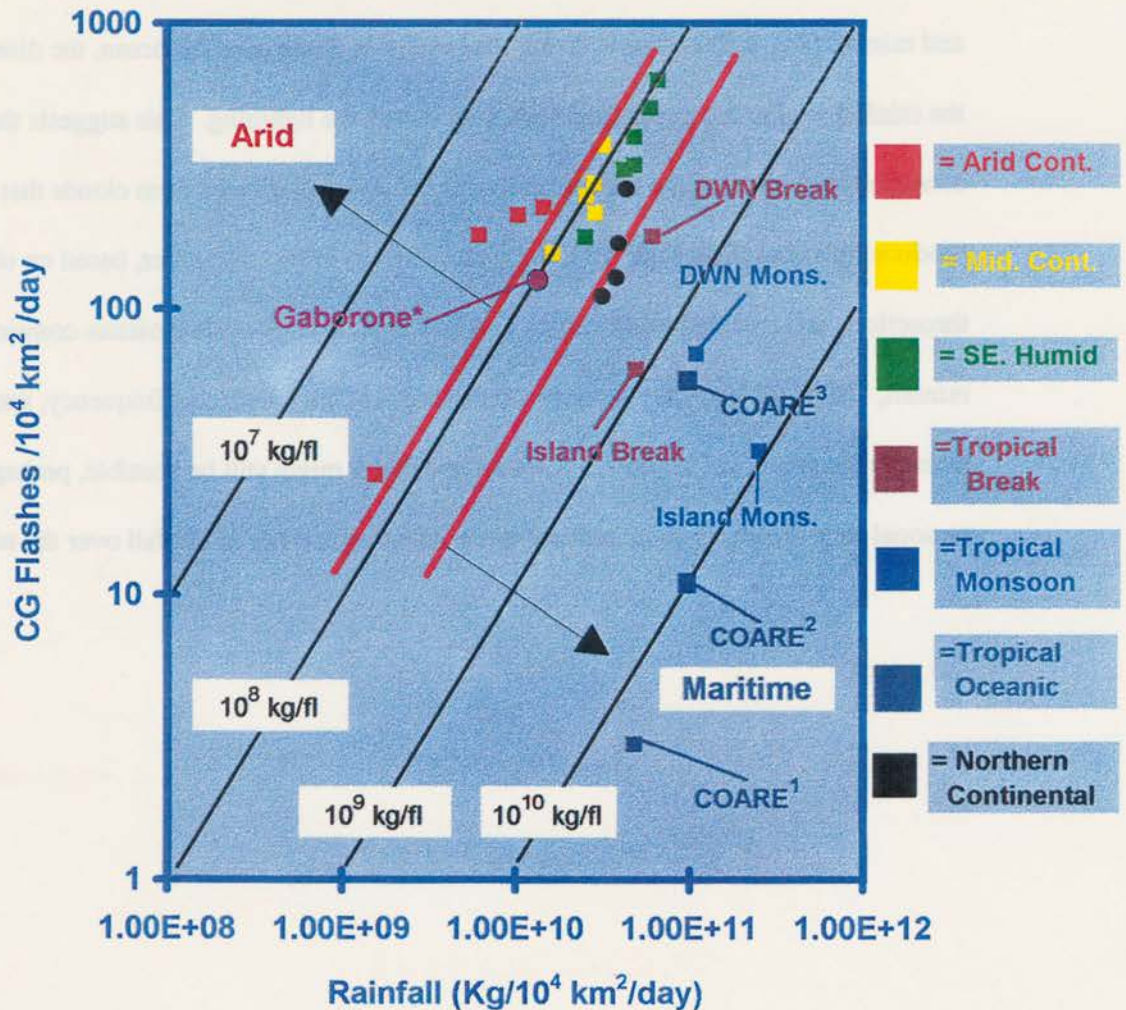


Figure 6.1 Warm-season normalized CG flash density (flashes/10⁴ km²/day; on the ordinate) vs. rainfall (kg/10⁴ km²/day; on the abscissa). Sloping solid black lines are contours of constant rain-yield (kg/fl). Each data point is colored and labeled (as indicated in the figure) according to each specific rainfall-regime discussed in the text. The large bold arrows indicate directions for which more "arid" or "tropical oceanic" rainfall-regimes are encountered. The two bold, red lines are subjectively placed to help delineate the arid and tropical/oceanic regimes from the combined mid-continent and southeast humid regimes.

Precipitation (m^3)

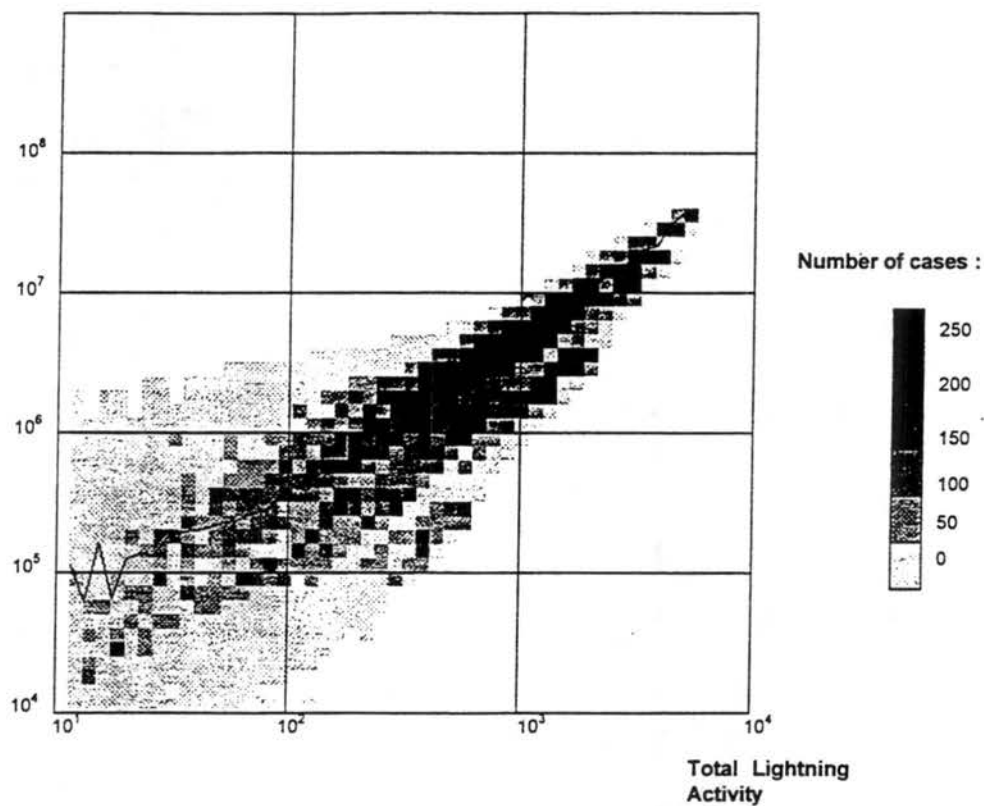


Figure 6.2 Scatter plot of rain-volume (ordinate) vs. total flash count (abscissa) for individual convective systems sampled in France (adapted from Richard and Lojou, 1996).

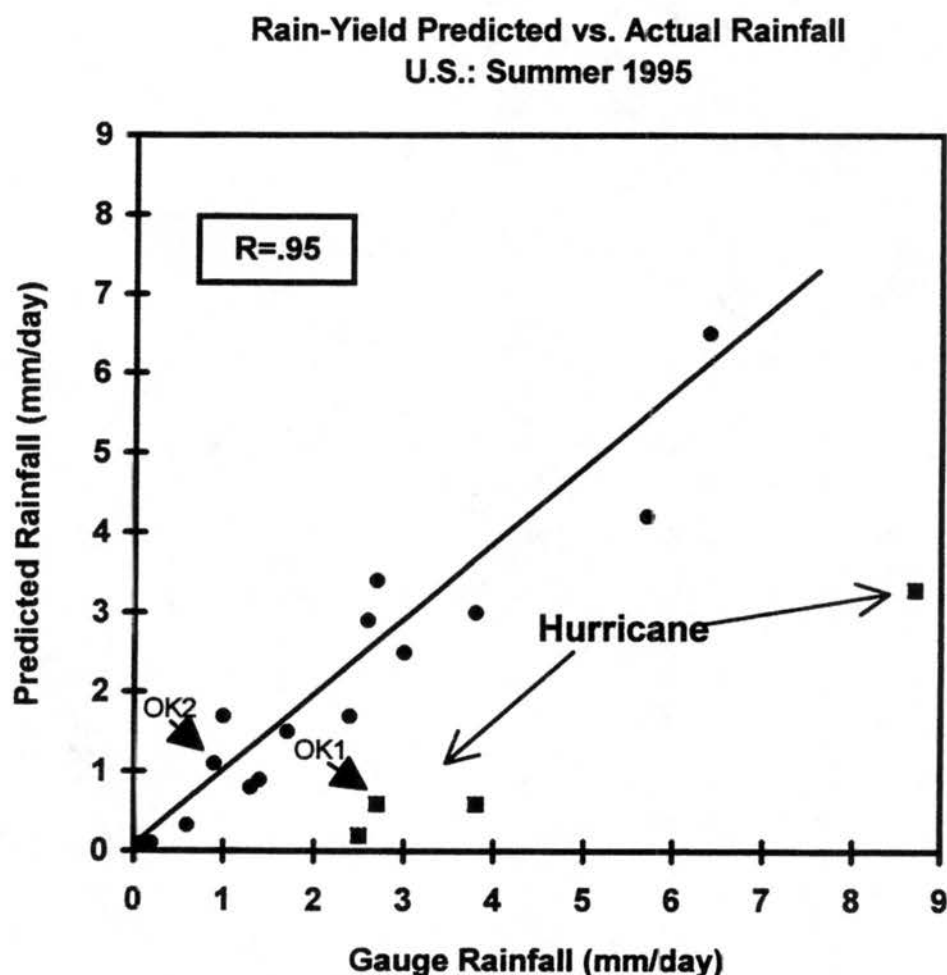


Figure 6.3 Scatter plot of rain-yield “predicted” rainfall (mm day^{-1}) vs. gauge measured rainfall (mm day^{-1}) for the summer-season of 1995 in the continental U.S. Hurricane/tropical storm data points (plotted as squares) are also indicated. Data points OK1 and OK2 represent rainfall values computed for western and eastern Oklahoma respectively, for the month of August, 1995. The “best-fit” line and linear correlation coefficient ($R=0.95$) for non-hurricane affected data points (solid circles) are also indicated in the figure.

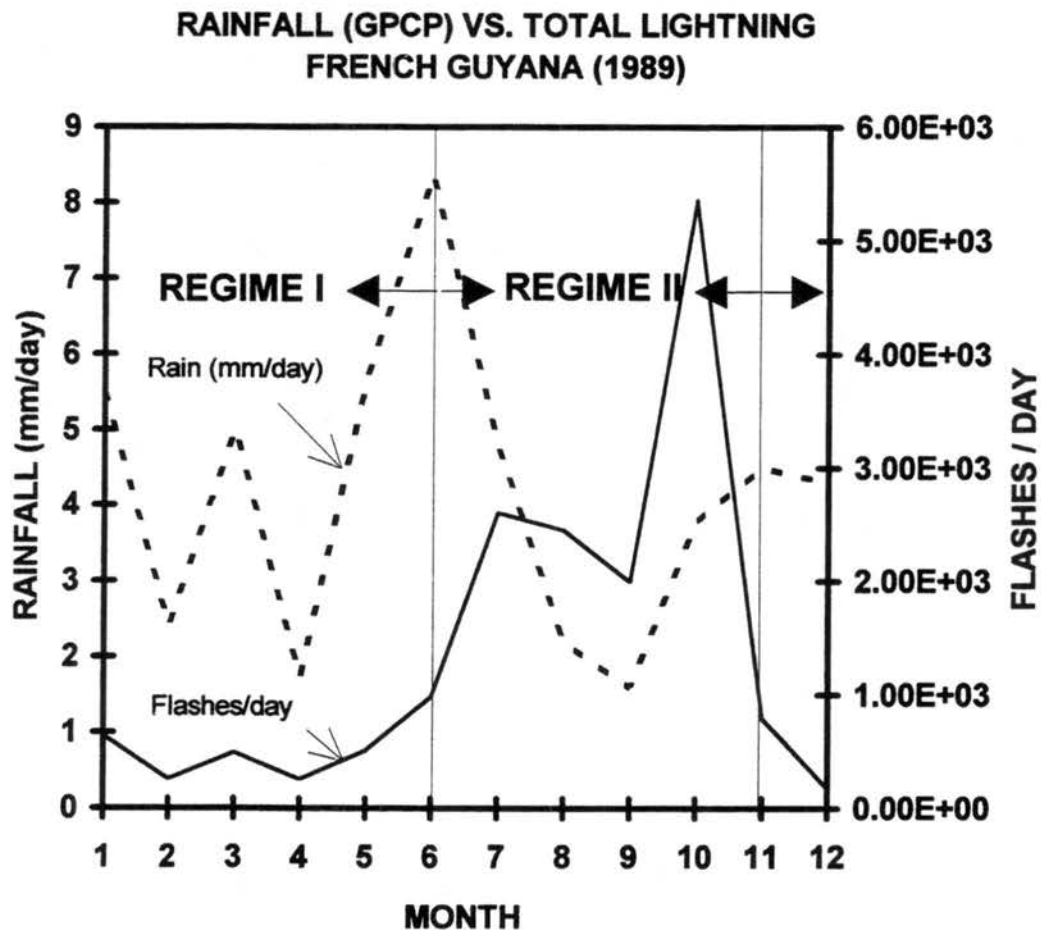


Figure 6.4 A time series of monthly rainfall (GPCP estimate; mm day^{-1} , left ordinate) and total lightning flash counts (flashes day^{-1} ; right ordinate) from 1989 for the vicinity of Kourou, French Guyana. The annual cycle is further subdivided into two rainfall regimes (Regime I and II) as described in the text. Bold arrows indicate the months associated with each specific regime.

a) **French Guyana (Jan-June; Nov-Dec, 1989)**
Norm. Rainfall (GPCP) vs. Total Lightning

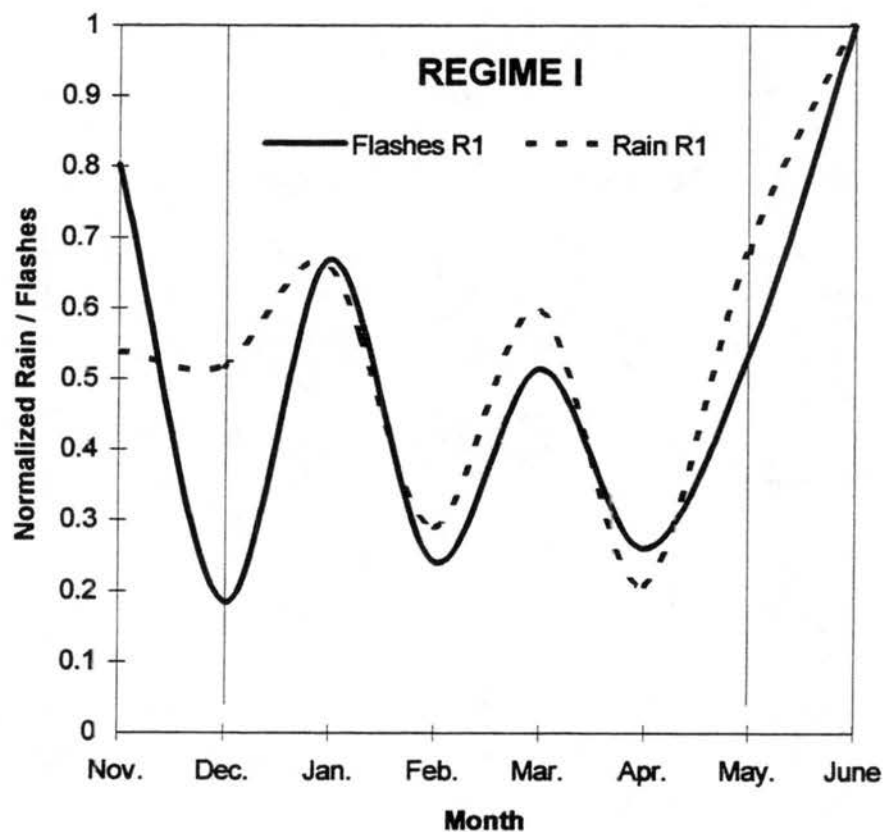


Figure 6.5 a) Time series of normalized rainfall (dashed line) and total lightning flash count (bold solid line) for Regime 1 of Fig. 6.4. Light solid lines indicate the bounds of Regime I (December-May) with transition periods during June and November. The months of November and December, 1989, are placed in sequence prior to that of January, 1989 for the purpose of presentation.

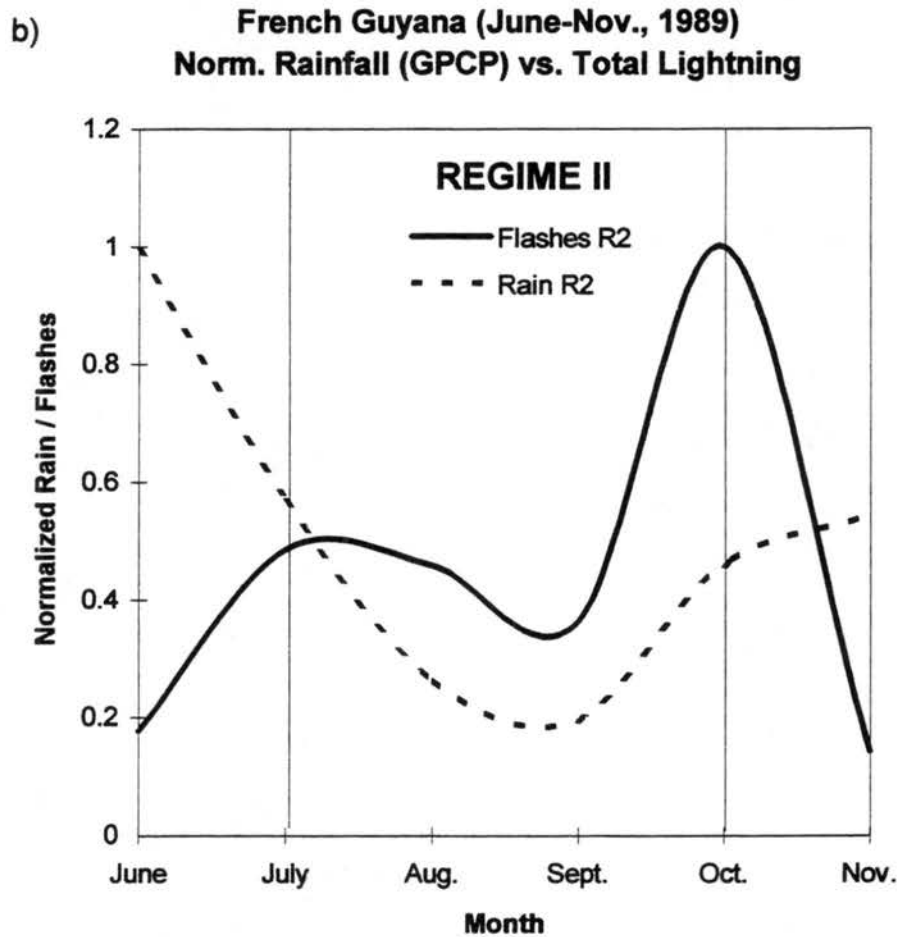


Figure 6.5 b) as in (a) but for Regime II (July-October).

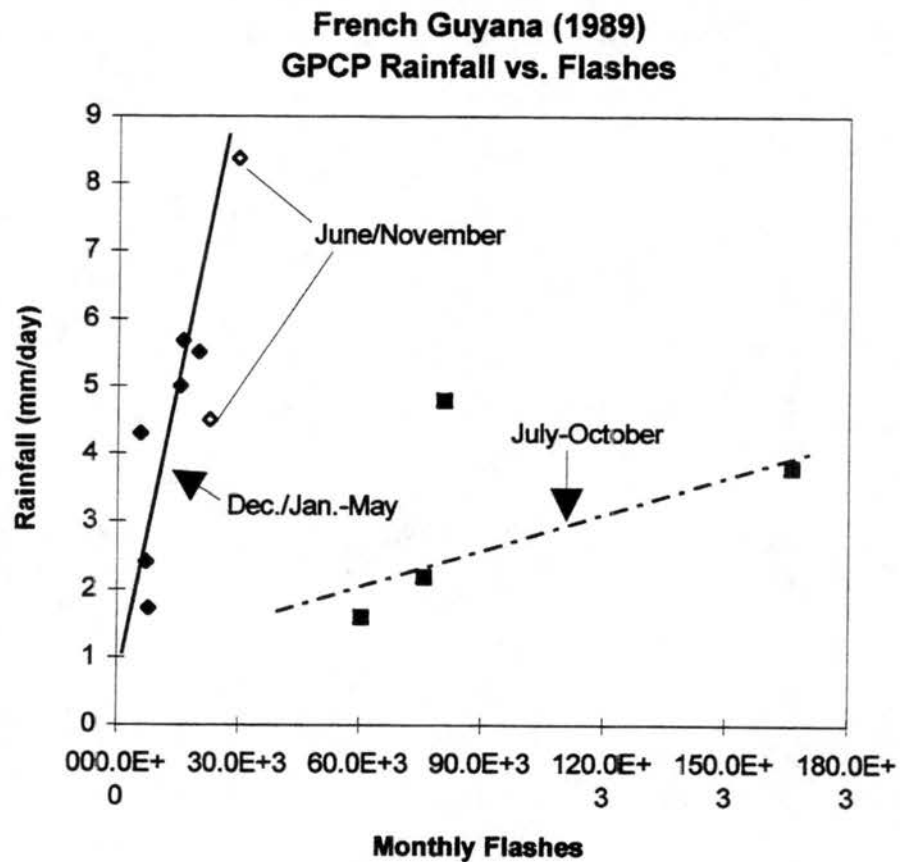


Figure 6.6 A scatter plot of rainfall (mm/day; ordinate) vs. monthly total flash count (abscissa) for Kourou, French Guyana, 1989. Months from Regime I (Fig. 6.4) are indicated with solid diamonds, Regime II (Fig. 6.4) with solid squares, and the transition months of June and November, 1989 are indicated by open diamonds. The "best-fit" lines for each regime are also shown (solid line- Regime I; dashed line Regime II).

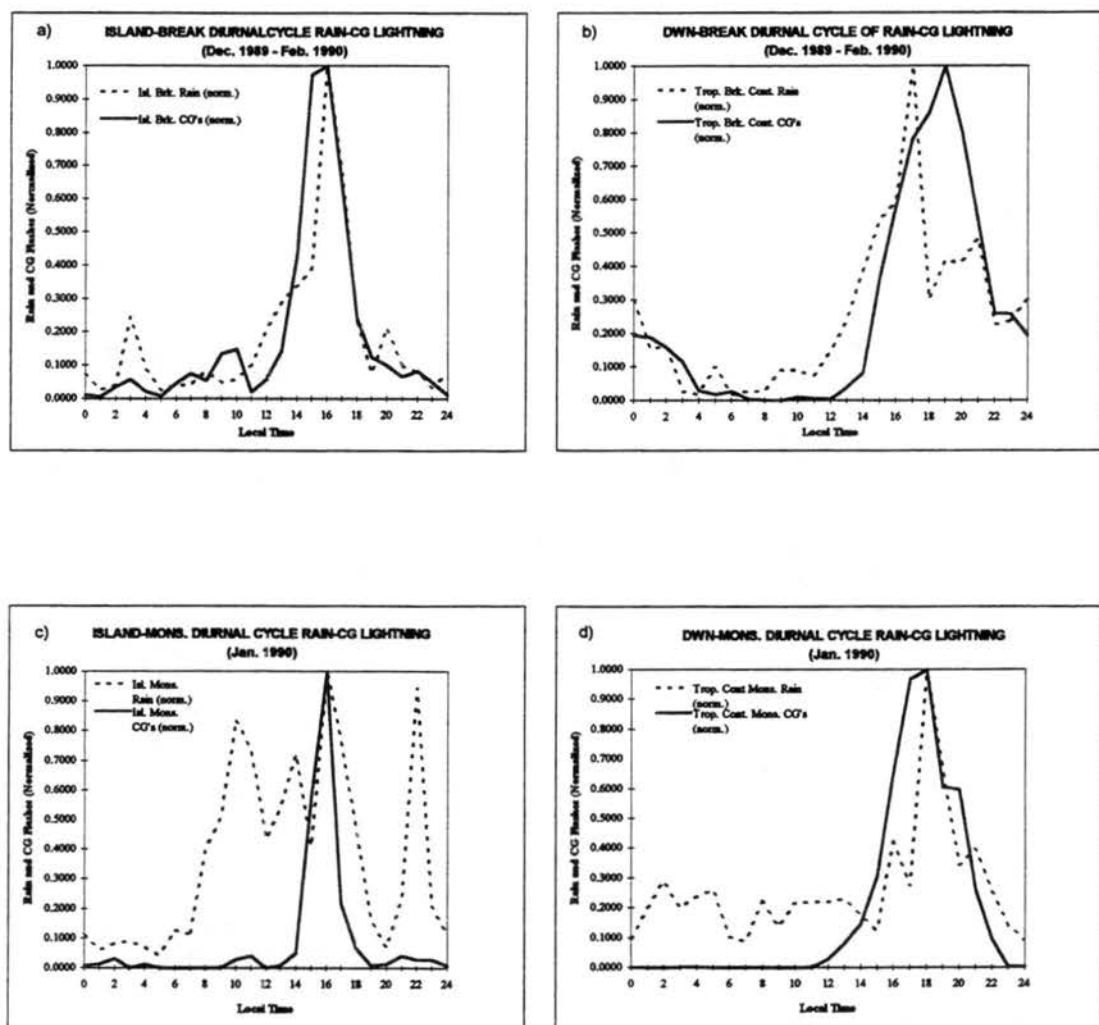


Figure 6.7 Diurnal cycles of rainfall (dashed line) and CG lightning (solid lines) for the tropical island and continental regions of northern Australia (Dec. 1989 - Feb. 1990). Normalized values of rainfall and CG lightning are plotted on the ordinate, local time on the abscissa. The diurnal cycles are partitioned into break and monsoon period rainfall regimes for the island (a) and (c), and continent (b) and (d) respectively.

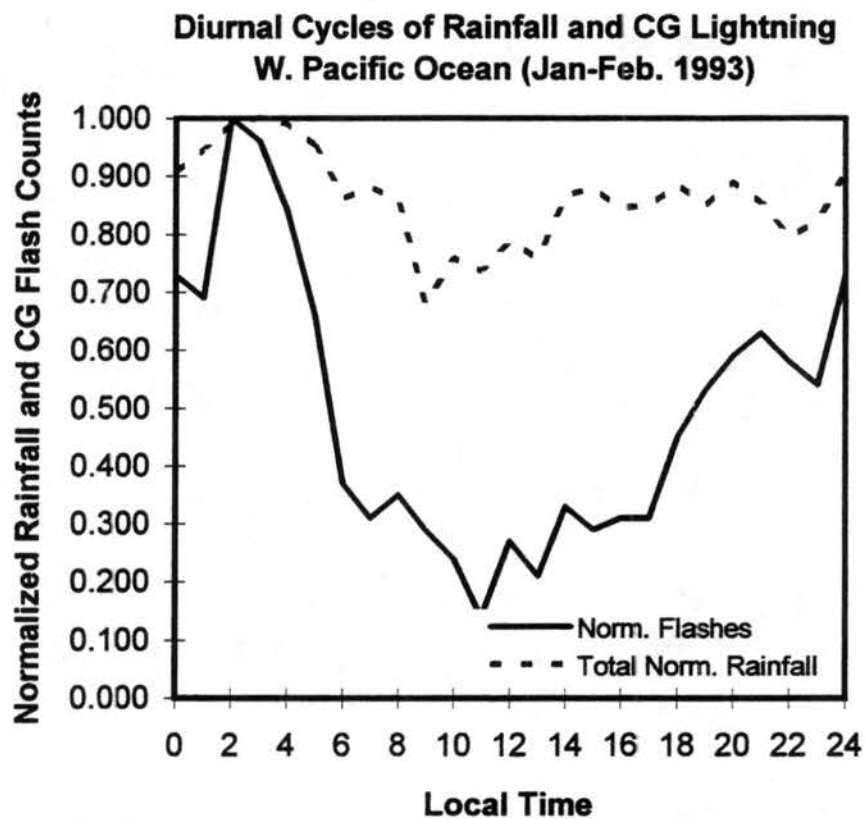


Figure 6.8 As in Fig. 6.7 but for the western Pacific warm-pool (Jan.-Feb. 1993).

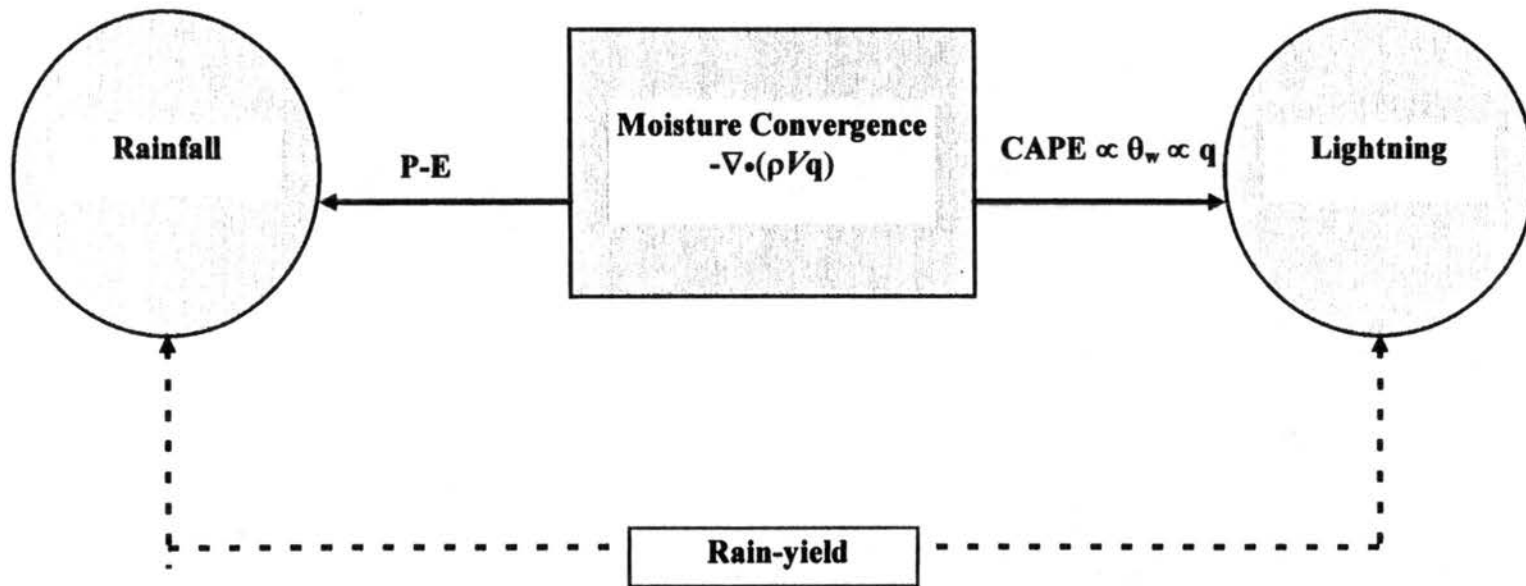


Figure 6.9 Conceptual model illustrating the relationship between lightning and rainfall over the tropical western Pacific Ocean as a function of the large-scale variables moisture convergence and CAPE (see text for discussion). In the diagram, “P-E” represents net surface rainfall, and “ θ_w ” and “ q ” represent boundary layer wet-bulb potential temperature, and mixing ratio respectively.

CHAPTER 7

SUMMARY AND CONCLUDING REMARKS

7.1 Summary

The primary question addressed in this study was: What do cloud-electricity (lightning) observations tell us about processes occurring over cloud and larger-scales in the tropical atmosphere (Fig. 1.1)? To address this question, new observations associated with lightning and non-lightning producing tropical oceanic convection were compared to similar observations from other rainfall-regimes (e.g. tropical continental and mid-latitude continental). Significant differences in cloud structure, evolution and electrical intensity were found to exist between, non-lightning and lightning-producing tropical oceanic convection (NACT and ACT respectively), lightning producing convection observed in other tropical regions, and (e.g., IT) mid-latitude locations (MC). An examination of composited sounding data associated with the environment of each class of convection (NACT, ACT, IT, MC) suggested that systematic differences existed in the thermodynamic structure (CAPE, shape of the thermal buoyancy profile) of the atmosphere associated with each regime.

Specifically, analysis of sounding data collected over both tropical northern Australia (DUNDEE) and over the tropical W. Pacific Ocean (TOGA COARE) suggested that variations in cloud-to-ground flash density *within* each regime were highly correlated to variations in the

magnitude of the CAPE (explicitly related to updraft strength). Over the tropical ocean, CAPE and the low-level (sfc-900 mb) moisture flux convergence varied coherently, with CAPE being highly correlated to the boundary-layer mean mixing ratio. The boundary layer θ_w was highly correlated to both the CAPE and PBL moisture, and the warm-pool CG lightning activity varied in a non-linear fashion with the θ_w (e.g., a 0.5°C increase in θ_w produced a factor of 5 increase in the CG flash density). Hence boundary layer moisture values exerted a strong control on the frequency of CG lightning detected over the warm-pool through increases in the PBL θ_w and the CAPE. Interestingly, temperature variations in the boundary layer, or at the sea surface, explained little of the variation in lightning activity over the ocean in contrast to what has been observed over continental and island locations (e.g., Lopez and Holle, 1986; Williams and Heckman, 1992; Williams, 1994).

The lack of a strong temperature dependence on CG lightning activity over the ocean became even more apparent when the diurnal cycles of lightning and precipitation were examined. The diurnal cycle of CG-lightning over the warm-pool exhibited a strong *nocturnal* peak (2 a.m. LST) when surface wet/dry-bulb temperatures were at a relative minimum, and a relative trough in activity during the middle of the day when the surface wet/dry-bulb temperatures were the largest. The CAPE, total and convective precipitation also exhibited similar nocturnal peaks (precipitation, 2 a.m.; CAPE between 11 p.m. and 5 a.m.) and mid-day troughs in activity, but the amplitude ratios of the precipitation cycle and the CAPE were much smaller than that of the CG activity (e.g., Rickenbach, 1996). The nocturnal-early morning peak in CAPE was the combined result of radiative cooling of the sounding aloft and low-level moistening of the sounding at pressures greater than 600 mb. The diurnal cycle of cold-cloud area for IR brightness temperatures less than 205 K reached a maximum in coverage three hours after the maximum in CG-lightning and convective rainfall (but during the broad peak in CAPE),

suggesting the presence of an extensive cirrus anvil. Hence, CG lightning data more accurately identified diurnal peaks in the convective and total rainfall (and hence convective heating) than did the IR brightness temperatures.

When daily counts of CG lightning flashes over the warm pool were plotted in time-longitude coordinates (similar to a Hovmöller diagram of OLR), the CG lightning data indicated the presence of propagating disturbances and the occurrence of a dry intrusion. We cannot say with certainty what fraction of the deep convection was producing lightning in the disturbances (statistics suggest that less than 10% of the 30 dBZ convective core heights reached the -10°C to -12°C level; DeMott, 1996), but we can say that the signature/location of “intense” convection forced by the disturbances was revealed in the lightning data. Periodic variations in the amount of lightning activity associated with the passage of disturbances over the warm-pool are consistent with the noted sensitivity between lightning flash density and the CAPE, θ_w , shear, MLMR and moisture flux convergence, all of which vary with the passage of a disturbance.

Though the amount of lightning observed over larger-scales was found to be well correlated to the CAPE *within* the convective regimes being examined, the word “*within*” should be emphasized because regimes such as the ACT, IT and MC all exhibited a similar CAPE but dissimilar amounts of CG lightning activity. For example, the ACT (IT) regime produced a factor of 10 (2) less lightning than the MC regime, but the CAPE’s only differed by $\sim 100 \text{ J kg}^{-1}$. The different CG flash densities observed in regimes with similar amounts of CAPE illustrates the importance of considering the *vertical distribution* of the thermal buoyancy (e.g., Lucas et al., 1994a). Distributions of thermal buoyancy (B_t) for the NACT, ACT, IT and MC regimes revealed small, but progressively larger increases in the peak B_t and in the slope of B_t at low to mid-levels of the atmosphere as the regime-mean CG flash density transitioned from lower (e.g., tropical oceanic NACT) to higher values (e.g., MC). The progressive increase in the slope of

the B_1 profile at low to mid-levels of the atmosphere should favor the development of stronger updrafts at low to mid-levels of the cloud. In turn, stronger updrafts are better able to counteract water-loading and entrainment, and in the case of tropical convection, more able to supply the necessary mixed-phase mass (e.g., exceed the $5\text{--}6\text{ ms}^{-1}$ “threshold” discussed above) for precipitation based charging of the cloud. Therefore, convection developing in the ACT or IT tropical environments were more likely to produce lightning (relative to the NACT regime).

Observations of cloud *internal* structure revealed striking differences between lightning producing (LP) and non-lightning producing (NLP) tropical oceanic convection. The tropical oceanic LP and NLP convection often had similar cloud-top heights, but radar data suggested that the LP clouds were associated with the presence of significant hydrometeor mass (as indicated by the presence of reflectivities $\geq 30\text{ dBZ}$) in the mixed phase region of the clouds. The mixed phase ice-mass present in LP oceanic convection appeared to be the result of an *upward* building reflectivity core that penetrated well into the cold regions of the cloud. Multiple Doppler-radar observations suggested that the mean updraft magnitude between the 0°C and -10°C region was *at least* $5\text{--}6\text{ ms}^{-1}$, large enough to loft millimeter diameter liquid, freezing and frozen drops together with sufficient cloud-water into the cold region of the cloud, thus providing conditions conducive to precipitation-based electrification of the cloud. Conversely, NLP tropical oceanic convection possessed mean updrafts less than $5\text{--}6\text{ m s}^{-1}$ at similar temperature levels. Therefore NLP clouds were not able to generate a robust mixed-phase ice-process in the cold region of the cloud, 30 dBZ reflectivities did not extend above the -10°C level, and electrical charging of the cloud was suppressed relative to LP clouds.

The tendency of an updraft to exceed any required “threshold” strength for electrification (lightning) to occur is affected by the thermodynamic structure of the ambient environment. This reasoning set the stage for modeling sensitivity studies whose purpose was to test the degree to

which the thermodynamic structure of the large-scale environment affected the electrical, microphysical and kinematic characteristics of *individual* convective clouds. These tests were performed by initializing a numerical cloud-model (with an electricity parameterization) with the composited storm environments from each respective regime (NACT, ACT, IT, MC). Based on the cloud-simulations, potential feedbacks (or “outputs”; Fig. 1.1) from the cloud to the large-scale such as diabatic heating and the microphysics of the rainfall process were examined and partitioned relative to the class (based on electrical intensity) of convection being simulated. Hence the modeling sensitivity tests directly related large-scale inputs to cloud-scale processing and outputs, while the observations provided a method to validate the model results with regard to the gross structural characteristics of the simulated clouds.

The sensitivity tests were completed using a modified version of the 1-D bulk microphysical cloud-model of Ferrier and Houze (1989) which included an upgraded four-ice category microphysical scheme and a parameterized non-inductive charging scheme. In addition to testing the sensitivity of the convective/electrical development to the large-scale thermodynamic structure of the atmosphere, we also examined 1) the role played by frozen drops in the electrification of tropical convection; 2) changes in the relative contribution of ice and warm-rain processes that occurred as clouds became more electrified; and 3) relative differences in the diabatic heating (excluding radiative effects) between lightning producing and non-lightning producing clouds.

Model simulations suggested that tropical cumulonimbi forming in environments of progressively larger CAPE became more electrified, and did so earlier in the simulation. Further, as the low to mid-level B_z profile gradually transitioned from the NACT to MC environments, the simulated convection became more vertically developed and more strongly electrified (as inferred by the time taken to achieve strong in-cloud electric fields prior to electrical break-down).

Consistent with large-scale observations of CAPE, CG flash density and boundary layer moisture variables over the tropical ocean, the simulated oceanic convection exhibited a slightly stronger sensitivity to perturbations in PBL moisture content than comparable perturbations in PBL temperature. In addition to thermodynamic considerations, the modeled convective updrafts were also sensitive to the magnitude of the gust front forcing applied to the cloud-base, suggesting that a slight increase in the dynamical forcing of the convection can compensate reduced values of CAPE with regard to the intensity of the convection.

Frozen drops (high-density ice) played an explicit role in the electrification of the simulated tropical convection. In all simulations where the model-cloud produced lightning, frozen drops and ice crystals were the major charge carriers in the cloud. It was also noted that the height of the 36 dBZ reflectivity contour in the simulated clouds (recall that the 30 dBZ contour was discussed in the observations) was correlated to the relative electrical intensity of the clouds and to the peak in frozen drop mixing ratios. In the simulated clouds, mean updraft velocities across the top 1 km volume of 36 dBZ reflectivity cores were very similar to the vertical velocity maxima observed in multiple-Doppler retrievals at the top of 30 dBZ cores in several cases from TOGA COARE. These drafts were large enough to loft 1-2 mm diameter frozen drops in the mixed phase region of the cloud.

Participation of the ice-process in the formation of rainfall also increased as the simulated clouds of each respective regime became more vertically developed and more intense from an electrical perspective. In the NACT, ACT and IT simulations the first well defined peak in surface rainfall occurred in association with the development of large rain mixing ratios between the 2-3 km levels as a result of warm-rain coalescence (no such "warm-rain" peak was observed in the MC simulation). A secondary peak in rainfall accompanied all of the tropical simulations that was associated with the fallout of rain formed higher in the cloud and the descent of melting

precipitation-ice (frozen drops and graupel) from above the 5 km level. As the NACT, ACT and IT simulations became progressively more intense, the percentage of total column integrated rainfall contributed by the ice-process (manifested in the secondary peak of rainfall) increased incrementally from 21% in the NACT simulation, to 34% in the ACT simulation, and to 42% in the IT simulation. In fact, in the lightning-producing IT cloud, the secondary peak in rainfall was slightly larger than the initial warm-rain peak. For comparison, in the MC simulation, melted ice accounted for 75% of the total rainfall produced in the column and was associated with the primary (and only) peak in surface rainfall.

It is interesting to note that as the secondary peak in rainfall became more pronounced with increasing cloud intensity (electrically and kinematically), the vertical distribution of diabatic heating and drying in the model cloud changed in a systematic manner. The transition in heating/drying profile was characterized by a juxtaposition in the relative magnitudes of the low-level (2-3 km) and upper level (5-7 km) heating peaks, and as a relative increase (decrease) in the upper (lower) level drying rates. As the simulated cells of each regime became incrementally more intense, the magnitude of the upper level heating (6-8 km) became larger than that of the low-level heating and occurred over a greater depth of the troposphere (as manifested by a decrease in the rate at which the heating fell off with height above the 6-8 km layer). The relative increase in the upper level heating associated with the ACT, IT and MC simulations occurred in response to an increase in condensational heating (water and ice) above the freezing level associated with the enhanced mixed-phase processes that existed in those clouds, and as a result of enhanced eddy heat flux convergence near the top of each of the clouds. The transition in the relative location of maximum heating with intensity in the simulated clouds is consistent with the modeled transition in rainfall process and the increasing importance of the secondary rainfall peak in the more electrified simulations.

Transitions in the rainfall process on the cloud-scale (as simulated and observed) were manifested on larger-scales when the "rain-yield" was computed. The rain-yield is a parameter developed to compare observed values of the CG flash density to warm-season area mean rainfall over large areas (10^5 km^2) and temporal domains (one month). Rain-yields computed for the summers of 1994-95 in the mid-latitude continental U.S. (latitudes $<40^\circ \text{ N}$) exhibited remarkable stability with values centered around $\sim 10^8 \text{ kg/fl}$. Correlations between rainfall and CG flash density were high in the southern U.S., especially in the southwestern U.S., where much of the summer rainfall is associated with lightning producing convection and an ice-based precipitation process. Indeed, for the mid-latitude continental regime it was subsequently demonstrated, that the product of the appropriate rain-yield and monthly CG flash density could produce a reasonably accurate rainfall estimate (within a factor of two of the gauge measurements) for many different locations.

A physically consistent transition from lower to higher rain-yields occurred as the large-scale environment in which the deep convection formed, transitioned from a mid-latitude continental regime (ice-based precipitation process, significantly higher flash densities; 10^8 kg/flash) to a tropical oceanic regime (larger contribution from warm-rain processes, less lightning; 10^{10} kg/flash). In the tropics, the relationship between rainfall and lightning (e.g., the rain-yield) is not as well defined as it was in several areas of the mid-latitudes because the tropical rainfall process has a substantial warm-rain component that is essentially decoupled from the electrification process. The analysis of monthly rainfall and total lightning flash counts over French Guyana coupled with analyses of the diurnal cycles of rainfall and lightning over the western Pacific warm-pool and Maritime Continent region of northern Australia, do suggest that lightning and rainfall are related in the tropics, but the relationship displays a *strong regime dependence*. For example, the positive correlation between lightning and rainfall was more

pronounced for continental rainfall regimes than for continental monsoon or oceanic rainfall regimes. Over French Guyana, the regime dependence appeared to be a function of the location of the ITCZ. Hence, a rain-yield approach to estimating monthly rainfall similar to that attempted for the continental U.S. *might* be feasible for continental locations of the tropics when the appropriate rain-yield for a given rainfall regime is selected. Within tropical oceanic regimes where the relationship between rainfall and lightning was weaker, observed and modeled relationships between large-scale moisture convergence, rainfall, CAPE (wet-bulb temperatures), and lightning frequency suggest that it might still be possible, perhaps over monthly or seasonal time-scales, to quantitatively relate lightning activity to rainfall.

7.2 Conclusions and future research

This study has demonstrated that cloud-electrical characteristics in the tropics such as the presence, absence and/or amount of lightning, can be used as a tool for remotely inferring characteristics associated with both the parent convection and the large-scale environment. Over large-scales, the amount of lightning observed in the tropics (assuming large-scale trends in CG lightning reflect trends in total lightning) at any given time is highly correlated to the magnitude and/or supply of CAPE specific to a given convective regime, which is consistent from a quasi-equilibrium standpoint because as the supply of CAPE increases, the amount or the intensity of the convection must also increase to maintain an equilibrium condition. However, a *strict* “equilibrium” condition in the CAPE (e.g., $\frac{\partial \text{CAPE}}{\partial t} = 0$), as reflected in the CG lightning activity, was never really observed in the tropical oceanic and continental sounding data presented in this study. Variations in the CAPE over the warm-pool produced small changes in

the boundary layer θ_w . In turn, small changes in boundary layer θ_w were associated with a disproportionate response in the CG flash density. Though the time-scale is much shorter, this result is in general agreement with the hypothesis of Williams (1992, 1994) regarding the use of lightning flash density as a sensitive indicator for changes in global surface properties such as θ_w . Additionally, since the observed dependence of lightning flash density on θ_w is amplified in tropical atmospheres (Williams, 1994), observations of lightning over extended time periods (e.g., months or years) in the tropics are more likely to reveal small changes or adjustments in the tropical atmosphere than in the mid-latitudes.

In addition to using lightning data for global change applications, the rapid response of lightning flash density to small changes in thermodynamic parameters (e.g., perhaps being forced by a migrating disturbance), and the unique characteristics of lightning producing clouds in the tropics, suggest the possibility that lightning data could eventually be used in real-time as an aid in the representation of rainfall, microphysics or sub-grid scale diabatic heating in numerical weather prediction (NWP) models (e.g., Puri, 1987). This suggestion is based on observations and model simulations that suggest lightning-producing tropical clouds exhibit different kinematic, microphysical and convective heating structures relative to deep, non-lightning producing clouds. Of course, the one-dimensional model of a single convective cell (e.g., not the 3-D ensemble) used in this study does not *presently* provide a sufficient *quantitative* basis for using lightning observations to infer large-scale heating profiles for real-time assimilation into NWP models. However, *we have demonstrated qualitatively* that the use of lightning observations in this fashion is *plausible* (and this will be a primary topic of future research).

One other potential utility of lightning data suggested by this study is the diagnosis of area mean rainfall in warm-season convective regimes. Results from rain-yield calculations suggest that under certain circumstances CG lightning may be used to diagnose rainfall over monthly

time scales and areas $\geq 10^5 \text{ km}^2$. The use of cloud-to-ground or total lightning flash density for describing rainfall statistics is an attractive concept because mapping lightning is done in a relatively straight-forward manner, giving this technique advantages over others. Also, widespread coverage is possible from lightning networks, even over the tropical oceans, using available VLF techniques or spaceborne sensors such as OTD or LIS. Accurate rainfall estimates in the tropics are critical for climate-based studies and it is these regions where rain measurement by gauge or radar is particularly difficult. Furthermore, while satellite-based rain estimates can provide the required coverage over remote regions of the tropical oceans, they are often susceptible to error when large regions of non-raining cold cloud are present (e.g., Ebert et al., 1997). Clearly, future research concerning the utility/application of the rain-yield and the conditions under which a rain-yield can be used to diagnose rainfall is warranted.

Important questions that are related to the comments above, and that need to be addressed in future research include: What percentage of a given ensemble of clouds in the tropics actually produces lightning, and what percentage of the total heating and rainfall produced in a cloud ensemble is associated with this lightning producing convection? Over the tropical ocean, are the very small fraction of undilute "hot-towers" previously postulated by Riehl and Malkus (1958) related in some way to tropical lightning producing "hot-towers" (e.g., Williams et al., 1992). In future research, we will attempt to address these questions through further observational and modeling studies. For example, using a cumulus ensemble model (cloud-resolving) modified to include a parameterization for electrification, it may be possible to obtain an estimate of both the percentage of lightning producing clouds in a cloud-ensemble and their relative contribution to the domain averaged heating/rainfall. Future observations using the TRMM-LIS or ground based VLF lightning detection networks together with other data platforms (e.g., radar, satellite passive instrumentation) will also enable statistics to be generated that describe the percentage of

lightning producing clouds present over the tropical ocean and/or continent at any given time.

Indeed, the broader implementation of lightning detection and location networks and spaceborne sensors now seems timely, especially to provide coverage in the tropics.

CHAPTER 8

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APPENDIX A

LIGHTNING LOCATION METHODS AND SITE ERROR DETERMINATION

Two return stroke location methods, magnetic direction finding and time-of-arrival, were used by researchers at both Colorado State University (CSU) and Texas A&M University (A&M) to compute the locations of return stroke positions associated with cloud-to-ground lightning flashes (positions were calculated for each return stroke in a flash). Since the majority of research and data collected during COARE concentrated on atmospheric and oceanic processes that occurred over and near the IFA, initial efforts were concentrated on calculating stroke locations for areas in the vicinity of the IFA. For example, at CSU stroke locations were initially computed for the area located to the east of a line connecting Kapingamarangi, Kavieng, Rabaul and a vector oriented along the 125° radial from Rabaul (Fig. 2.2a).

A.1 Flash positions computed using triangulation

Cloud-to-ground return stroke locations were initially calculated independently at A&M and CSU using magnetic direction finding techniques (i.e., triangulation; DF). At CSU the triangulated positions were calculated using the intersection of azimuth bearings from two ALDF antennas (e.g., Krider et al., 1976) and the Law of Sines on a sphere to determine a return stroke latitude and longitude. Since there were three ALDF antennas, when all three detected the same

return stroke, three separate triangulation's were possible. This yielded a set of three different positions for a given stroke that were initially averaged to arrive at the final stroke location. The DF calculation was completed in a similar fashion at A&M with the exception that a flat earth was assumed in the equations (in contrast to spherical geometry used at CSU). When the calculated positions were compared between the two groups, reasonably good agreement was found between the two sets of flash locations (i.e., locations were generally within 1-5 km of each other).

When all of the positions were calculated for a given time interval they were overlaid on radar and satellite data to provide additional checks on the location accuracy. As an example, Fig. A1 shows a horizontal cross-section of radar reflectivity at a height of 2 km collected on 1 February 1993 (0542 UTC) by the MIT C-band Doppler radar aboard the R/V Vickers (located approximately 400 km from the center of the ALDF network). The locations of cloud-to-ground lightning return strokes that occurred within a 20 minute period centered on the radar analysis time are also indicated. It is clear that many of the triangulated return stroke positions were incorrect since it is unlikely that the lightning discharges were associated with areas of very low reflectivities, or areas where no radar echoes were detected. Yet, the more general area of occurrence for the flashes appears plausible given the magnitude of the radar reflectivities (i.e., 45-50 dBZ) observed for some of the cells located in the same general vicinity.

To further examine the relative accuracy of the triangulated CG positions, we overlaid triangulated CG return stroke locations and GMS satellite imagery collected on 9 February 1993 (1245 UTC; Fig A2). Again, note the *broad* agreement near 6°S, 158°E between satellite brightness temperatures (proxy for cloud-top height) and the CG return stroke pattern. However, the triangulated lightning locations appear suspicious due to the relatively large scatter of the stroke locations in both cloudy and clear areas of the satellite imagery. Clearly, a method

for more accurately locating the return stroke positions including quantification of the site errors associated with each antenna location was required.

A.2 Development of TOA methods for locating cloud-to-ground flashes

To calculate more accurate CG stroke locations it was necessary to develop a method which did not rely solely on the azimuth bearings to flashes since the bearings were contaminated by site errors. This was accomplished independently at CSU and A&M by developing algorithms for locating CG return strokes based on time-of-arrival methods (which do not require azimuth bearings). At A&M, the return stroke positions were computed using analytical solutions to the polynomial equations governing the intersection of two hyperbolas (hyperbolas determined by differences in the stroke detection time between two pairs of antenna's). At CSU the same polynomial equations governing the intersections of two hyperbolas were developed, but the stroke locations (roots to the equations) were computed in a "hybrid" algorithm that used stroke locations previously calculated by triangulation as an *initial guess* in the time of arrival type solution. We use the word "hybrid" because TOA techniques typically rely exclusively on differences in arrival time between at least three (or more) antennas and the intersection of the resulting two (or more) hyperbolas to determine a flash location (e.g., Bent and Lyons, 1984; Holle and Lopez, 1993).

For the "hybrid" approach, the triangulated stroke locations were used to specify an initial interval (100 km radius from the triangulated position) for finding the root of the equation governing the intersection of two hyperbolas defined by the specific differences in time of arrival for a flash signal at different pairs of ALDF's. In this manner, the roots to the polynomial equations (i.e., the stroke locations) could be computed quickly and efficiently using the specified

interval in a simple bisection algorithm. Since the initial guess for the return stroke position was assumed to be relatively close to the actual location of the stroke (i.e., within 100 km), the probability of calculating incorrect solutions due to the geometrical ambiguities involved (shown in Fig. A3) when using only three antennas for a TOA-type solution (Holle and Lopez, 1993) should be reduced. Indeed, with the exception of strokes that occurred very close to ALDF sites ambiguous location errors do not appear to be a problem. The method employed in this "hybrid" technique is conceptually similar to the IMPACT method discussed in Cummins et al., (1995) for the NLDN lightning location network in the United States.

Since the stroke positions were computed independently using slightly different algorithms (i.e., at CSU and at A&M), it was possible to compare the positions. The location comparisons revealed differences in CG stroke positions of only 1-3 km within a range of approximately 500 km from the center of the network. At longer ranges (i.e., 500-900 km from the network), the computed positions differed by as much as 7-10 km. Subsequent analysis revealed that the differences found to exist between the computed stroke locations in the two algorithms (especially at long ranges from the network) were due primarily to small differences in the constants and precision defined in the computer programs used to solve the intersection equations (e.g., the radius of the earth, rotation angles etc.). Given the relatively small differences in stroke location as a function of range between the two algorithms, it is clear that the two methods produced stroke locations that compare quite favorably for ranges approaching 900 km from the network.

An example of the marked increase in location accuracy due to implementation of the TOA algorithm can be inferred from Fig. 4.7 (case discussed in Chapter 4, Sec. 4.3.2), which is the same plot of radar reflectivities shown in Fig. A1, with the important exception that the overlaid CG stroke positions were computed using the hybrid TOA method. Note that the stroke

positions are now clustered around more intense convective cells, located approximately 10 and 60 km northeast of the radar ($x=10$ $y=15$ and $x=55$ $y=45$ respectively). In particular, the cell and associated CG's closest to the radar were observed both from a visual and electronic perspective, the latter by an electric field mill operating on board the R/V Vickers.

We can gain further appreciation for the increased accuracy of the TOA techniques by examining the same satellite data shown in Fig. A2, but with the positions of CG flashes calculated by the hybrid TOA technique overlaid (Fig. A4). In Fig. A4 the CG flashes situated under the eastern edge of the deep cloud system at 6°S 158°E are now tightly grouped into a mesoscale region (compare to Fig. A2). Note that the absolute number of CG flashes located with the TOA technique (157) is smaller than the 204 flashes located using triangulation. This is a reflection of the requirement for at least three antennas to detect a CG before that flash can be located using a TOA algorithm (approximately 50 of the CG's in Fig. A2 were detected by only two antennas). Over a large sample of CG flashes ($\sim 25,800$) we found that approximately 65% of the flashes detected by two of the COARE ALDF antennas were detected by all three antennas.

Based on the algorithm error checks and on subsequent plots of lightning, satellite, and radar information, we estimate the location accuracy of the TOA methods for the COARE ALDF network to be within 10 km of the true stroke location at ranges of 500-900 km from the network, and within 2-5 km of the true location for ranges within 500 km of the network. These errors are remarkably low given the large distances over which cloud-to-ground lightning was detected by the network. Given the relatively small errors estimated for the location accuracy, and the fact that no site error measurements could be performed for the network during installation, it seemed reasonable to use the TOA solutions for estimating antenna site errors.

A.3 Estimation of antenna site errors

For a given CG return stroke detected by all three antennas, the algorithm for computing site errors (systematic errors in the measured azimuth to a flash for a given antenna) assumes that the stroke location determined by the TOA method is the actual location of the stroke. Subsequently, the azimuth to the assumed "true" location of the stroke can be computed for each antenna and then compared to the measured azimuths. The *mean* site error (with outliers removed) was then computed for 72 five degree intervals (72 bins) by averaging the number of azimuth errors per bin for 26 separate days over a six month time period from January-June, 1993. Subsequently, a site error curve (i.e., site error as a function of azimuth) and associated statistics such as the standard deviation of the site error for each azimuth bin could be created for each antenna (e.g., Kavieng ALDF; Fig. A5).

After examining the standard deviations of the site errors for many different azimuth bins (e.g., Fig. A5) and then plotting the frequency distributions of the site errors for different azimuths, we found that: 1) the error distributions were associated with standard deviations of approximately 2 degrees in the majority of the bins; and 2) in some bins the number of individual azimuth errors was not a large enough sample (i.e., < 30) to obtain accurate estimates of the mean and standard deviation of the site error for the bin. Visual inspection of the distributions (where a considerably higher number of strokes occurred) indicated that at least 30-50 strokes were required to obtain an accurate estimate of the mean site error for a bin. Standard deviations of 2° in the site error distribution for a particular azimuth seem reasonable (e.g., Tyahla and Lopez, 1994) given that random errors in the measured azimuth for a single ALDF antenna are approximately 1° , and an additional error is introduced by our assumption that the TOA positions are the "true" locations of the strokes (i.e., in reality there is also an error

associated with the locations calculated using the TOA method, though we have no way of directly assessing the value of that error).

Once site error curves were created, the errors were applied to the ALDF azimuths in an attempt to correct the antenna azimuths used in the triangulation algorithms. The triangulated positions were then recalculated and compared to the TOA solutions again. We found that for ranges < 300 km from the center of the network, the corrected triangulation's were within approximately 10-15 km of the TOA solutions, on average. However, the average errors found to exist between the corrected triangulations and the TOA solutions were typically much larger than 10 km at ranges greater than 300 km from the network. This would be expected even if all of the site error standard deviations were within 2° of the mean (i.e., a 2° error in azimuth places an azimuth bearing approximately 10 km from the true location of the stroke at a range of 300 km). Thus for ranges *exceeding* 300 km from the center of the TOGA-COARE ALDF network, the accuracy of CG stroke positions calculated using *only triangulation* limits the application of the data to studies on larger scales that do not require accuracies of better than 10 km.

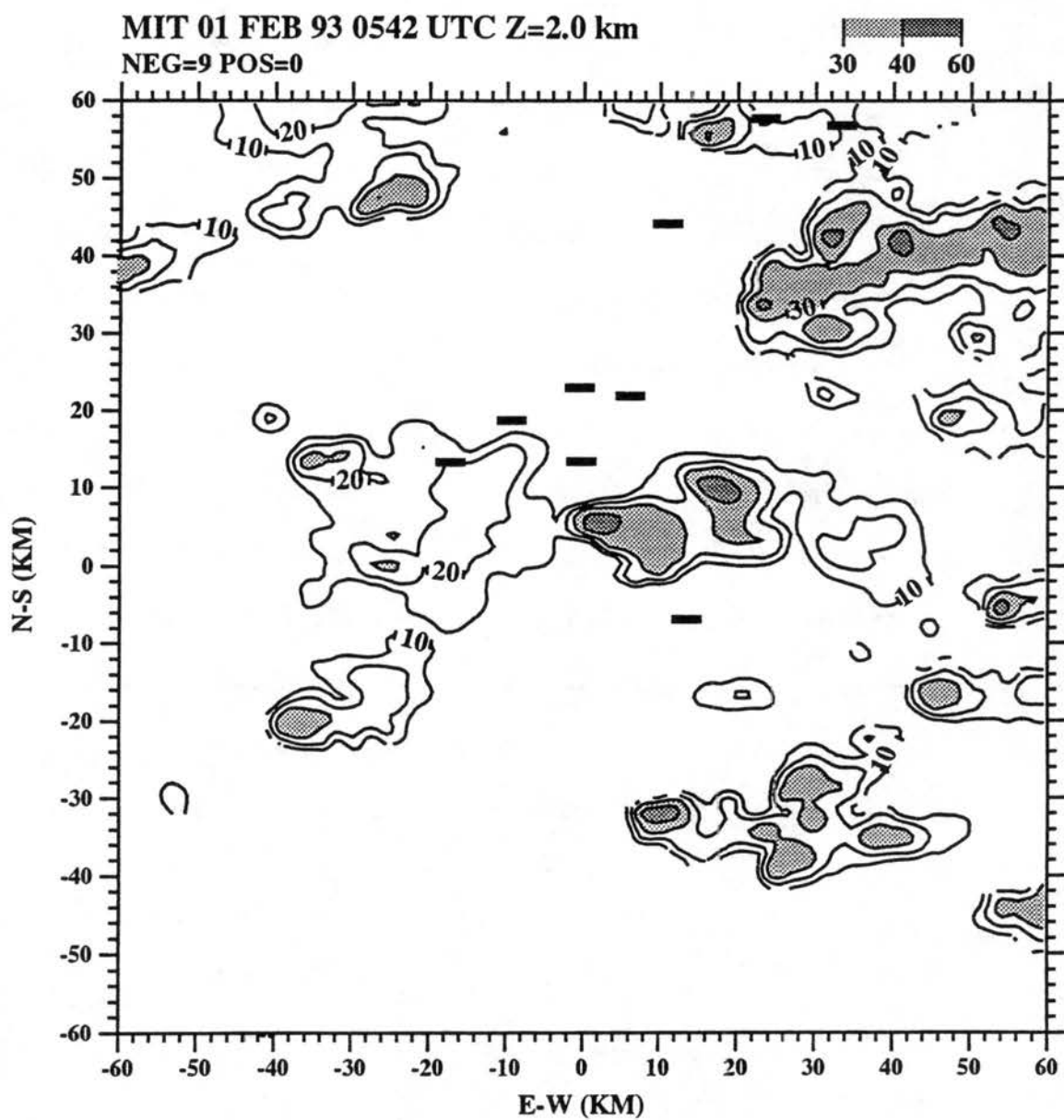


Figure A1. As in Fig. 4.7 but with CG return stroke positions calculated using triangulation.

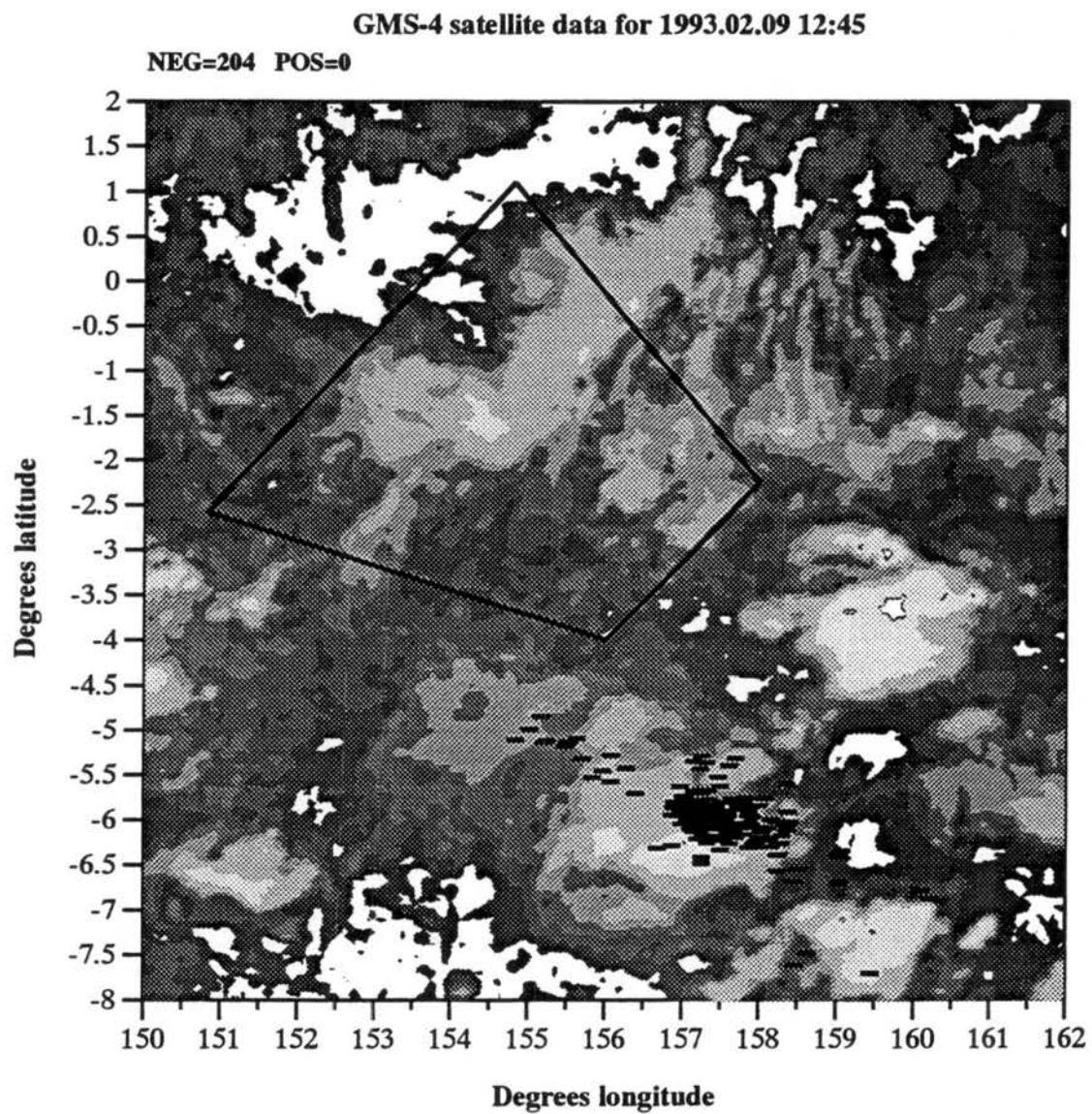


Figure A2. As in Fig. 4.1 but 1245 UTC. Lightning positions were calculated using triangulation.

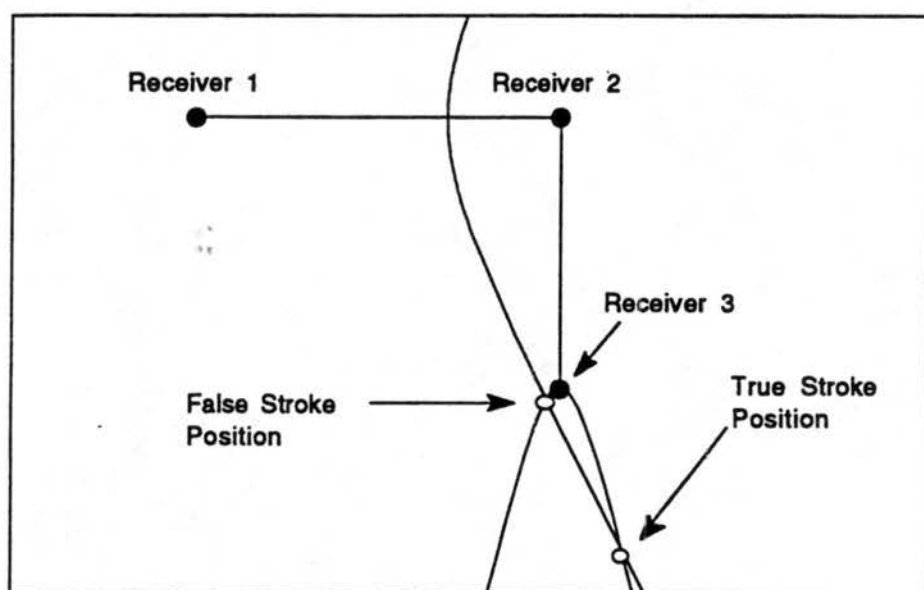


Figure A3. Illustration of ambiguous location errors that can be encountered when only three DF antennas are used with the time-of-arrival solution. Note the two intersections of the hyperbola branches, and the true stroke position behind receiver 3. (reproduced from Holle and Lopez, 1993).

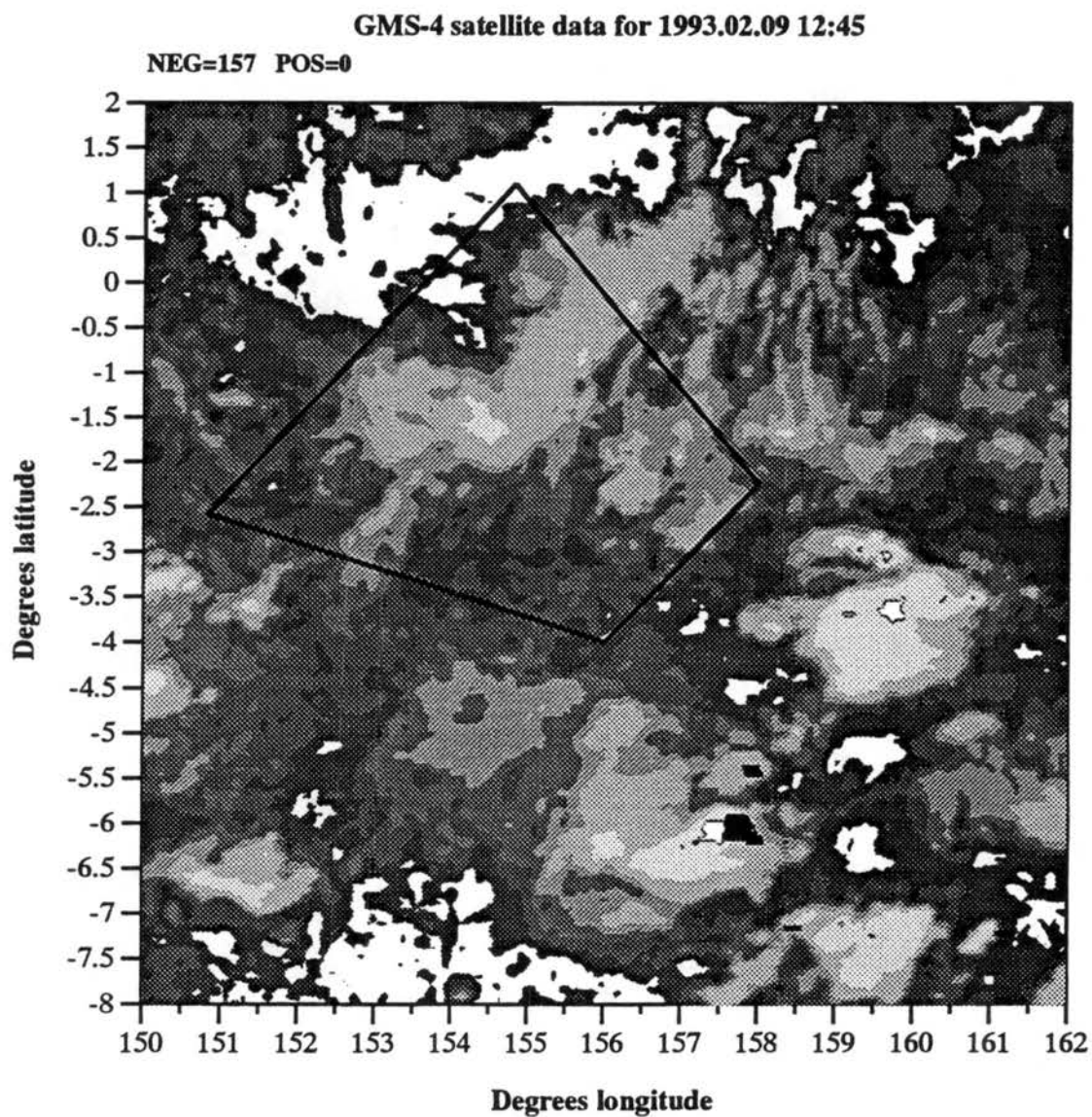


Figure A4. As in Fig. A2 except the return stroke positions were calculated using the hybrid TOA algorithm.

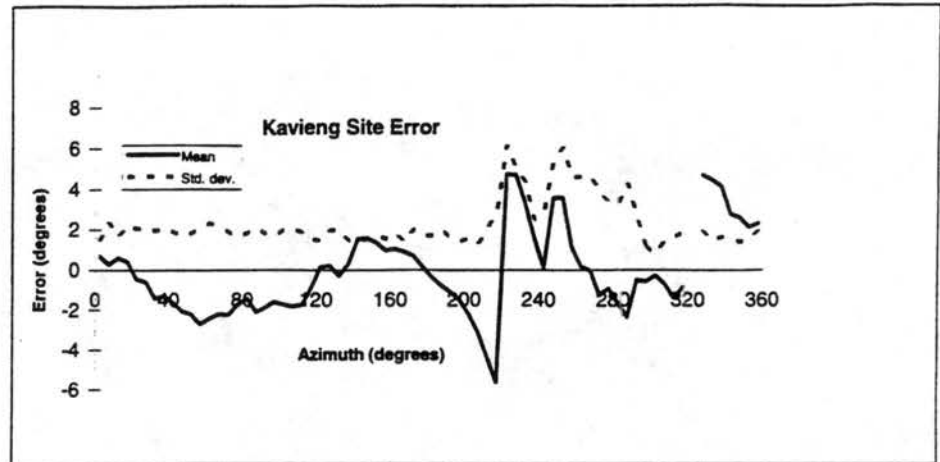


Figure A5. An example of the site error curve (solid line) vs. azimuth for Kavieng. The curve is based on data from the 26 most active lightning days from January through June, 1993. The site error is evaluated for each five degree bin from zero to 360 degrees. The standard deviation (dashed curve) for each five degree azimuth bin is evaluated and plotted as a function of azimuth.