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14 **Abstract**

15 When cumulonimbus clouds aggregate, developing into a single entity with precipitation
16 covering a horizontal scale of hundreds of kilometers, they are called mesoscale
17 convective systems (MCSs). They account for much of Earth’s precipitation, generate
18 severe weather events and flooding, produce prodigious cirriform anvil clouds, and affect
19 the evolution of the larger-scale circulation. Understanding the inner workings of MCSs
20 has resulted from developments in observational technology and modeling. Time-space
21 conversion of ordinary surface and upper-air observations provided early insight into
22 MCSs, but deeper understanding has followed field campaigns using increasingly
23 sophisticated radars, better aircraft instrumentation, and an ever-widening range of
24 satellite instruments, especially satellite-borne radars. High-resolution modeling and
25 theoretical insights have shown that aggregated cumulonimbus clouds induce a mesoscale
26 circulation consisting of air overturning on a scale larger than the scale of individual
27 convective up- and downdrafts. These layers can be kilometers deep and decoupled from
28 the boundary layer in elevated MCSs. Cooling in the lower troposphere and heating aloft
29 characterize the stratiform regions of MCSs. As a result, long-lived MCSs with large
30 stratiform regions have a top-heavy heating profile that generates potential vorticity in
31 midlevels, thus influencing the larger-scale circulation within which the MCSs occur.
32 Global satellite data show MCSs varying in structure, depending on the prevailing large-
33 scale circulation and topography. These patterns are likely to change with global
34 warming. In addition, environmental pollution affects MCS structure and dynamics
35 subtly. Feedbacks of MCSs therefore need to be included or parameterized in climate
36 models.

1. Introduction: Hints of mesoscale convection

When Huckleberry Finn told the story of floating down the Mississippi River on a raft, he said, “The fifth night below St. Louis we had a big storm after midnight, with a power of thunder and lightning, and the rain poured down in a solid sheet.” Before he wrote this story in 1884, Mark Twain had worked for some years on Mississippi River boats—long before scientific research on storms had gotten underway and revealed that *mesoscale convective systems* (MCSs) have a tendency to form just east of the Rocky Mountains in the afternoon or evening and propagate eastward and produce heavy rain over the Mississippi basin during nighttime. The keen observations of this experienced riverboat man presaged a century of research on the nature of convective storms over the central U.S. and elsewhere in the world. We have since learned that MCSs are not confined to the U.S. Midwest, but actually are important elements of the global circulation and water cycle as well as one of the world’s most significant producers of extreme weather. We also know from satellites and radars that MCSs take on a variety of forms, depending on whether they occur over ocean or land, near mountains, over plains, in the paths of moist jets, or in a variety of other circumstances. This review chronicles what have learned about these important atmospheric phenomena and how this knowledge has arisen from technological advances in meteorological observations over the last century.

In the 19th century, a few hints about mesoscale organization of convection also cropped up in the scientific community. By studying some of the very first cloud photography, the Scottish meteorologist Ralph Abercromby, in collaboration with the

Swedish meteorologist Hugo Hildebrand Hildebrandsson (Abercromby 1887; Hildebrandsson 1887), identified the cloud form that is now generally recognized as cumulonimbus. Earlier, Luke Howard (1803) had identified cumulus as a basic cloud type seen by eye from ground or ship. Upon studying his photographs, many taken by himself in his travels around the world, Abercromby thought that the term cumulonimbus was needed to identify the “mountainous cumulus which discharges rain in showers and thunderstorms” to separate this cloud type from “the cumulus of a fine day from the similar cloud of showery day.” It is perhaps less well known that these pioneers also identified the anvil clouds of cumulonimbus. Hildebrandsson described the upper part of the cloud as “towering up to colossal proportions as mountain ranges, or a gigantic mushroom, with a flat layer of ‘false cirrus’ around or on the top.” But in this early age, without satellites, radar, or aircraft, it was hard for the extreme horizontal dimensions of these storms to be fully recognized.

As a result, these 19th century literary and early scientific harbingers mostly did not comprehend the massive scale on which convective clouds can become organized. One exception appeared when, in 1883, at a meeting of the American Association for the Advancement of Science in Minneapolis, University of Iowa physicist Gustavus Hinrichs gave the first known description of the straight-line winds behind gust fronts, which are a common feature of MCSs moving across the plains states of the Midwestern U.S. He found that these gust fronts extended over 10s to 100s of kilometers.² He formally published his work in a two part article in the long defunct *American Meteorological Journal* (Hinrichs, 1888a,b). Another early study of note was that of W. J. Humphreys

² He even introduced the use of the Spanish term *derecho*, which is now commonly used to describe certain long-lived windstorms.

(1914), who worked with the nascent U.S. Weather Bureau under the title of Professor of Meteorological Physics. He speculated on the structure of a thunderstorm with a sketch indicating a period of heavy rain followed by lighter rain, but the essential horizontal and vertical scale and circulation characteristics of MCSs would not be fully recognized for another thirty years.

2. World War II and the amazing work of Hamilton and Archbold

Working in Nigeria during WWII, two Royal Air Force officers supporting wartime air operations developed an extensive manual on the meteorology of tropical western Africa (Hamilton and Archbold 1945). Based on pilot reports, balloon soundings, and surface meteorological data, they determined the mesoscale nature of convective storms characterized as spatially extensive convective “disturbance lines.” From the observed speed and duration of the events passing a point, it was apparent that the rainfall zones were ~75-150 km in width. Synoptic analysis of surface station reports showed that the leading windshift lines were up to 1000 km in length and bowed outwardly in the direction of storm motion. These rudimentary observations indicated that the convection was organized on a spatial scale much larger than that of isolated convective clouds yet considerably smaller than synoptic scale, i.e. the disturbance lines were the leading edges of mesoscale storms.

The early paper by Hamilton and Archbold (1945) also captured the essential character of the storm-scale overturning air motions within these mesoscale systems. Panel a of Fig. 1 shows the wind pattern at the ground as they deduced it from surface meteorological stations in West Africa, while panel b is an accompanying vertical section showing the cloud outline and system-relative air motions in a plane normal to the line.

Hamilton and Archbold did not indicate the horizontal and vertical scales in these particular figures, but from their maps of specific events and statements in the text, we know that the windshift line and width of the circulation features normal to the line are mesoscale in horizontal extent. They did not have radar or other observations to show the embedded convective scale structures. As we will see, much of the research on the mesoscale aspects of MCSs subsequent to the study of Hamilton and Archbold has employed information from radar, aircraft, satellite, and numerical models to provide details and variations on the theme comprised by the cloud outline and circulation features illustrated in Fig. 1, but modern technology has not changed their fundamental conclusions. Nonetheless, details are important, and we have learned much more about the characteristics of MCSs and their role in the large-scale circulation of the atmosphere since the time of Hamilton and Archbold.

3. The emergence of mesometeorology and radar in the 1950's

Unfortunately, the study of Hamilton and Archbold that had laid such a clear foundation for understanding the mesoscale organization of atmospheric convection was barely noticed and largely forgotten after 1945—possibly as part of the general global societal chaos and reconfiguring of the scientific community after WWII. Research on convective storms in midlatitudes nevertheless was taking hold in the U.S. with the massive post-war Thunderstorm Project led by the University of Chicago Professors Byers and Braham (1949). In what remains one of the largest and most innovative field projects in the history of meteorology, they deployed aircraft, radars, and radiosonde

units left over from the war to observe storms in Ohio and Florida.³ The Thunderstorm Project revealed the nature of the individual up- and downdrafts embedded in convective storms. The mesoscale characteristics of the convection observed in the Thunderstorm Project were harder to decipher because, unlike the tropical convection studied by Hamilton and Archbold, midlatitude convective processes were difficult to disentangle from other factors—especially, frontal dynamics in Ohio and sea breeze dynamics in Florida.

Polar-front theory had dominated midlatitude meteorological thought since the 1930s, and a first impulse of researchers was to apply frontal reasoning to large convective systems. It was widely recognized by this time by U.S. weather analysts that thunderstorms in midlatitudes lined up parallel to, but often ahead of, cold fronts and that these lines of storms extended over meridional distances of thousands of kilometers. An example of applying frontal reasoning to these great lines of storms is the study of a case observed in the Thunderstorm Project by Chester Newton (1950), one of the greatest meteorologists of the era. He presented the cross section analysis of Thunderstorm Project radiosonde data collected in Ohio in Fig. 2 in which he drew isotherms according to the rules of standard synoptic frontal analysis, with first order discontinuities defining the boundaries of a frontal zone (indicated by heavy lines). Reliance on synoptic surface meteorological observations and polar front thinking also led Tepper (1950) to try to

³ The Thunderstorm Project used twenty-two freight cars full of instrumentation, including radiosondes and radars, plus numerous trucks and jeeps. In addition ten Northrup P-61C Black Widow aircraft, flown by “highly competent instrument pilots of the Air Force” and gliders flown by pilots from the Soaring Society of America volunteered for aircraft operations. Although enormous in terms of platforms, the instruments used were generally of a primitive nature, thus limiting the success of the program. Nearly three decades went by before another massive effort was undertaken to understand convection.

145 explain the pressure rise accompanying the line of storms ahead of a front as the result of
146 a gravity wave being triggered at the cold front and moving out ahead of the front. We
147 now know that this pressure rise is mainly associated with the spreading downdraft cold
148 pool of the storms, and that deep convective clouds breaking out by the release of
149 instability in the warm sector air ahead of a cold front tend to arrange themselves in lines
150 ahead of an approaching cold front for a variety of synoptic and mesoscale reasons that
151 would have been difficult to decipher by standard synoptic analysis and frontal reasoning
152 alone.

153 A paradigm shift in analysis methods occurred when Tetsuya (Ted) Fujita
154 emigrated from post-WWII Japan and joined the Chicago School of Byers and Braham.
155 In an innovative paper (Fujita 1955), he introduced the community to *mesometeorology*,
156 in which meteorological observations were painstakingly combined by space-time
157 conversion to produce horizontal and vertical cross sections that revealed meteorological
158 structures on subsynoptic scales. It was in this way that he showed that a line of
159 thunderstorms along or ahead of a cold front was not a kind of front in the sense of polar-
160 front thinking, but rather a lining up of convective entities organized individually on the
161 mesoscale. In today's lexicon, we would say that he identified the MCS as the building
162 block of the prefrontal or frontal squall line. As Fujita put it, "The mesosynoptic
163 disturbances greatly influence the situation as viewed on the regular synoptic scale,
164 which is about 10 times the mesoscale, and make conventional [synoptic] analysis
165 hopelessly difficult." Figure 3 shows the conceptual model of an MCS as inferred by
166 Fujita. An important feature of this conceptual model is the horizontal scale of the
167 storm's footprint, which was ~300 km. This model was thus comparable in scale and air

motion characteristics to the tropical systems analyzed by Hamilton and Archbold (1945). Fujita's detailed analysis of an example case in his 1955 paper showed that the line of storms occurring just ahead of a cold front consisted of a series of entities like that of the conceptual model rather than resembling a variety of polar front as implied by analyses such as that of Fig. 2. Because time series often capture mesoscale events better than scheduled observations, time-to-space conversion has remained an important method of mesometeorological analysis through the years; notable examples would be the highly referenced work of Fujita's protégé Roger Wakimoto (1982) and the more recent study of Adams-Selin and Johnson (2010).

4. The rise of radar meteorology

Even using his clever time-space conversions, Fujita could not reveal anything about the internal structures of MCSs on smaller scales because his mesometeorological methods did not benefit from targeted aircraft measurements like those in the Thunderstorm Project, or from radar, which was still emerging from its development in WWII to become one of the most important meteorological instruments of the last 100 years. One of the pioneers of radar meteorology was the relatively unheralded Herbert Ligda, who worked at the Massachusetts Institute of Technology, Texas A&M, and Stanford, but published relative little in the formal literature. A remarkable finding of Ligda during his time at Texas A&M was his discovery of the internal structure of the same types of MCSs described both by Hamilton and Archbold and by Fujita. Figure 4 from a Texas A&M project report was published in the non-refereed proceedings of a conference of glider pilots (Ligda 1956). It showed in schematic but amazingly precise form the details of the typical radar echo pattern of an MCS of the type analyzed by

Hamilton and Archbold (1945) and Fujita (1955). Notable features were a narrow sharp line of weak echo (A) marking the gust front immediately ahead of a convective line (B), which was advancing with an eastward component of motion and consisted of numerous intense convective elements, each elongated northwest to southeast. The feature D was a trailing region of stratiform precipitation, separated from the line of convective cells by a zone of weak echo (C). Houze et al. (1990), still unaware of Ligda's early study hidden in non-standard literature, arrived at an essentially similar conceptual model of the radar echo in an MCS of this type. Radar was thus beginning to show that an MCS, which had been determined by Hamilton, Archbold, and Fujita, to be of a horizontal scale of a few hundred kilometers, contained important substructures on a range of smaller scales, all of which have turned out to be important elements when considering the role of MCSs as weather producers and elements of larger scale circulations.

At this point of the review, I would like to emphasize that MCSs are *not always in the form of lines of intense convective cells followed by a zone of stratiform precipitation*, as in the examples studied by Hamilton and Archbold, Newton, and Ligda. Such storms, often called *squall-line MCSs*, attract attention because they are dramatic—as they pass over, they are marked by a predictably ordered sequence of events: a sudden windshift (called a “squall front” or “gust front”), followed by a short period of heavy rain or hail, and then a longer period of quasi-steady lighter stratiform rain. As such, this type of MCS is especially amenable to analysis and modeling—and great fun to analyze. However, MCSs of similar dimension can exhibit a variety of more complex patterns of convective cells and stratiform rain. Tropical studies carried out long after Hamilton and Archbold's (1945) work have shown both squall-line and non-squall-line MCSs (Leary and Houze

1979b; Houze and Betts 1981; LeMone et al. 1998; Kingsmill and Houze 1999; Yamada et al. 2010; Barnes and Houze 2016). In midlatitudes, Houze et al. (1990) noted that non-squall-line MCS were seen in about 1/3 of the major springtime rainstorms in Oklahoma.

Much of the radar-based work on the convective/stratiform patterns of MCS precipitation has been over the central U.S., which raises a cautionary concern because this region of the world is unique so that the MCSs in this region are likely not universally representative of MCSs around the world. The MCSs in the U.S. are affected by significant geographic factors that affect the forms taken by the convective storms. The Rocky Mountains to the west, the mountains of Mexico to the southwest, and the Gulf of Mexico to the south are the most important of these geographical features. Combined with the semi-permanent Atlantic subtropical anticyclone, these geographical features favor a moist low-level jet that systematically feeds warm, moist air into MCSs in this region. The only other somewhat similar region on Earth is Argentina where the South American low-level jet emanating from Amazonia and guided by the Andes to the west feeds MCSs in that region. Rasmussen and Houze (2011) found that MCSs in subtropical South America take on forms very similar to those over the central U.S. In general, however, MCSs are a global phenomenon, and while the evolution and forms of MCSs seen over the central U.S. and subtropical South America may be useful for forecasting in these particular regions, some of the characteristics of MCSs seen in these regions may not apply universally over the globe.

With this geographical caveat in mind, we note some milestones of radar-based studies of MCS structure over the U.S. Although the convective cells in MCSs are not always arranged in lines, such lines of convection are nevertheless of special interest to

forecasters. Bluestein and Jain (1985) found that over the central U.S., lines of deep convective cells form in four ways, but they did not consider the relation of the cells to stratiform precipitation in MCSs. Houze et al. (1990) examined the convective/stratiform configurations of the precipitation in MCSs over Oklahoma. The example from their study in Fig. 5a has leading-line/trailing-stratiform structure of the type described by Ligda (1957). Houze et al. (1990) called this form “symmetric.” However, the radar echoes in Fig. 5b were said to be of an “asymmetric” form in which the stronger convective cells were on the equatorward end of the line and the stratiform region was trailing the poleward portion of the line. Skamarock et al. (1994) found the asymmetric structure to be favored by the action of the Coriolis force acting on the time scale of the MCS, which probably accounts for this asymmetric structure not appearing (at least to this author’s knowledge) in the tropics. Loehrer and Johnson (1995) and Parker and Johnson (2000) further examined the symmetric and asymmetric paradigms in midlatitude continental MCSs over the region east of the Rocky Mountains in the U.S. They found that the most common life cycle scenario was an initial line of convective cells developing a stratiform region, first in a symmetric juxtaposition with the line and then evolving into an asymmetric form, as evidently the Coriolis force had longer to act. The second most common evolution had the only stratiform precipitation forming on the northeast end of the convective line, as old cells weakened and new ones formed on the southwest end of the line.⁴ Also analyzing radar data over the central U.S., Jirak et al. (2003) found that MCSs evolving from linearly arranged convective cells were longer-lived, more severe, and more effective at producing precipitation than MCSs that

⁴ This behavior is similar to the rainbands in tropical cyclones (Hence and Houze 2008; Houze 2010; Didlake and Houze 2013).

developed from areally arranged convection. That result is probably very specific to the U.S. Schumacher and Johnson (2005) and Peters and Schumacher (2015, 2016) have distinguished additional variations on the convective/stratiform structures of MCSs, especially in flood-producing MCSs. However, the structures and circulations these authors identified remain entangled with midlatitude frontal activity that also strongly influences convection over the region east of the Rocky Mountains.

5. The 1960s–early 80s: the beginning of the satellite era

Ground-based radar studies, such as those described above, are intrinsically regional in scope. The global significance of MCSs and their convective/stratiform structures has followed from developments in satellite meteorology, which allows global analysis of the frequency of occurrence of MCSs of various types. This opportunity arose with the launch of the first weather satellite in 1960 (Anderson 2010). One of the first things to be noticed was that cirrus shields in regions of deep convection were mesoscale in extent. Martin and Karst (1969) and Martin and Suomi (1972), using digitally enhanced visible images, were able to track “cloud clusters” which lasted for 3-6 days. These clusters were regions of high cloud that were 3000-7000 km in horizontal scale, i.e. they were nearly synoptic-scale features. However, these early investigators also noted bright cores within these cloud shields that lasted 1-4 h, moved slower than the clusters, and sometimes were in the form of bands. These bright cores were evidently the active convective entities that we would now identify as MCSs. Another important early satellite study was that of Frank (1970), who found that cloud clusters over the tropical Atlantic were systematically associated with the troughs of synoptic-scale easterly waves, which in turn are the foci of tropical cyclone genesis. But this study did not address the

subsynoptic substructure of the cloud clusters.

Visible satellite imagery is limited to daytime and is largely non-quantitative. The first clear identification of MCSs in satellite data arose from the use of infrared imagers on satellites, which provide an indication of cloud-top temperature. Using early infrared imagery, Maddox (1980) identified what he called *mesoscale convective complexes*, or MCCs. These entities are defined as large, circular, cold cloud tops whose infrared brightness temperatures are lower than -32°C over an area $>100,000\text{ km}^2$ (radius of 178 km) and lower than -52°C over an embedded area $>50,000\text{ km}^2$. The peculiar choices of thresholds of -32 and -52°C arose from the fact that the work was done using photographic operational satellite products that used these particular isotherm values, as can be seen in the illustration of an MCC in Maddox's classic paper (Fig. 6).

An ensuing series of papers (synthesized by Laing and Fritsch 1997) showed where MCCs identified in this way occur around the world. However, we now know that MCCs are an extreme form of mesoscale convection. Many deep convective systems bearing a large amount of precipitation and high-impact weather have horizontal scales in which precipitation covers areas $\sim 100\text{-}500\text{ km}$ in horizontal scale without satisfying the strong cold cloud top criteria of Maddox (1980), and the term MCS is commonly used (including in this review) to encompass this broader population of mesoscale entities of which MCCs are an extreme subset. Anderson and Arritt (1998) also pursued this type of analysis. They provided a catalog of information on the seasonal and diurnal frequency occurrence of MCCs as well as of MCSs of similar intensity but whose upper cloud shields were not circular but elongated.

Visible and infrared imagery were just the beginning of the impact of satellites on

the study of MCSs. As we will see in later sections of this review, the late 1990s and 2000s saw the introduction of radars on satellites, which further revolutionized understanding of MCSs and their global importance. But first we need to note the importance of the grand field experiments in the tropics.

6. The 1970s: The far reaching impact of the Global Atmospheric Research Programme's Atlantic Tropical Experiment (GATE)

By the 1970s, Lorenz's (1963) work on predictability had led to the idea that prediction of global weather up to two weeks in advance could be accomplished if convection in the tropics could be better parameterized. This grand hypothesis required a grand experiment to achieve better understanding of the interaction of convective and synoptic scales of motion in the tropics, and an ambitious international field project was organized. That project was the 1974 Global Atmospheric Research Programme's Atlantic Tropical Experiment (GATE, Kuettner and Parker 1976), carried out over the eastern tropical Atlantic Ocean, off the western coast of Africa.

GATE was designed to reveal simultaneously properties on scales ranging from convective up- and downdrafts of the type documented in the Thunderstorm Project to the synoptic scale motions of near-equatorial easterly waves that control the occurrence of convection in this part of the tropics. A massive effort was required to document the large-scale environment by a shipborne sounding network, the convective structures by radars on ships, mesoscale air motions and in-cloud properties sampled by aircraft, and a variety of radiation and boundary-layer flux measurements. The Thunderstorm Project in the late 1940s had shown the effectiveness of deploying such special observations in an organized community effort. In the decades following the Thunderstorm Project,

instrumented aircraft had further proved themselves by flying into hurricanes (Dorst 2007), and some small field programs involving radars and/or aircraft in the tropical Pacific (Zipser 1969), the Caribbean (Holland 1970), and the Atlantic (Shupiatsky et al. 1976a,b) had been successful. GATE brought such fieldwork to an absolutely massive scale. Gathered in the eastern tropical Atlantic were the atmospheric research resources of 20 nations, deploying 12 large research aircraft and 40 research ships (www.ametsoc.org/sloan/gate). Four of the ships hosted a network of three-dimensionally scanning weather radars making and recording quantitative reflectivity data over a four-month period (Hudlow 1979) while soundings were launched from 14 ships surrounding the area of radar observations (Thompson et al. 1979). By this time, geosynchronous weather satellites were in orbit, providing the evolving cloud context within which the measurements were made. The coordinated soundings, radars, aircraft and satellite measurements in GATE led to a new appreciation of MCSs.

Ironically, the experiment design of GATE was founded on the conceptual model of convective/synoptic interaction being between convective updrafts and synoptic-scale motions, as expressed by Yanai et al. (1973) and Arakawa and Schubert (1974). In this view, mesoscale convection was not considered part of the interaction. However, the shipborne radar network showed that both squall-line and non-squall MCSs were major components of the cloud population observed in the GATE observational array (Houze and Betts 1981), and the extensive multi-scale observational array of GATE laid a foundation for research on MCSs that continues to this day. GATE studies produced fundamental knowledge of the thermodynamics, precipitation mechanisms, air motions, microphysical processes, dynamics, life cycle stages, heating processes, and relationship

to larger scales of motion. While knowledge of each of these aspects of MCSs arose out of GATE studies or work inspired by GATE, it has advanced in concert with technological developments in radar, modeling, satellite remote sensing, and additional field programs—lesser in scope than GATE but more specifically targeted, with more sophisticated instruments, and including midlatitude venues. The next topics of this paper will therefore begin with GATE as the starting point and then discuss how knowledge of the topic has been refined since that time.

7. Precipitation in MCSs: The universality of the convective/stratiform paradigm

A major result of GATE was the finding based on the four shipborne quantitative radars that the precipitation consisted of two primary components, convective and stratiform. Prior to GATE, there was no recognition that any tropical precipitation was stratiform. Each of the GATE radars was operated in a tilt-sequence mode, in which the antenna rotated 360° in azimuth at a series of elevation angles. The three-dimensional echo coverage obtained continuously in this way showed that when MCSs passed over a ship, short periods of heavier precipitation had radar echoes in the form of vertically oriented cores, but that some 40% of the precipitation was of a more moderate nature and was characterized on radar by a horizontally oriented layer of maximum reflectivity just below the 0°C level (Houze 1977; Leary and Houze 1979a; Cheng and Houze 1979). The schematic in Fig. 7 illustrates the “bright band” produced by melting of large ice particles, which was a clear indication that a large proportion of the rain was stratiform in nature; i.e. the vertical air motions were generally not strong enough to support vertical advection of precipitation particles. Instead, ice particles were systematically drifting downward and turning into light-to-moderate rain over very large areas. Figure 8

illustrates how the horizontally integrated rainfall (sometimes called volumetric rain) from the MCS begins as convective and then develops a stratiform component that over time becomes equivalent in magnitude to the convective component. Following Houze (2014), precipitation is stratiform if the mean vertical air motion in a cloud is much smaller than the fall velocities of ice particles, so that ice particles are drifting downward, melting, and falling out as rain over a broad area. One way that stratiform precipitation forms is when active updrafts in a region of convection weaken and the precipitation particles aloft slowly fall out (Houze 1997). This process is how the stratiform regions of MCSs arise and is discussed further in Section 13. Oceanic MCSs of the type seen in GATE have now also been seen over the West Pacific (Petersen et al. 1999; Houze et al. 2000), the Maritime Continent region (Houze et al. 1981), and the central Indian Ocean (Barnes and Houze 2014, 2016).

GATE studies included studies not only of oceanic MCSs but also studies of MCSs over the African continent (Fortune 1980; Payne and McGarry 1977). These continental African MCSs have been documented in detail in a variety of studies (e.g. Fink and Reiner 2003; Schumacher and Houze 2006; Futyán and Del Genio 2007; Cetrone and Houze 2011; Powell et al. 2012), which show well-defined convective and stratiform behavior mainly of the leading-line/trailing-stratiform type.

GATE thus launched a direction of research on MCSs in the tropics that has paralleled studies of the structures of MCSs in the midlatitude region of the U.S. east of the Rocky Mountains, as was discussed in Section 3. Unlike the findings from the midlatitude studies, fronts and other baroclinic effects did not affect MCS structures in the tropics. The common denominator appearing from both the midlatitude and tropical

studies was upscale growth of a region of deep convection so that the rain covered an area of mesoscale extent, and the existence of a mesoscale region of stratiform precipitation occurring in association with the deep convection.

This convective/stratiform behavior was thus being documented over disparate parts of the globe, and always the MCSs exhibited precipitation distinctly subdivided into convective and stratiform regions. The development of satellite-borne radar has provided a way to link the nature of MCSs over the globe. The Tropical Rainfall Measuring Mission (TRMM) satellite launched in 1997 provided the first radar observations of the three-dimensional structure of precipitation from space (between 35°N and 35°S). TRMM orbited until 2014, but in that year the Global Precipitation Mission (GPM) satellite began providing similar radar measurements (between 65°N and 65°S). U.S. and Japanese teams responsible for processing the TRMM and GPM radar data have routinely separated the radar echoes into stratiform, convective, and “other” categories. The convective/stratiform separation algorithm used on the TRMM radar data (Awaka et al. 1997) was based both on bright band detection and on the horizontal texture analysis of Houze (1973), Churchill and Houze (1984), and Steiner et al. (1995). Mapping of TRMM radar data (Schumacher and Houze 2003) showed that ~30-70% of tropical rainfall was stratiform, with lower fractions over land and higher fractions over the oceans (Fig. 9). The “other” components of the TRMM radar echo patterns accounted for only a small proportion of the total rain over the low latitudes (Houze et al. 2015). Because the only source of stratiform rain in the tropics is primarily if not exclusively that occurring in MCSs, the large fractions in Fig. 9 showed that the latent heating in the tropics is substantially influenced by MCSs (see further discussions of TRMM in Sections 16 and

420 17, below.)

421 **8. Scales of air motions and thermodynamical processes within an MCS**

422 Another paradigm-shifting result of GATE was the realization that two types of
423 downdrafts occur in MCSs (Zipser 1977; Houze 1977). As indicated in Fig. 7, one type of
424 downdraft is highly local and occurs in connection with the locally intense convective
425 precipitation; it is forced by the weight of hydrometeors and can be locally strong. The
426 denser air from aloft forms a density current with a leading edge called the “gust front.”
427 The other type occurs in the stratiform precipitation areas and is less intense but more
428 widespread; it is driven by primarily by evaporation of falling hydrometeors, as first
429 demonstrated by Brown (1979) but also by sublimation and melting of ice particles
430 falling beneath the mesoscale stratiform cloud of the MCS (e.g. Braun and Houze 1997).
431 The difference between convective and mesoscale downdrafts in an MCS was elaborated
432 in the GATE study of Zipser (1977), which showed how both types of downdraft
433 transport air of lower equivalent potential temperature (θ_e) downward, with the
434 convective downdraft generally penetrating all the way to the surface (Fig. 10).
435 Sometimes a mesoscale downdraft that has formed in the stratiform region extends
436 downward and merges with the convective downdraft.

437 The updraft motions in an MCS also occur on two scales. This result was
438 substantiated with the maturing of Doppler radar technology in the 1980s. One of the
439 early dual-Doppler studies of MCSs was the Convection Profonde Tropicale 1981 field
440 program carried out with two Doppler radars in West Africa (Sommeria and Testud
441 1984). Dual-Doppler synthesis by Roux (1988) of the wind field in the convective region
442 of an MCS showed that cells of upward motion on the scale of a few kilometers and

corresponding to reflectivity cells of similar dimension were superimposed on a broader region of sloping ascent over the downdraft density current at lower levels (Fig. 11). A similar dual-Doppler experiment in the 1985 Oklahoma–Kansas Preliminary Regional Experiment for Stormscale Operational and Research Meteorology (PRESTORM, Cuning 1986) led to the conceptual model in Fig. 12, which describes a squall-line MCS. The Doppler radar syntheses showed that a general, mesoscale flow of air originating in the lower troposphere ascends through the convective region, where stronger local updrafts are superimposed, and into the stratiform region where the air is still generally buoyant with respect to the large-scale environment but without the strong intense convective cells seen in the convective region.

In the stratiform region, the general ascending mesoscale flow lies over the layer of general descent of the mesoscale downdraft, which is drawn in from the midlevels of the environment and is driven downward diabatically by sublimation, melting, and evaporation of precipitation. In the case of a squall-line MCS of the type conceptualized in Fig. 12, the midlevel inflow feeding the mesoscale downdraft was dubbed the “rear-inflow jet” by Smull and Houze (1987), who showed that this inflow is a ubiquitous feature of squall-line MCSs and may be intensified if the environmental shear favors strong flow relative to the storm in midlevels. As shown in Fig. 12, the stratiform region downdraft of a squall-line MCS flows horizontally toward the convective region, where the precipitation-driven downdrafts of convective cells dominate in the low levels. Doppler radar has shown that the mesoscale downdraft of the stratiform region of a squall-line MCS sometimes merges with the convective downdrafts in the leading line of convection and that these mergers can produce strong effects, with the gust front surging

forward and triggering new convection in the form of a “bow echo.” An example of a bow echo is shown in Fig. 13a from Davis et al. (2004) and Jorgensen et al. (2004). Doppler radar observations showed strong midlevel flow toward the back edge of the curved convective line. In Fig. 13b, a vertical cross section through the bow-echo portion of the line shows the descending rear inflow from the stratiform region of an MCS penetrating into the convective region, where it combined with the convective-scale downdraft and pushed the gust front forward, underneath the main updraft cell. These bow echo events are a major forecasting concern, as they are often associated with violent, damaging surface winds.

The broad, mesoscale overturning of the MCS is a major mechanism for vertical mixing of the atmosphere. Because it enters from midlevels, the mesoscale downdraft transports a large amount of midtropospheric low- θ_e air to lower levels, while the generally ascending mesoscale flow brings high- θ_e air from low levels to the upper troposphere. This mesoscale overturning is often thought of as a two-dimensional process as it occurs in the squall-line type of MCS. However, Doppler radar observations have shown that the mesoscale overturning actually occurs in a variety of three-dimensional configurations. In the 1980s, Doppler radar was implemented on aircraft to observe the winds in tropical cyclones (Marks and Houze 1983), and by the 1990s airborne Doppler was a proven technology, with dual beams allowing for dual-Doppler synthesis of the airflow in convection. The second major tropical oceanic field campaign was the Tropical Ocean Global Atmosphere Coupled Atmosphere Ocean Response Experiment (TOGA-COARE, Webster and Lukas 1992; Godfrey et al. 1998), and it capitalized on airborne Doppler radar technology. From 25 flights around the stratiform regions of MCSs,

Kingsmill and Houze (1999) compiled statistics on the characteristics of the mesoscale airflow and found that midlevel inflow consistently entered from above and descended to below the 0°C level (Fig. 14), but that this inflow entered from a wide variety of directions, as indicated schematically in Fig. 15a. The midlevel inflow in Fig. 15a is completely consistent with the squall-line configuration of Fig. 12 but the observations show a more general result because the inflow can come from any direction, and Kingsmill and Houze (1999) found the midlevel inflow direction to be determined by the large-scale environmental wind direction.

From another set of 25 flights in TOGA COARE, Kingsmill and Houze (1999) determined statistics of the ascending flow entering the convection regions of fully developed MCSs (Fig. 16). The low-level flow came from various directions relative to the MCS but in all cases it was coming from a layer which could be as deep as 4.5 km, implying that the convection was drawing in a layer of air much deeper than the boundary layer and that, as such, the mature MCS is not simply a collection of deep but weakly entraining deep convective towers arising from the boundary layer, as is often assumed in parameterization schemes. In addition, as indicated in Fig. 16, the lower-tropospheric layer of air entering the MCS's convective region usually has a negative vertical gradient of θ_e . When this layer of potentially unstable air rises and saturates, the air becomes absolutely buoyantly unstable. Doppler radar observations show that despite its absolute instability, the inflow layer maintains a coherent structure as a sloping front to rear ascent. Evidently, the release of the instability is not rapid enough or deep enough to destroy the basic layer structure. The term Moist Absolutely Unstable Layer (MAUL) has been used to describe such layers (Bryan and Fritsch 2000). Extremely high

resolution modeling reveals that within the layer of generally rising air, the buoyant elements overturn in rolls aligned along the shear (Bryan and Fritsch 2003), consistent with radar data, which show that the cells in convective zones of MCSs are often elongated, as in the early study of Ligda (Fig. 4) and later documented in the Oklahoma study of Houze et al. (1990).

The overturning in MCSs in the form of a circulation of a horizontal scale of the entire system and in the form of deep layers of air is a clear departure from the view that deep convection consists entirely of buoyant plumes arising from the planetary boundary layer. That view has traditionally dominated convective parameterization schemes. However, recent studies are finally considering how to parameterize layered MCS circulations into climate models (e.g. Moncrieff et al. 2017).

9. Wind shear in the environment in relation to the structure of MCSs

Over the summer season of GATE, it became apparent that two types of MCSs occurred within the oceanic data array. These became known as squall-line systems (Houze 1977; Zipser 1977) and non-squall events (Leary and Houze 1979b). Houze and Betts (1981) described the GATE observations of these two types of systems. The squall-line MCSs moved faster than other MCSs and featured a leading line of deep convective clouds followed by a broad region of stratiform precipitation, the latter accounting for ~40% of the rain from the MCS. The squall-line form of MCS has received much attention, largely because of its relationship to the large-scale environmental shear. The propagating line is usually somewhat curved convexly in the direction of system motion and generally normal to the lower-tropospheric wind shear, so that the relative wind is toward the line at low levels, switching to a front-to-rear direction aloft. Prior to GATE,

Moncrieff and Miller (1976) and Moncrieff (1978) had begun to show from vorticity consideration how the overturning in a convective line should be organized into deep overturning layers, as shown by the colored shading in Fig. 12. These papers indicated specifically that the overturning was related the wind shear ahead of the line. The work of Moncrieff and Miller was highly influenced by the earlier work of Browning and Ludlam (1962), who showed that the overturning in a convective storm, regardless of scale, should consist of such overturning layers. Moncrieff and Miller's insights are not restricted to tropical convection. Squall-line MCSs occur at all latitudes and over both land and ocean. The relationship to shear is always that indicated in these early papers.

After GATE, questions remained about how the convective cells embedded in the upward branch of the overturning layer of a squall-line MCS relate to the environment shear. In a highly influential paper, Rotunno et al. (1988) described how the line forming in an environment of shear like that considered in the Moncrieff and Miller papers first becomes erect. The ideas in this paper have become known as RKW theory, and Weisman (1992) and Weisman and Rotunno (2004) have published extensions and updates. The basic notion of RKW theory is that the convective cells in the line become erect in response to a balance of horizontal vorticity associated with the updrafts' positive buoyancy and the convective downdrafts' negative buoyancy (Fig. 17a-b). As the line ages, and the stratiform region grows, additional vorticity associated with the stratiform region's mesoscale downdraft leads to the convection becoming tilted rearward (Fig. 17c). When the mesoscale downdraft emanating from the stratiform region becomes extremely intense, its vorticity can push the cold pool boundary and vorticity balance illustrated in Fig. 17 far ahead of the main line in the form of a bow echo like the

example in Fig. 13.

Again, it is important to note that squall-line MCSs are a subset of all MCSs, and when the environmental shear is different from that of the Moncrieff-Miller paradigm, very different arrangements and interactions of the convective and stratiform portions of MCSs can occur. In many situations the convective cells in an MCS are not even arranged in a line (Houze et al. 1990). Bluestein and Jain (1985) and Parker and Johnson (2000) have examined cases exhibiting line structure of deep convection over the U.S. and have found a variety of patterns, all systematically related to environmental shear. Parker and Johnson (2000) found that the tropospheric shear determined whether the precipitating cloud deck of an MCS formed ahead of, along, or behind a line of convection (Figure 18). LeMone et al. (1998) examined lines of convection over the western tropical Pacific, and Johnson et al. (2005) documented lines over the South China Sea. Both studies showed several patterns of convective/stratiform organization that bore systematic but complex relationships to the environmental shear. Some lines were parallel to shear, while others were perpendicular. In an especially interesting variant of MCS structure over the tropical Indian Ocean examined by Yamada et al. (2010), the environment had low-level westerlies and upper-level easterlies, and convective and stratiform regions that separated over time by moving in opposite directions, with the stratiform regions moving systematically westward and ingesting eastward-moving deep convection.

Robe and Emanuel (2001) carried out idealized simulation of clouds in a radiative convective equilibrium setting. They provide some insight into how convective lines form in relation to both low-level and midlevel shear. For example, they find that if the

low-level shear becomes too large, lines may rotate away from the direction orthogonal to the shear. Nevertheless, Johnson et al. (2005) note that while some of forms of convective line organization are understandable via conventional reasoning, many remain puzzling and in need of further study.

Despite all of these variations of line structure and horizontal patterns of the convective and stratiform components of MCSs, similarities remain in the precipitation-forming mechanisms. When Barnes and Houze (2014, 2016) analyzed, simulated, and compared squall and non-squall MCSs over the tropical Indian Ocean, they found similar kinematic and microphysical processes to be operating in both types of MCSs.

10. Boundary layer and cold pools below MCSs

One of the results of GATE was the realization that MCSs over tropical oceans completely alter the character of the planetary boundary layer. One of the innovative instruments deployed in GATE was a shipborne vertically pointing acoustic sounder. This instrument, analogous to radar, received reflections of emitted sound waves from irregularities in the lowest 800 m of the boundary layer (Mandics and Hall 1976; Gaynor and Mandics 1978). Clear-air convective plumes, small convective clouds, and stable layers were the primary identifiable echoes. Figure 19a from Houze (1977) shows the structure of the undisturbed boundary layer ahead of the MCS. The vertically oriented spikes of high acoustic reflectivity seen emanating from the sea surface are clear-air turbulent plumes. The echo spikes with bases at about 300 m appearing intermittently above the layer containing the surface-based plumes are low-level small non-precipitating cumulus clouds. During the rainy period no signal was obtained. Figure 19b shows the boundary-layer structure in the wake of the MCS. The stable layer capping the cool

604 downdraft air that had entered the boundary layer was marked by the continuous, dark,
605 undulating streak of echo between 200 and 400 m. As this cold air moved over the warm
606 ocean, turbulent mixing was stronger than that ahead of the MCS, and the pattern of
607 plume echoes seen immediately after 1820 was, therefore, denser than before the squall.
608 Soundings obtained in the cold pool were consistent with the acoustic sounder data, and
609 calculations of surface fluxes corroborated the increased turbulence. Bulk flux
610 calculations for MCS wake conditions over tropical oceans readily show that latent heat
611 fluxes dominate in cold pools over tropical oceans. Using soundings and flux
612 measurements obtained aboard the GATE ships, Johnson and Nicholls (1983) mapped the
613 boundary-layer structure throughout the cold-pool region of one of the tropical oceanic
614 MCSs observed in GATE. Figure 20 shows the mapped latent heat flux. Tethered balloon
615 data take from GATE ships during convective system passages were consistent with the
616 surface fluxes (Barnes and Garstang 1982).

617 Cold pools from MCSs, as seen in GATE, have a major impact on the boundary
618 layer in ocean regions where MCSs are frequent. Using the GATE acoustic sounder
619 measurements, Gaynor and Mandics (1978) estimated that 30% of the area over the ocean
620 in the eastern Atlantic Intertropical Convergence Zone is occupied by cold pools at any
621 given time, and because of their large areal coverage, MCSs account for a large
622 proportion of that cold pool area. Over the western tropical Pacific, shipboard
623 measurements of fluxes analyzed by Young et al. (1995) and Saxen and Rutledge (1998)
624 showed increased fluxes similar to those seen in GATE during the passage and in the
625 wakes of MCSs. The latent flux increases were mainly due to increased wind speed while
626 the sensible heat fluxes were also affected by both wind and the temperature drop in the

627 cold pools. Examining fluxes measured on moorings over the tropical Pacific, Esbensen
628 and McPhaden (1996) found that the cold pools of MCSs enhanced surface fluxes by 10–
629 30%, depending on wind conditions.

630 One of the important effects of convective downdraft cold pools is that they act as
631 density currents and trigger new convection at their leading edges. In this sense, the cold
632 pools can be thought of as a medium of communication between existing and future
633 convection. Capturing cold pool formation and interaction is thus a key to representing
634 accurately whole populations of convection. In the Dynamics of the MJO (DYNAMO)
635 field experiment over the central tropical Indian Ocean, Feng et al. (2015) have shown
636 the characteristics of a population of cold pools filling the boundary layer and colliding
637 with each other and triggering new convection. They found that the strongest new
638 convection occurs at points of collision. Starting with a population of only a few small
639 precipitating clouds, ever-larger new clouds form, with some eventually reaching MCS
640 proportions with very large cold pools (Rowe and Houze 2015).

641 The flux characteristics, formation, growth, and dissipation of MCS cold pools
642 over land has been harder to determine because of the wide variety of land surface
643 characteristics, and much remains to be determined. Unlike over oceans, where the latent
644 heat flux dominates, over land, the surface sensible heat flux is important in modifying
645 the cold pools. Modeling has produced a few insights, but no holistic understanding of
646 MCS cold pools over land. Grant and van den Heever (2016) have performed model
647 calculations indicating that both surface sensible heat flux and turbulent entrainment of
648 environment air is important in MCS cold pool dissipation over land. Model results of
649 Gentine et al. (2016) indicate that surface roughness is an important factor to consider in

cold pool evolution, but their calculations do not necessarily apply to MCS cold pools. Simulations by Drager and van den Heever (2017) have shown that passage of MCS cold pools over wet patches of ground mitigates the cold pool dissipation.

11. Elevated MCSs

In midlatitude continental regions, boundary layer factors sometimes cease to be a consideration, as the layered overturning of an MCS is sometimes completely aloft, disconnected from the planetary boundary layer. Trier and Parsons (1993) noted this behavior in an MCC documented in PRE-STORM. Marsham et al. (2010) developed a conceptual model of such an “elevated MCS” (Figure 21). In this model, the MCS overrides a layer of stable air, located on the cold side of a front or over a stable layer produced by nocturnal cooling of a land surface. Warm unstable air running over the top of the stable layer provides the layered ascent of the mature MCS lying atop the stable layer. The mesoscale downdraft emanating from midlevels lies under the mesoscale ascent, as in Fig. 12, but may remain atop the stable layer, as depicted in Fig. 21. Also indicated in the conceptual model is a perturbation of the stable layer underneath the MCS. The updraft layer of the MCS rises over this bump in the stable layer, which is thought to be due to a gravity wave or bore propagating through the stable layer. In a model simulation, Parker (2008) obtained an MCS initiated before nocturnal cooling produced the stable layer, and the MCS subsequently became elevated above the stable layer. The perturbations of a stable layer, associated with a front or strong nocturnal cooling over land may also trigger new convection that may evolve into an elevated MCS (e.g., Parker 2008), and the downdrafts of a developing MCS triggered aloft may penetrate to the surface and change an elevated MCS into an ordinary MCS (Marsham et

al. 2011). These various scenarios led to a field project in 2015 called the Plains Elevated Convection at Night (PECAN), which aimed to study elevated deep convection in its various forms, with an emphasis on nighttime conditions when stable layers are most pronounced (Geerts et al. 2017).

The occurrence of elevated MCSs in connection with fronts is sufficiently common in midlatitude continental regimes that it led Fritsch and Forbes (2001) to identify them as a special category of MCS, which they called “Type-1,” in contrast to “Type-2” MCSs that occur in more barotropic environments, such as the tropics. Wilson and Roberts (2006) estimated that about half of the MCSs over the central U.S. are elevated. Other regions of the world such as the Pampas of Argentina, which has MCSs similar to those in the central U.S. (Rasmussen and Houze 2011), likely have similar occurrences of elevated MCSs. The existence of elevated MCSs poses a complex forecast problem and a particular complication for projecting climate change. Global models that are used to project long-term changes in the general circulation and climate still depend on parameterization of deep convection, and existing parameterization schemes assume deep convection to be rooted in the planetary boundary layer. The existence of elevated mature MCSs like that illustrated in Fig. 21 and the further complication of deep convection being initiated above stable layers implies the inability of existing parameterizations to represent elevated forms of deep convection. Clearly radical changes in parameterization methods will be needed for global models used for climate projections over long time periods, or advances in computing power will need to allow such models to run with sufficiently fine grid spacing for MCSs to be fully resolved as they are in “cloud resolving models,” which are presently being run regionally over

modestly long climatic periods (e.g. Prein et al. 2015, Yang et al. 2017, and others).

12. Canonical lifecycle of an MCS, modes of heating, and potential vorticity generation

One of the significant results of GATE was the recognition that MCSs undergo a canonical lifecycle, which is illustrated schematically in Fig. 22. This lifecycle has been confirmed and elaborated by subsequent field campaigns using satellite imagery and radars on ships, aircraft, islands and continental landmasses. Fig. 22a shows a single “cell” of deep convection at this early stage, but in reality a group of such “cells” occurring in close proximity is an essential feature for the system to reach mesoscale proportions. The group is often in the form of a line, but the cells may be grouped in other patterns. Note in Fig. 8 how the precipitation is almost completely convective in the early hours of the MCS. Over time, the initial cells weaken while new ones form. As the older cells weaken, their vertical motions no longer support vertical advection of hydrometeors, and the precipitating ice particles aloft generally drift downward through the environment of weakened mean upward motion (Fig. 22b). They melt as they fall below the 0°C level and turn into rain. The raining layer is dominated by downward motion where the drops are falling through a layer of subsaturated air and evaporating at rates calculated by Brown (1979). This period when both active convection and stratiform precipitation are present characterizes the mature stage of an MCS. In Fig. 8 the mature stage would be the period ~1700-2300 GMT. After the mature stage, the convective and stratiform portions of the MCS decline together. Figures 22c and d illustrate this period of decline, which can last well beyond the time when precipitation is reaching Earth’s surface. The remaining ice cloud aloft can last 10s of hours. An

important detail in Fig. 22 is that the cloud top continues to rise. Turbulent motions in the stratiform cloud layer cause the top to rise as the cloud layer entrains dry air from above. The mixing is strongest at night when radiative cooling in the upper portion of the cloud layer is actively destabilizing the layer. The high thin cloud layer left aloft remains for long periods of time and has climatological implications for radiative transfer in the upper troposphere (Hartmann 2016).

The heating of the atmosphere by an MCS over its lifetime is a combination of latent and radiative heating. The latent and radiative heating are interrelated through the water budget of the MCS. The fallout of rainwater from the MCS is proportional to the net latent heat released into the atmosphere by the MCS, and the ice cloud aloft affects radiative transfer in the upper troposphere. Furthermore, the MCS dynamics and physics affect the distribution of both latent and radiative heating by the system. GATE and subsequent field studies led Houze (1982) to realize that the distribution of the latent heating with height differs between the convective and stratiform portions of an MCS. The latent heating is related to the water budget of the MCS. Figure 23 illustrates the bulk water budget over the area and lifetime of an MCS. The bulk components of the water budget are shown as if they were combined into one convective region, one associated stratiform region, and anvil clouds associated with both the convective and stratiform regions. The variables represent the disposition of condensed water mass, which is interrelated by water budget equations (Houze et al. 1980). The net heating in the convective region is dominated by condensation and sublimation of vapor in the convective updrafts (C_{cu}), which must account for all of the convective rain (R_c), as well as the condensate transferred from the convective to the stratiform region (C_T) and rain

742 re-evaporated in the convective downdraft (E_{cd}). Freezing of supercooled water in the
 743 convective updrafts further increases heating in the upper part of the convective cells
 744 (Zipser 2003). The heating profile in the convective region tends to be positive
 745 throughout the troposphere (Fig. 24). Sometimes the maximum is somewhat below the
 746 middle troposphere and sometimes somewhat higher, but generally not in the upper
 747 troposphere (Houze 1989). The net heating and cooling processes in the stratiform region
 748 are dominated aloft by sublimation of vapor onto ice particles (C_{su}) in the weakly buoyant
 749 stratiform updraft region at upper levels and evaporation (E_{sd}) of stratiform precipitation
 750 particles in the mesoscale downdraft occupying the lower part of the troposphere. In
 751 addition to the evaporation, cooling by melting occurs in a ~ 0.5 km thick layer just below
 752 the 0°C level in the stratiform zone (Leary and Houze 1979a). As a result of these
 753 processes, the latent heating profile is positive at upper levels and negative at lower levels
 754 (Fig. 24a). In addition to the latent heating and cooling processes, convergence of
 755 radiative fluxes in the long-lived and extensive ice cloud “anvils” of the MCS leads to
 756 further heating aloft, making the net heating profile “top-heavy.” As shown in Fig. 24b,
 757 the larger the stratiform proportion of MCS rainfall, the more the overall profile is “top-
 758 heavy,” meaning that heating is more concentrated aloft than would be the case if all the
 759 precipitation were occurring in convective towers. The top-heaviness of heating profiles
 760 associated with MCSs was first pointed out by Houze (1982), and Hartmann et al. (1984)
 761 showed that the general circulation of the tropics was substantially impacted by the top-
 762 heaviness of MCS heating. Schumacher and Houze (2003) used data from the TRMM
 763 satellite radar to show that across the tropics rainfall from convection ranges between
 764 ~ 20 -70% stratiform, implying that as a result of MCSs heating profiles across the tropics

tend to be top-heavy. The stratiform rain fraction tends to be higher over the oceans (see Section 17). Similar stratiform proportions occur in connection with midlatitude MCSs.

The top-heaviness of the heating profile of an MCS has important dynamical implications. The quasi-balanced flow of the large-scale circulation of the atmosphere is in a constant state of adjustment to the potential vorticity field. When diabatic processes occur, potential vorticity is not conserved, but rather it is affected by spatial gradients of the heating. In particular, the time rate of change of potential vorticity is directly proportional to the vertical gradient of heating. Thus, as can be inferred from Fig. 24, the more top-heavy the MCS heating profile, the stronger the feedback to the large-scale. This feedback is sometimes felt locally in the form of a “mesoscale convective vortex (MCV),” identified and named by Bartels and Maddox (1991) and illustrated schematically in Fig. 25 for a case analyzed by Fritsch et al. (1994). Raymond and Jiang (1990) pointed out how the environment shear profile typical of certain MCSs can interact with the potential vorticity anomaly associated with the heating profile to prolong the lifetime of the MCS. The importance of the MCV associated with the potential vorticity anomaly to the duration and precipitation productivity of MCSs has been shown by Trier and Davis (2002) and Schumacher et al. (2013). On seasonal time and space scales, it has been found that more realistic structures of the mean Walker Cell (Hartmann et al. 1984) and El Niño-Southern Oscillation (Schumacher et al. 2004) are obtained in simple climate models if it is assumed that the net effect of convection on the large scale circulation has an imprint of the top-heavy stratiform heating of MCSs.

13. Microphysics and the connection between convective and stratiform precipitation of MCSs

Although GATE was massive in terms of the number of observational platforms brought to bear on tropical convection, instrumentation on those platforms as well as numerical modeling were limited in sophistication at that point in time, and analysts had to rely on inference more than is the case today. The only particle-scale microphysical measurements obtained were raindrop-size distributions detected by ship-based electromechanical disdrometers (Austin and Geotis 1979) and by a foil impactor on a DC-6 aircraft (Cunning and Sax 1977). Nevertheless, on the basis of this primitive information on drop-size distributions measured below the melting layer of the stratiform regions of several MCSs observed in GATE, Leary and Houze (1979a) inferred aspects of the ice-phase microphysics of the stratiform regions of MCSs in GATE. Combining the drop-size measurements with quantitative radar reflectivity patterns and sounding data, they concluded that the ice particles just above the melting layer in the stratiform region were most likely rimed aggregates and/or graupel (Fig. 26). These inferences have been generally confirmed as technology has improved (Barnes and Houze 2014).

Just 5 years after GATE, another field campaign of the Global Atmospheric Research Programme was the Monsoon Experiment (MONEX), in which MCSs similar to those seen in GATE were investigated over the Bay of Bengal by some of the first flights of an instrumented National Oceanic and Atmospheric Administration P-3D aircraft, on which some of the earliest optical Particle Measuring System probes were mounted to obtain shadow-gram images of ice particles through a wide range of altitudes (i.e. temperatures) in the deep layer of stratiform cloud above the melting layer. Most

particle images were of indiscernible shape but occasionally displayed certain crystal habits and structures. Houze and Churchill (1987) determined the frequency of occurrence of identifiable ice particle types seen at different flight level temperatures (Fig. 27). They found that the frequencies were consistent with ice particles growing by vapor deposition in a cloud in which vertical air motions were weak enough for the particles to be growing by vapor deposition and aggregating while drifting downward toward the melting level. The most frequent particle types at each level were consistent with the pattern of ice particle growth, development and fallout in a stratiform cloud of weak upward air motion. In addition, the occurrence of nearly round particles just above the 0°C level would be consistent with riming producing some graupel or other rimed particles just above the melting layer as suggested by Fig. 26.

In the 1980s and 90s, intensive observations of MCSs turned to midlatitudes, and Doppler radar was beginning to be used in research on precipitation cloud systems. The 1985 PRESTORM program was one of the first large field campaigns to employ multiple research Doppler radars. Working with data from an MCS over Kansas, Braun and Houze (1995) retrieved the patterns of freezing and melting in the MCS by applying thermodynamic and bulk microphysical equations to the air motion field synthesized from dual-Doppler radar observations. They found that the melting in the stratiform region was occurring in a thin horizontal layer ~200 km in breadth, while the freezing was in a vertical column corresponding to the intense active convective cells of the MCS (Fig. 28). As noted in Section 12, the cooling due to melting in middle to low levels of the stratiform region and the column of freezing in the upper parts of convective cells both contribute to the top-heavy heating profile of the MCS.

In 2011-12 another technological advance led to a significant increase in the observational knowledge of the microphysical processes in the convective and stratiform regions of MCSs, when dual-polarization S-band Doppler radar measurements were made during a nearly 4-month period over the equatorial Indian Ocean to study the Madden-Julian Oscillation, or MJO (Madden and Julian 1971, 1972), in the Dynamics of the MJO (DYNAMO) field experiment (Yoneyama et al. 2013).⁵ MCSs similar to those seen in GATE and MONEX occur in that tropical oceanic region, and Barnes and Houze (2014) analyzed statistics of the radial velocity and microphysical particle type retrieved from the dual-polarization radar data. They composited the patterns of identified particle types in relation to the mesoscale flow patterns in both convective and stratiform regions for data obtained in two DYNAMO MCSs. The composite for the stratiform region (Figure 29a) validated the general pattern of microphysics in the stratiform region that had been suggested in GATE and MONEX (cf., Figs. 26 and 27) in that rimed particles occurred intermittently just above the melting layer at the base of a deep layer of snow. Barnes and Houze (2016) carried this work further by assimilating Doppler radar velocities observed in two DYNAMO MCSs into a cloud-resolving model and found that the same general layering of microphysical processes was found in both observations and simulations, although the model results varied substantially with the particular choice of microphysics parameterization. The model results generally showed deposition everywhere above the 0°C level, aggregation at and above the 0°C level, melting at and below the 0°C level, and intermittent riming near the 0°C level, consistent with the GATE and MONEX results seen in Figs. 26 and 27.

⁵ DYNAMO was part of a multi-agency program, and was also known by the abbreviations CINDY and AMIE, which are explained by Yoneyama et al. (2013).

854 Barnes and Houze (2014, 2016) also composited the dual-polarization radar
855 results for convective regions of DYNAMO MCSs (Fig. 29b). The convective regions
856 had a vertical column of graupel, extending a few km above the 0°C level and signifying
857 the occurrence of freezing associated with riming in that zone—consistent with Fig. 28.
858 The composite also shows wet (i.e. melting) aggregates just below the 0°C level,
859 consistent with the pattern of melting below the column of freezing in Fig. 28. A
860 composite similar to those in Fig. 29 has not yet been compiled for midlatitude MCSs,
861 but based on the fact that continental convective updrafts are generally stronger than
862 those over oceans (Zipser and LeMone 1980; LeMone and Zipser 1980), we would
863 expect the signal of graupel (and hail) and associated freezing to be stronger over
864 continental regions, except perhaps for the Amazon environment, which has some
865 maritime-like characteristics.

866 It has also been apparent since the time of GATE that the microphysical processes
867 in the convective and stratiform regions of an MCS are not independent but instead
868 closely related. The term C_T in Fig. 23 represents condensate generated in the convective
869 region that later becomes part of the precipitating stratiform cloud. That figure, however,
870 does not indicate the process by which condensate initiated in convective cells becomes
871 part of the stratiform region. Results from GATE hinted at the process when Leary and
872 Houze (1979b) described the evolution of radar echo structure in an MCS. They found
873 that over time an area of active convective echoes became stratiform while new
874 convective elements formed in the near vicinity of the now-stratiform cells. As this
875 process repeated, an increasingly large portion of the MCS radar echo became stratiform.
876 Figure 30 illustrates the process schematically. Note how ice particle 1 generated in an

active cell eventually falls out and melts in the stratiform region partially formed by its parent cell. From this schematic it is evident that the term C_T in Fig. 23 does not represent a spatial advection so much as a temporal transformation that occurs while the ice particles are falling out. It is also important to note that the buoyant elements rising in active convective towers expand as they rise, as illustrated schematically in Fig. 31. Yuter and Houze (1995) dubbed these elements “particle fountains.” Ice hydrometeors suspended in the updraft therefore fall out over an ever-expanding area as the buoyant element approaches its maximum level of ascent. As more and more such widened elements arrive aloft and convective towers weaken, a mesoscale region of the mid-to-upper troposphere becomes filled with old, weakly buoyant elements depositing ice particles. Thus, a stratiform region forms from the remains of old and weakened but still buoyant convective elements.

Note that the process in Fig. 30 requires no shear of the environment. Thus, in principle, an MCS with convective and stratiform regions can develop without shear. However, MCSs often form in the presence of environmental shear, which affects the arrangement of the convective and stratiform components of the MCS. A common and oft-studied form of MCS is the leading-line/trailing-stratiform squall line (Fig. 12), in which the particle fountain elements are systematically advected rearward in the front-to-rear ascending flow. Thus, the widened elements of weak buoyancy and slowly falling ice particles are systematically moved rearward of the leading line where they form the upper part of the stratiform region, as illustrated in Fig. 32.

14. Modeling of individual MCSs

At the time of GATE, “cloud resolving” or “convection permitting” models of the

type that are ubiquitous today were in their infancy, so only very primitive numerical modeling of convective clouds was part of GATE research (e.g. Simpson and van Helvoirt 1980). Rapid development of models in the 1980s-1990s, however, quickly led to many simulations of MCSs. Almost all such modeling has been focused on the type of MCS that has a leading line and trailing stratiform region (e.g. Fovell and Ogura 1988; Lafore and Moncrieff 1989; Skamarock et al. 1994). One exception is the study of Barnes and Houze (2016). They successfully ran and compared simulations of both squall and non-squall MCSs observed in DYNAMO in which the air motions in convective and stratiform regions were assimilated into the model. Results were consistent with the observational results of Kingsmill and Houze (1999) (Figs. 14 and 16) in that the of the non-squall MCS had a midlevel downward-sloping flow below the melting level in the stratiform region, and a deep upward sloping flow in the convective region.

An example of a simulation of an MCS of the leading-line/trailing-stratiform type, provided to the author by Professor Robert Fovell, is shown in Fig. 33. It exhibits the main features of the MCS: A vertical tower of high reflectivity is centered in the convection zone at ~340 km on the horizontal axis; behind the convective zone is the stratiform region with a bright band at the melting level. The negative values of the horizontal wind field show the front-to-rear flow ascending from low levels ahead of the storm, through the convective zone, and into the stratiform region. The consistency with the conceptual model in Fig. 12 is evident. The positive values show the midlevel flow entering from the rear of the storm, descending to the surface below the melting level, and feeding into the bulbous head of the gust front in the convective zone. The exact characteristics of the model's radar reflectivity field depend on the microphysical

parameterization and radar simulator used in the calculations. Effort directed toward determining the most accurate representation of the cloud and precipitation fields of MCSs continues to be an active area of research (e.g. Morrison et al. 2009; Powell et al. 2012; Van Weverberg et al. 2013, Barnes and Houze 2016; and others).

The ability to simulate MCSs accurately is important diagnostically because some features of MCSs are almost impossible to observe. For example, one of the most important properties of an MCS is its thermal structure. Yet the thermal properties of an MCS are nearly impossible to measure directly, especially in stratiform regions where temperature perturbations are small but important. Temperature may be retrieved from dual-Doppler radar observations, but such measurements are seldom available and that methodology is fraught with assumptions and far from precise. Accurate model simulations are at present the best indicators of MCS thermal structure. The simulation in Fig. 33 shows how the upper portion of the stratiform region is filled with weakly buoyant air previously located in the convection zone, as suggested by Figs. 31 and 32.

15. Dynamical interpretations

The ability to model MCSs fairly accurately should help to understand MCSs on a theoretical basis. Nevertheless, such basic questions as, “Why do MCSs exist?” or “Why do they have a preferred scale of a few hundred kilometers?” have yet to be fully answered. A reason for this state of poor understanding is that MCSs do not lend themselves easily to fluid dynamical explanations as do, for example, baroclinic instability waves. At horizontal scales of a few hundred kilometers, MCSs occur near the boundary between two-dimensional balanced flow and three-dimensional turbulent flow (Tulloch and Smith 2006). Theoretically accounting for MCSs in this scale range is

946 especially complicated because MCSs are not simply fluid dynamical entities.
947 Thermodynamics, cloud microphysics, turbulence, and probably radiative transfer
948 complicate theoretical explanations accounting for the existence of MCSs in the
949 atmosphere. For example, the horizontal scale of an MCS is affected strongly by ice-
950 phase cloud microphysics. Because an MCS is composed of the aggregate of convective
951 and stratiform regions of the system, its horizontal area coverage depends in part on the
952 fallout trajectories of ice particles (Fig. 30). The maximum horizontal extent reached by
953 an MCS also depends on it having a continual supply of moisture to supply the formation
954 of new convective elements while old ones weaken and become part of the stratiform
955 region of the MCS, and a host of topographic and radiative factors impact the moisture
956 supply.

957 Despite these obstacles, progress has been made on some aspects of dynamical
958 understanding of MCSs. More or less contemporaneously with GATE, but not as part of
959 the GATE scientific activity, Moncrieff and Miller (1976) had been considering the
960 question of what factors determine the mesoscale circulation of an MCS, which they
961 realized needed to be viewed as layered overturning rather than as convection in the form
962 of bubbles of buoyant air rising from the boundary layer. A series of papers by Moncrieff
963 and colleagues (Moncrieff and Miller 1976, Moncrieff 1978, 1981, Thorpe et al. 1982,
964 Crook and Moncrieff 1988, Moncrieff 1992; Moncrieff et al. 2017; see also pp. 497-505
965 of Cotton and Anthes 1989 and Chapter 9 of Houze 2014 for synopses of the Moncrieff
966 work) have quantified this view, for the case of an idealized steady-state two-dimensional
967 MCS. A key assumption of the theory is that the storm may be characterized by a
968 prescribed decrease in hydrostatic pressure across the updraft at midlevels. If the large-

scale environment is sheared as well as unstably stratified, air must flow through the storm along a unique set of streamlines. The geometry of the streamlines is deduced from conservation of entropy, mass, momentum, and vorticity along streamlines. Similar reasoning is employed to determine the streamlines of the downdraft fed by midlevel inflow on the rear side of the storm. For a typical environment of strong low-level shear, the updraft consists of a layer ascending on a slantwise path through the storm (Fig. 34). The mathematical basis of the Moncrieff theory is summarized in Chapter 9 of Houze (2014). Models of MCSs generally confirm the layered overturning model of Moncrieff (Fig. 35).

Because the Moncrieff theory is for steady-state conditions and prescribed environmental shear and stability, it provides no explanation for why the layer inflow and ascent initially develops in an MCS. Explanations that have been offered usually involve gravity-wave thinking. Such thinking traces all the way back to Hamilton and Archbold's (1945) groundbreaking study (Section 2). They posited that the disturbance lines of West Africa were an atmospheric manifestation of a gravity wave. In their words, the "disturbance line" behaved as a traveling disturbance, which "shifts the various air particles while they are under its influence, somewhat after the manner of a wave at sea." Regarding factors controlling the movement of the system, they noted: "It is tempting to argue that the disturbance line must be carried along in the prevailing upper wind current, where the bulk of its cloud is." They "very tentatively" compared the motion of the disturbance line to that of a simple gravity wave in a channel of stratified liquid and concluded that the motion of such a wave was not unlike that of the observed disturbance lines. Several decades passed before gravity-wave thinking re-emerged in relation to

MCS dynamics. In the 1970s and 1980s, a flurry of papers were inspired by Charney and Eliassen (1964)'s concept of "conditional instability of the second kind" (CISK) to describe an hypothesized cooperative interaction in which friction-layer convergence drives deep convection and associated heating to strengthen and/or perpetuate a warm core cyclone. Applying this idea to gravity waves, Lindzen (1974) dubbed the interaction of wave dynamics and convective clouds "Wave-CISK" on the basis that the convergence and upward motion in the circulation of an inviscid mesoscale or larger scale wave (rather than frictional convergence) can maintain a deep convective heat source, which in turn strengthens or maintains the wave. Raymond (1984) applied Wave-CISK to try to study the scale of wave response to an imposed heating profile. The resulting dispersion relation did not definitively determine a wave response that would correspond to MCSs because the results are sensitive to the details of the parameterization of the heating. However, the calculations show that under environment wind and thermodynamic stratification characteristic of squall lines, a dominant growing mode appears that has the layer lifting structure seen in real squall-line MCSs (Fig. 36).

Schmidt and Cotton (1990) and Pandya and Durran (1996) took another analytic approach based on gravity wave thinking to explain the layer overturning in an MCS. These studies ran nonlinear high-resolution models to simulate the realistic detailed behavior of an MCS consisting of a squall line and trailing stratiform region. Pandya and Durran (1996) averaged the diabatic heating field in the region of the simulated convective line over a two-hour period (Fig. 37a). Then they input the averaged heating field into the model and let an initially undisturbed field respond. The result yielded the horizontal wind field shown in Fig. 37b. They noted that this field is consistent with

Moncrieff's steady-state model, and they showed theoretically that the layered overturning in Fig. 37b is a gravity wave response to the mean heating in the convective line. This result indicates that the deep-layer inflow occurs after the convective cells have "organized," i.e., clustered into one mesoscale group that constitutes a quasi-steady heat source. Layers of inflow are then drawn up and down through the system as a gravity wave response to the heating, as in Fig. 34. Thus, it is seen that the mesoscale circulation is not the sum of convective elements but rather develops over time as a broader circulation induced by the aggregated heating in the region of grouped convection.

16. Relationship of MCSs to synoptic-scale and larger circulations

a. MCSs in equatorial waves and the MJO

In the years just preceding GATE, visible and infrared satellite imagery were showing that high clouds atop deep convection in the tropics bore a systematic relationship to synoptic-scale easterly waves in the tropics, over both the Atlantic and Pacific (Frank 1970; Chang 1970; Chang et al. 1970; Reed and Recker 1971). GATE showed the relationship of MCSs ("cloud clusters" in the terminology of that time) in relation to tropical easterly waves of African origin (Payne and McGarry 1977). Studies of satellite imagery of tropical convection continued into the 1980s. Nakazawa (1988) found that deep convective clouds seen in satellite imagery over the tropical western Pacific moved with propagating motion systems on multiple scales in a kind of interference pattern. These early satellite studies did not, however, by themselves isolate the specific importance of MCSs in the relationship of convection to the larger-scale envelopes of waves in other tropical disturbances.

1037 In the tropics, the upward air motion and associated heating processes in synoptic-
1038 and larger-scale waves are highly concentrated within MCSs and other convective clouds.
1039 After the GATE and MONEX campaigns in the tropics, it was becoming apparent that
1040 the heating processes occurring in MCSs are top-heavy as a result of cooling by melting
1041 and evaporation at lower levels (Section 12), and that this top-heavy heating was
1042 significantly influencing the feedback of MCSs to the larger-scale circulation (Houze
1043 1982, 1989; Hartmann et al. 1984; Schumacher et al. 2004). Within an equatorial wave or
1044 other large-scale circulation feature numerous convective cloud entities occur. The
1045 vertical profile of net vertical air motion (and hence heating) in a sector of a near-
1046 equatorial large-scale wave thus depends on the mix of clouds of different types and sizes
1047 in that sector. For this reason, Mapes et al. (2006) pointed out that a tropical large-scale
1048 wave (synoptic and larger) must have a varying population of convective clouds
1049 (“building blocks”) to account for the net upward air motion in any wave sector. They
1050 hypothesized that the population varies across a wave as indicated in Fig. 38. Panel a
1051 shows building blocks that are mostly shallow convection. That population is followed in
1052 panel b by an ensemble of building blocks of different sizes and depths, with some of the
1053 blocks being MCSs with large stratiform components. A third population in panel c is
1054 dominated by a large proportion of MCSs. Observations in DYNAMO were consistent
1055 with this building-block idea as synoptic-scale waves repeatedly passed over the
1056 observational network during the four months of the campaign. Fig. 39a from Zuluaga
1057 and Houze (2013) composites reanalysis data temporally centered on the maximum
1058 precipitation of a DYNAMO wave event. The large-scale convergence was initially
1059 concentrated in low levels, extended through a deep layer near and just after the time of

maximum rainfall, and finally was finally concentrated aloft with divergence in lower levels. Fig. 39b shows that the precipitating cloud population (determined by radar) exhibited a behavior consistent with the building-block hypothesis. Four categories of radar echoes were compiled statistically relative to the wave. In the early stages of the wave passage, shallow isolated convective cells reached their maximum occurrence. As time went on and the convective population changed, the maximum frequency of deep but relatively isolated convective cells occurred. Just before the peak rainfall, the convective elements that were both deep and wide achieved maximum frequency of occurrence, indicating that the convection elements were systematically increasing in scale to mesoscale proportions. These elements were indicating the presence of developing MCSs. Finally in the later time period echoes with broad stratiform precipitation areas dominated, indicative of a large presence of mature MCSs. The cloud population was thus in sync with the large-scale vertical motion in Figure 41a—39a. The shallow cloud population in the early stages corresponding to the convergence and upward motions was concentrated in the low troposphere. The population in middle stages was dominated by deep and wide convective elements of growing MCSs, when the upward motion extended through a deep layer. The late stages had a predominance of broad stratiform echo consistent with the wave's upward motion aloft and downward motion in the lower troposphere.

Thus, the vertical motion in the composite synoptic-scale wave, like that in a gravity wave (Fig. 36), is similar to the layer lifting in an individual MCS but on a larger scale. Such similarity also occurs on scales larger than gravity waves and synoptic-scale waves, namely the MJO (Moncrieff 2004; Moncrieff et al. 2017). Figure 40 shows

layered overturning in the MJO, qualitatively similar to an MCS but on a scale of thousands of kilometers. The vertical motions of the MJO are again related to the populations of convective clouds in the different sectors of the large-scale disturbance. Barnes and Houze (2013) used a statistical analysis of TRMM Precipitation Radar (PR) data to show how the precipitating cloud population changes with the stage of the MJO. Figure 41 compares the frequencies of occurrence of two types of radar echoes seen by the TRMM PR over the central Indian Ocean. An isolated shallow echo is an echo covering no more than two 5 km x 5 km pixels and having tops at least 1 km below the 0°C level and may be thought of as showers of “warm rain.” A broad stratiform region is a contiguous stratiform echo covering $>30,000 \text{ km}^2$. From Figure 41, it can be seen that the active phase of the MJO has maximum frequency of precipitation from the broad stratiform regions of MCSs and the lowest frequency of shallow isolated precipitating convective clouds. Using the TRMM latent heating product, Barnes and Houze (2016) determined the latent heating profiles associated with different categories of radar echoes as a function of MJO phase. Barnes et al. (2015) found that the top-heavy heating profile associated with the stratiform regions of MCSs was by far the most pronounced during the active phase of the MJO (Fig. 42). Virts and Houze (2015) showed that during the MJO active periods lightning decreases as MCSs increasingly dominate the cloud population.

b. MCSs in relation to midlatitude baroclinic waves and low-level jets Since the 1980s, tropical studies of MCSs in relation to the larger-scale circulation were proceeding largely in parallel with similar studies of midlatitude MCSs. Studying continental convection over the central United States, Maddox (1983) published a landmark

1106 observational paper showing that MCCs occurred systematically ahead of synoptic-scale
1107 baroclinic troughs in the westerlies (Fig. 43). Yang et al. (2017) have confirmed this
1108 behavior in a modeling study verified by satellite and radar data. They showed that MCSs
1109 in free-running cloud-resolving simulations over two warm seasons systematically
1110 formed ahead of large-scale troughs in the westerlies (Fig. 44). Trier and Parsons (1993)
1111 noted how a trough moving over the Rocky Mountains and into the Great Plains area
1112 strengthens the climatological southerly low-level jet that feeds moisture into MCSs
1113 forming over the central U.S. (Figure 45). A similar behavior occurs in South America,
1114 where the South American Low-level Jet (SALLJ) flows southward along the eastern
1115 edge of the Andes from the moist Amazon region to feed MCSs in the region centered on
1116 Argentina (Nogués-Paegle and Mo 1997; Douglas et al. 1998; Saulo et al. 2000; Marengo
1117 et al. 2004; Vera et al. 2006; Salio et al. 2007; Rasmussen and Houze 2016). As shown
1118 by Bonner (1968), these low-level jets are stronger at night, which gives nocturnal
1119 preference for MCSs over the central U.S. (as noted by Huckleberry Finn, see
1120 Introduction). Dai et al. (1999) showed how the diurnal and semidiurnal processes favor
1121 large-scale convergence over the Rockies during the day and over the plains to the east at
1122 night. These processes assure that the enhanced jet associated with an approaching trough
1123 has its maximum effect on MCSs at night in the central U.S. Data from the U.S. radar
1124 network show that MCSs developing from diurnally triggered convection over the
1125 Rockies and propagating eastward maximize at night over the central U.S. (Carbone et al.
1126 2002), in conjunction with the nocturnal maximum of the low-level jet in that region.
1127 Feng et al. (2016) found that an increase in MCS activity over the central U.S. over the
1128 last has been accompanied by strengthening of the low-level jet and its moisture transport

1129 over the past 30-40 years.

1130 In an early modeling study, Stensrud (1996) found that the persistent occurrence
1131 of parameterized deep convection feeds back in a way that strengthens the parent trough.
1132 The modeling study of Yang et al. (2017) suggests that the mechanism by which the
1133 longer-lived MCSs strengthen the trough is the top-heavy heating profile resulting from
1134 the large stratiform regions of long-lived MCSs (Section 12).

1135 **17. Global distribution of mesoscale convection: Variability, and impacts on the** 1136 **global circulation**

1137 At the time of GATE, in the mid 1970s, satellite meteorology was still in its
1138 infancy. For the next 2-3 decades, satellite studies relative to MCSs (e.g., Laing and
1139 Fritsch 1997) were limited to two-dimensional visible and infrared imagery. Fritsch et al.
1140 (1986) used these images to show that MCCs, as identified by Maddox (1980, see Section
1141 5) account for about 50% of warm season rainfall over the central U.S. The first use of
1142 passive microwave sensors on satellites in the 1980s added the ability to detect
1143 precipitation from space (Wilheit 1986). By combining infrared brightness and passive
1144 microwave data from different satellites, Yuan and Houze (2010) used infrared and
1145 passive microwave imagery to define and identify a mature MCS by multi-sensor satellite
1146 analysis. They used the fact that an MCS has both a large cold cloud shield (minimum
1147 brightness temperature < 220) and a large rain area ($> 2000 \text{ km}^2$ in passive microwave
1148 rain detection), with the additional criterion that the large cold cloud shield contain some
1149 intense rain, also identified by passive microwave data. MCSs so defined accounted for
1150 56% of tropical rainfall over the years of the Yuan and Houze study. This objective
1151 method is consistent with others, who have, on the one hand, used large cold cloud

shields to distinguish MCSs (e.g., Maddox 1980) or, on the other hand considered an MCS to be a cumulonimbus system that has become large enough to support a contiguous rain area of mesoscale dimension (e.g., Mohr and Zipser 1996; Nesbitt et al. 2000). Using this objective method, Yuan and Houze (2010) were able to show the variability of MCS occurrence across the tropics (Fig. 46). The patterns of frequency of occurrence of MCSs differ from land to ocean and by size of MCS. In general, the largest ones occur over oceans and smaller ones concentrate over land.

In late 1997, the launch of the TRMM satellite provided the first precipitation radar (the Ku-band TRMM PR) in space. In 2006, the launch of CloudSat provided the first satellite-borne cloud radar (W-band). With these radars in orbit, it became possible to observe the *three-dimensional* structures of cloud systems around the entire globe. Yuan et al. (2011) used the CloudSat radar data to analyze the internal microphysical characteristics of the large anvil clouds of MCSs. However, CloudSat's radar is highly attenuated in precipitation, and for determining MCS characteristics it is crucial to observe the three-dimensional structure of radar echoes to distinguish convective and stratiform portions of MCSs. TRMM PR with its Ku-band radar has been the basis of determining the internal structures of precipitating convective systems from space. Although Ku-band suffers from attenuation, effective methods for correcting the attenuation have been applied to the TRMM PR data, and they have been a most effective way to characterize convective precipitation systems in low latitudes. Figure 47 is an example of the TRMM PR data in an MCS located over the tropical West Pacific. In this example the echo between ~70 and 200 km on the horizontal axis has a bright band and is clearly stratiform, while the deep vertical column of echo centered at ~210 km is

obviously convective. The National Aeronautical and Space Administration (NASA) and Japan Aerospace Exploration Agency (JAXA) routinely apply an algorithm that uses the three-dimensional echo structure to classify all of the TRMM PR echoes as convective, stratiform, or other (Awaka et al. 1997). The “other” category is only a tiny fraction of all the echoes seen by TRMM (Schumacher and Houze 2003; Funk et al. 2013).

Using only the TRMM radar echoes identified as convective, Liu and Zipser (2013) analyzed the frequency of occurrence of convective features that were mesoscale in horizontal dimension (contiguous convective echo covering $>1000 \text{ km}^2$). Such echoes are almost certainly indicating the existence of MCSs. Figure 48a shows the frequency of such echoes that were roughly circular in shape, while Fig. 48b shows where the elongated (“linear”) mesoscale convective echoes occur. This mapping led to several important conclusions that show how MCS characteristics vary regionally. Figures 48a and b both show that mesoscale convective echoes tend to be larger over land. However, using radar data from the GPM satellite, Liu and Zipser (2015) have shown that the precipitation areas are overall bigger over oceans (Figure 49). Combined with Liu and Zipser’s (2013) finding that the convective echo features are smaller over ocean, we may therefore infer that the stratiform regions of MCSs are larger over the tropical oceans than over tropical landmasses. Liu and Zipser (2013) also reported that the mesoscale convection is less deep over the oceans, especially in the intertropical convergence zone of the East Pacific, and as indicated in Figure 48b oceanic regions at low latitudes have a higher fraction of linear mesoscale convective entities than over land. Their results also show orientations of linear systems being controlled by factors such as warm ocean currents and fronts.

1198 Making further use of the three-dimensional radar echo fields provided by the
1199 TRMM PR, Houze et al. (2015) defined four categories of radar echoes: isolated shallow
1200 echo (ISE) is the same type of isolated warm-rain shower echo as that analyzed by
1201 Barnes and Houze (2013) in Figure 41; the deep convective core (DCC) is a three-
1202 dimensional echo object consisting entirely of echo > 40 dBZ extending to >10 km in
1203 maximum height), a wide convective core (WCC) is a three-dimensional echo object
1204 consisting entirely of echo > 40 dBZ and covering 1000 km² in horizontal dimension),
1205 and a broad stratiform region (BSR) is a contiguous stratiform echo covering >50 000
1206 km²). While these characteristics do not alone define an MCS, an MCS would normally
1207 contain a WCC. Indeed from results such as Pandya and Durran (1996), it is essential that
1208 intense convection must be grouped together to induce a mesoscale circulation of the type
1209 that characterizes an MCS (see Section 15). Thus a WCC is a proxy for a developing
1210 MCS. A BSR on the other hand would be found in an especially robust MCS and is a
1211 proxy for a mature MCS. Therefore, mapping the frequency of occurrence of WCCs and
1212 BSRs around the globe informs us as to the variability in location and occurrence of
1213 MCSs that appear to be in developing and mature stages. The definition of the DCC does
1214 not have an areal coverage requirement and thus does not necessarily indicate the
1215 presence of an MCS although such deep convection often precedes MCS development.
1216 The thresholds of 10 km, 1000 km² and 50,000 km² given in the definitions above are
1217 extreme and are referred to as “strong” thresholds. The same echo-object categories can
1218 be redefined with thresholds relaxed to “moderate” values of 8 km, 800 km² and 30,000
1219 km², respectively. The strong thresholds tend to give better depictions over land, while
1220 the moderate thresholds better characterize oceanic convection. However, Houze et al.

(2015) applied both thresholds over the entire band of TRMM PR observations in low latitudes.

Figure 50 contains analyses of the frequency of occurrence of the four categories defined by Houze et al. (2015) for the austral summer season. The warm oceans host dense populations of shallow isolated cumulonimbi (Fig. 50a). The most frequent occurrence of ISEs is in trade wind zones, where deeper convection is suppressed. However, it is also apparent from comparing Figs. 50a and d that ISEs are populous in zones where WCCs occur, i.e. where MCSs are most likely developing. It is thus possible that the self-aggregation process (Wing and Emanuel 2013; Emanuel et al. 2014) is operating to produce MCSs in these tropical oceanic regions. However, the land areas at low latitudes have almost no ISEs. So if the self-aggregation process operates over the land areas it must initiate with non-raining clouds or deeper precipitating convective elements. Comparison of Figs. 50a and b shows that the strong deep convective cores, where 40 dBZ echo extends above 10 km, occur almost exclusively over land, consistent with the findings of Zipser et al. (2006) and Liu and Zipser (2015), who have pointed out that these deepest and most intense cells do not correspond to where rainfall is maximum. For example, the DCCs defined by the 40 dBZ strong threshold do not occur over the Amazon region, which is where the South American rainfall is greatest, while they occur with high frequency over the relatively arid region of Argentina. However, if the thresholds for intensity and maximum height are relaxed to 30 dBZ and 8 km, respectively, DCCs are seen over the ocean and Amazon (Fig. 50c). In addition, wide convective cores, which are the convective areas of MCSs, generally occur at the locations of the moderate-threshold DCCs (cf., Figs. 50c and d). Thus, we see that MCSs

do not require the presence of the very deepest and most intense convection. MCSs form prolifically in the regions of moderately intense convective cores. Note: Fig. 50d shows the pattern for moderate threshold WCCs; the pattern for strong threshold WCCs (not shown) is similar and the same conclusions apply.

Houze et al. (2015) also mapped the occurrence of broad stratiform echo objects over all of the low latitudes. The BSRs and WCCs are both characteristics of MCSs, and Houze et al. (2015) found that the occurrence of BSRs maximizes in the same locations as WCCs. Hartmann et al. (1984) and Schumacher et al. (2004) have shown that more realistic representations of the mean general circulation of the tropics are obtained when the heating profile is top-heavy in a way that is consistent with the presence of MCS-generated stratiform precipitation. Figure 51 shows that the upper-level circulation across the Indo-Pacific Ocean sector is more robust when the precipitation is assumed to be 40% stratiform (Fig. 51b) versus 0% stratiform (Fig. 51a). This difference is consistent with the increase in potential vorticity in midlevels accompanying the more top-heavy heating profile associated with 40% of the precipitation being stratiform. However, in reality the stratiform contribution to the heating varies regionally. Schumacher and Houze (2003) showed stratiform rain fractions ranging from ~20–70% across the tropics. Confirming the inference from Liu and Zipser (2013), Houze et al. (2015) found that the frequency of stratiform precipitation associated with mesoscale convection in the tropics is greater over the oceans than over continental regions. Schumacher et al. (2004) showed that taking account of east-west variation in the stratiform rain fraction was important in representing El Niño circulation over the tropical Pacific. These findings indicate the importance of taking account of MCS occurrence in accurately representing the global

1267 circulation.

1268 **18. Representing MCS dynamics in global climate models**

1269 Planning of GATE in the early 1970s was motivated by the need for an
1270 observational basis for parameterizing convective clouds in global atmospheric models.
1271 The prevailing view (e.g. Yanai et al. 1973; Arakawa and Schubert 1974) was that the
1272 gap of scale between convective updraft plumes from the boundary layer and large-scale
1273 motions allowed for such parameterization. GATE then showed that the scale separation
1274 did not exist and that MCS were important forms of convective clouds (Houze and Betts
1275 1981). However, computing technology has improved and many numerical weather
1276 prediction models now resolve MCSs. But climate projection over tens to hundreds of
1277 years is still not feasible at global cloud-resolving resolution, and some way of
1278 representing MCSs in global climate models therefore will remain a need for some years
1279 to come.

1280 The previous section of this review has discussed how patterns of occurrence of
1281 MCSs vary over the globe (Section 17). We have also seen that large-scale environment
1282 dynamics control when and where MCSs form and grow (Section 16) and that the MCSs
1283 feed back to and influence their parent large-scale circulations (Section 12). In climate
1284 models running at coarse resolution, the controls and feedbacks between MCSs and the
1285 large-scale flow need to be parameterized in a way that is consistent with the layered
1286 overturning flow characterizing MCSs, the potential vorticity generation that occurs in
1287 the stratiform regions of longer-lived MCSs, and the ability of MCSs to disconnect from
1288 the boundary and occur in elevated form (Section 11). Climate models have particular
1289 difficulty in representing the diurnal cycle of precipitation in continental regions such as

the central U.S. (Dai 2006; Klein et al. 2006; Van Weverberg et al. 2017), and the inability to include MCSs, especially elevated MCSs (Section 11) and their relationship to diurnally varying low-level jets (Section 16) in coarse-resolution global models is likely to be part of the explanation for this shortcoming.

New approaches are being developed to accommodate MCS characteristics in climate models. One line of work builds on the Mapes et al.'s (2006) building block concept in which a population of convective clouds consists of three types of cloud: congestus, deep convection, and precipitating stratiform elements (Fig. 38). Khouider et al. (2010) formulated a stochastic model by dividing a coarse grid into a lattice of small-scale sites, each subject to a probability that a congestus, deep, or stratiform element will form. The probabilities for congestus and deep convection are based on Convective Available Potential Energy (CAPE⁶) and dryness is based on large-scale CAPE and midlevel dryness. If middle levels are dry, vertical development is restricted so that congestus can form but deep convection cannot. Once deep convection has formed at a lattice point it may develop over time into stratiform based on specified factors. All the cloud elements have prescribed lifetimes. This approach does not use large-scale shear environmental shear in formulating its probabilities, yet the large-scale shear is important in establishing the layered circulation and associated transport properties of an MCS.

In another approach, Moncrieff et al. (2017) allow a conventional convective parameterization to operate but apply an additional parameterization representing the layered overturning of MCSs. This additional parameterization consists of adding a top-heavy heating profile to the convective heating profile and a corresponding momentum

⁶ CAPE is the vertically integrated buoyancy of an undiluted parcel lifted from somewhere in the planetary boundary layer.

transport profile that is consistent with the momentum transport by the layered flow, as seen in data by Houze et al. (2000), in cloud resolving modeling by Mechem et al. (2002, 2006), and inferred from a general circulation model by Moncrieff and Klinker (1997). The profiles are applied wherever the convective parameterization is activated and are designed to be consistent with the Moncrieff et al. (1992) layered overturning MCS model. The profile magnitudes are controlled by tunable multiplicative coefficients. These coefficients have the potential of being functions of the large-scale shear, thus making this MCS parameterization consistent with the effects of shear in controlling MCS dynamics. This parameterization approach can be visualized conceptually as in Fig. 52, where a layered overturning on a larger-than-convective scale is applied where a convective population is occurring in an environment of large-scale shear. A notable feature of this approach is that the overall scale of the overturning may vary with the extent of the field of parameterized convection. Thus, the self-similarity of overturning in dynamic systems of various scales containing convective population (Section 16) is represented in the output of the climate model in which the parameterization is applied. Based on an analysis of data obtained in the Amazon region, Schiro et al. (2018) have found evidence of layered overturning dominating convective mixing and argue that approaches to parameterizing MCSs that emphasize deep-inflow mixing are desirable.

These new approaches to parameterizing MCS effects in climate and general circulation models, which take into account the layered overturning nature of MCSs, are still in their infancy but promise to provide ways of representing MCSs in models that must be run in coarse resolution to project changes in Earth's climate, water cycle, rainfall, and severe weather.

19. Current and future research on MCSs

Since the time of Hamilton and Archbold (1945), details of MCSs have emerged in tandem with developments in observational technology and modeling. They are seen as important elements of the global circulation and climate and producers of precipitation and flooding around the world. MCSs occur in disparate climatic regimes and take on a variety of forms, but have the common denominators of mesoscale horizontal scale and the development of stratiform portions that shift latent and radiative heating feedbacks upward into the mid-to-upper troposphere. Knowledge of these systems and how to represent them in forecasting and climate models remains an area of active and urgent research. The following are some areas of current and future research aimed at advancing knowledge, understanding, and the ability to represent MCSs accurately:

a. Initiation and upscale growth of convection to form MCSs.

Forecasting weather from MCSs and representing this form of convection in global climate models depends critically on determining when and where they will occur. All MCSs arise from an initial outbreak of deep convection, and the environmental factors leading to deep convection are well known. The crucial question is how deep convective elements, which each have a horizontal scale of only 1s to 10s of kilometers, organize themselves horizontally into a group of elements closely spaced within a zone ~100 km. It is this aggregation of deep convective elements that can then induce the mesoscale circulation of an MCS. The idealized “self aggregation” in an oceanic region of convection studied by Wing and Emanuel (2013) and others may be a partial answer to the initiation question. Their work especially highlights feedbacks between moisture

convergence and differential radiation between cloudy and less cloudy zones in sweeping convective elements into mesoscale zones. Whether such a process operates over land remains a question. Over land, inhomogeneities in land surface and topography may affect both initiation and early upscale growth. The effect of topography in the upscale growth of MCSs initiated near the Andes has been noted by Rasmussen et al. (2014), and two simultaneous field projects to investigate the upscale growth of these Andes systems (RELAMPAGO—Remote Sensing of Electrification, Lightning, and Mesoscale/microscale Processes with Adaptive Ground Observations, and CACTI—Clouds, Aerosols, and Complex Terrain Interactions) will take place in 2018.⁷

b. Forms of MCSs over the globe

As noted in this review, the horizontal patterns of convective elements within an MCS are varied. The canonical leading-line/trailing-stratiform organization is but one spatial arrangement. These variations in patterning are symptoms of different internal dynamics of the MCSs. The variations need to be understood if MCSs are to be predicted accurately either explicitly or in parameterized form. These variations can affect the degree to which the MCSs influence the potential vorticity of the large-scale environment and how they produce severe weather locally. Parker and Johnson (2000), Schumacher and Johnson (2005), and Peters and Schumacher (2015, 2016) have identified variations on the convective/stratiform structures of MCSs over a large portion of the U.S. They find that these variations are sometime related to the baroclinity of the synoptic environment. Rasmussen and Houze (2011) found variations in MCS structure over subtropical Pampas of South America to be similar to those seen over the U.S. Great

⁷ <https://publish.illinois.edu/relampago/>

Plains. However, the various structures and internal dynamics of MCSs seen in other regions such as tropical oceans (Leary and Houze 1979b; Yamada et al. 2010; Barnes and Houze 2014, 2016) and in the Maritime Continent (Houze et al. 1981) need to be unified into a common understanding of MCS internal structure and dynamics. The existence of datasets from past, current, and future satellite-borne radars (Liu and Zipser 2013; Houze et al. 2015) will be important in tying the different structures of MCSs seen around the world into a common unified global understanding of the occurrence of MCSs and their relationship to the large-scale circulation of the atmosphere.

c. MCSs and tropical cyclone genesis

As noted in the early study of Frank (1970), MCSs concentrated in the troughs of synoptic-scale tropical waves are precursors of tropical cyclones. In the modern era, Dunkerton et al. (2008) have emphasized how the growth of MCSs in environments of synoptic scale vorticity can concentrate vorticity and lead to tropical cyclone development. Ritchie et al. (2003) and Houze et al. (2009) have described examples of how MCSs rotate around a common center of larger-scale vorticity just prior to tropical cyclogenesis. Houze et al. (2009) used airborne Doppler radar to show that the convective updrafts in the region prior to the formation of the tropical cyclone were excessively strong and wide and thus capable of concentrating vorticity, as suggested by Montgomery et al. (2006). Figure 53 illustrates how the convective population within a pre-existing zone of cyclonic vorticity consists of both isolated convective elements and MCSs in different stages of development. The MCSs contain MCVs in midlevels due to the generation of potential vorticity in their stratiform regions (Section 12) while the convective-scale updrafts consist of vortical hot towers. Ritchie and Holland (1997),

1402 Simpson et al. (1997), and Ritchie et al. (2003) hypothesized that a tropical cyclone
1403 forms when the MCSs spinning around the low center reinforce the larger-scale
1404 depression. Bister and Emanuel (1997) suggested further that the stratiform-region
1405 vorticity builds downward as the region of MCSs becomes a tropical cyclone.
1406 Montgomery et al. (2006) argue on the other hand that the concentration of vorticity in
1407 the VHTs is critical to the cyclogenesis. The relative importance of these two
1408 hypothesized processes remains a key question in understanding the role of MCSs in
1409 tropical cyclogenesis.

1410 *d. Extreme precipitation and flooding and societal impacts*

1411 MCSs are often implicated in flooding, both of the slow rising and flash flood
1412 types. Schumacher and Johnson (2005) and Peters and Schumacher (2015, 2016) have
1413 shown some very specific types of MCS behavior that can inform flood forecasting in the
1414 central U.S. Rasmussen et al. (2015) have shown how the large-scale conditions favoring
1415 floods in the Asian monsoon can be anticipated 7–10 days in advance in global model
1416 ensemble forecasts, but that understanding the exact locations and details of the flooding
1417 events requires information on topography, soil moisture conditions, river drainage
1418 basins, and knowledge of how MCSs form in the specific synoptic conditions dominating
1419 at the times of the flood occurrences. Further research seamlessly combining global
1420 numerical prediction models, regional cloud resolving models, and hydrologic models is
1421 needed to improve forecasting of floods in regions vulnerable to flooding by MCSs,
1422 especially regions such as that of the Asian monsoon, where population and living
1423 conditions are highly weather sensitive.

1424 *e. Scale interaction and parameterization in global models*

1425 Besides forecasting flooding and other severe weather conditions on the
1426 timescales of days and weeks, global models are critical to projecting climate changes
1427 over longer time periods, for which convection-resolving models are at present still
1428 impracticable. As discussed in Section 18, it remains necessary to parameterize deep
1429 convection in models used for assessing global climate change. No presently used
1430 parameterization scheme represents deep convection in the form of MCSs, although
1431 several methods, noted in Section 18, are under development. Such research on
1432 parameterization of MCSs is essential because of the importance of MCSs in determining
1433 where rainfall occurs on average and because (as discussed in Sections 11-17) the
1434 dynamical feedbacks of MCSs to the large-scale are a factor in determining the evolving
1435 large-scale circulation. The parameterization methods in Section 18 will therefore remain
1436 an important and urgent area of research over the coming years.

1437 The underlying difficulty in establishing these parameterization methods is the
1438 interplay of different scales in MCSs. The primary characteristic of MCSs is the layered
1439 overturning that occurs on horizontal scales ~ 100 s of kilometers (Section 8). However,
1440 deep, narrow columns of buoyant convective updrafts ~ 1 -10 km are embedded within the
1441 general overturning. In the boundary layer, cold pools < 1 km in depth spread forward
1442 and rearward for 100s of kilometers. In addition, liquid and ice phase microphysics and
1443 associated radiative processes must be parameterized on all of these scales. The
1444 propagation of MCSs is affected not only by spreading of cold pools but by the
1445 movements of gravity waves or fronts with which the MCSs are entangled. When
1446 elevated MCSs occur, cold pools in the boundary layer become irrelevant, so that

parameterizations based on cold pool dynamics do not apply. Until further research establishes how all of these processes interact, parameterization cannot be based on firm understanding. Further field campaigns designed to enlighten these processes may be necessary. Virtual field programs involving only modeling and existing datasets (Moncrieff et al. 2012; Waliser et al. 2012) offer additional opportunities to test parameterization concepts. Much research in the area of overlapping and interacting scales within and connected with MCSs thus remains.

f. Aerosol, global warming, and MCSs in a changing climate

At the same time that understanding of MCSs and their importance has been increasing, the environment in which they occur has been changing in ways that can affect their dynamics and microphysics, as well as where and when they occur. In an environment of global warming, circulation patterns may change such that the favorable locations for MCSs in relation to land, sea, and mountains may shift. In particular, the changing juxtaposition of the general circulation and mountain ranges may affect the occurrence of flooding in the monsoon regions of South Asia (Rasmussen et al. 2015). It is therefore important to continue to monitor MCS occurrence by satellite, as examined in the studies of Liu and Zipser (2013) and Houze et al. (2015), and to use global modeling to project future seasonal and geographic patterns of occurrence of MCSs.

In addition, the environmental conditions affecting the nature and intensity of individual MCSs are changing. In studies such as those of Moncrieff and Zipser, described in previous sections of this review, the structure and dynamics of MCSs have been determined as functions of the temperature, water vapor, wind shear of the larger-scale environment. Because of increased anthropogenic pollution of the atmosphere, the

aerosol content of the surroundings has become an additional important environmental determinant of MCS structure. Research on this topic is still in its infancy; however, certain factors are beginning to emerge, and the results are showing that changes in the nature of MCSs as a result of aerosol in the environment are nonlinear and subtle. When considering ensembles of convective clouds, as occur over tropical oceans, some results indicate that aerosol indirect effects associated with shallow cumulus may offset or partially compensate for the aerosol indirect effects associated with congestus and deep convective clouds (van den Heever et al. 2011). It is therefore important to consider MCSs as a specific cloud type when evaluating aerosol environmental effects. Fan et al. (2013) found that a primary effect of aerosol in deeper convective systems is to alter the nature of the stratiform anvil. Their study indicates that increased environmental aerosol leads to smaller but more numerous ice particles in the anvil clouds. The anvils become broader and thicker under these conditions. Because of the importance of stratiform regions in the feedback of MCSs to the large-scale flow (Section 12), this finding has important implications for projecting circulation and climate changes. Saleeby et al. (2016) found that in highly polluted environments, riming of ice crystals in an MCS is greater, leading to less lofted cloud water and lower mixing ratios of ice in the anvils but nevertheless more numerous small ice crystals in the anvil and wider anvil areal coverage, with correspondingly significant effects on radiative transfer through the anvil, which, according to the discussion of Section 12, has implications for the heating profiles within MCSs. Clavner et al. (2018a, b) have found in the modeling study of one case of an MCS that a more polluted environment led to greater area of convective precipitation and a smaller area of stratiform rain. Marinescu et al. (2017) have shown that the vertical

profile of aerosol concentration in the environment is especially important in determining how aerosols affect MCSs (e.g., Fig. 54). Most aerosol observations are near Earth's surface, and Marinescu et al.'s (2017) study indicates how it may be ill advised in modeling studies to extrapolate surface measurements to higher altitudes. In a review of studies of aerosol and convective clouds, Fan et al. (2016) argued that there is an urgent need for observations of vertical profiles of aerosol in the environments of convective clouds. In a world of increasingly polluted environments, further examination of MCS response to the aerosol environment is one of the most urgent needs for future research. Areas such as India and China, where MCSs occur with severe consequences, the degree of pollution is some of the greatest in the world, and forecasting and projecting MCS occurrence cannot be accomplished without taking account of the aerosol environment.

20. Conclusion: The characteristics and importance of MCSs

This review has chronicled the history of the study of mesoscale convection. The MCS is the largest of the convective cloud phenomena. With its horizontal dimension of hundreds of kilometers, it exists near the energy-spectrum boundary between two and three-dimensional atmospheric turbulence. An MCS occurs when deep convective clouds congregate in a region $\sim 500\text{-}1000\text{ km}^2$. As the congregated clouds heat the troposphere by latent and radiative process, they induce a larger circulation, which is mesoscale in dimension and consists of layers of the atmosphere overturning: the rising layer emanates from the lower troposphere while the subsiding layer is drawn from midlevels. The lower-tropospheric layer feeding the ascending branch of the circulation can be several kilometers deep, i.e. the rising air is not necessarily rooted in the boundary layer. Sometimes the entire mesoscale overturning circulation lies above a layer of stable air,

1516 completely uncoupled from the boundary layer. This mesoscale layered circulation sets
1517 the MCS apart as a distinct phenomenon with its own set of dynamics.

1518 MCSs are familiar to weather forecasters and other students of severe weather.
1519 They account for a large portion of midlatitude and tropical rainfall (Fritsch et al. 1986,
1520 Yuan and Houze 2010), they produce severe weather and flooding (Houze et al. 1990;
1521 Rasmussen and Houze 2011), and they affect the larger-scale circulation of the
1522 atmosphere (Hartmann et al. 1984, Schumacher et al. 2004). It has taken the
1523 meteorological community over 70 years—since the time of Hamilton and Archbold
1524 (1945)—with ever improving observational methods and increasing sophisticated
1525 models, to reach this level of understanding.

1526 Satellite observations have shown the ubiquity of MCSs. Studies of spaceborne
1527 radar observations on board the CloudSat, TRMM, and GPM satellites have shown how
1528 MCSs (or their proxy radar-echo signatures) are distributed over the globe (Liu and
1529 Zipser 2013; Houze et al. 2015). These studies further show that that MCSs are not all
1530 alike. MCSs may or may not contain convection arranged in lines or bands. Over oceans,
1531 MCSs have somewhat less intense convective elements but larger and more robust
1532 stratiform regions than do MCSs over continents. Some regions, such as Amazonia, have
1533 MCSs whose characteristics lie between those of purely oceanic or dry continental
1534 MCSs.

1535 Despite such morphological differences from one climatological regime to
1536 another, MCSs always contain stratiform as well as convective precipitation. The
1537 stratiform may be in the form of a stratiform zone following a propagating squall line, but
1538 often the stratiform precipitation forms more or less in place where active convection

weakens while new convection forms nearby. The presence of a robust stratiform region implies that the net heating profile of the MCS (latent and radiative heating combined) is top-heavy in proportion to its stratiform rain fraction. This top-heavy profile generates potential vorticity in midlevels. Sometimes this potential vorticity is manifested in a mesoscale convective vortex in midlevels of the stratiform region. Whether or not such a vortex is manifest, the net effect on the large-scale circulation of an MCS is to inject potential vorticity at midlevels and thus affect the future course of the large-scale circulation in which it is embedded. This heating profile of MCS is one of the most crucial features of MCSs to be captured in a large-scale model, especially if that model aims to understand the role of MCSs in further changes of the global climate. Models with high resolution capture MCSs fairly accurately, and as cloud microphysical parameterizations improve these simulations, will become even more accurate. Such models now being run over limited regions of Earth are showing aspects of the role of MCSs in climate change (Prein et al. 2015, Yang et al. 2017).

Models with coarser resolution, such as climate models used to project changes over the entire globe over long periods, must either wait for computing technology to advance sufficiently to accommodate global high resolution models running over centuries of model time, or develop appropriate parameterizations of MCSs. Convective parameterizations based on a separation of scales between convective and synoptic scales will not suffice because of the horizontal dimension of MCSs. Several methods of parameterizing the heating and momentum transport profiles of MCSs in climate models are under development. Successful development of climate models that include MCSs, whether by cloud resolving modeling or parameterization, is critical because MCSs

remain an important societal problem. They are implicated in flooding and severe weather around the world. For example, floods in India and Pakistan, which result in massive casualties and human suffering, often involve MCSs (e.g. Houze et al. 2011; Rasmussen et al. 2015). MCS characteristics are affected by the increasingly polluted aerosol environment in many locations of the world, and as Earth warms the patterns of MCS occurrence will likely shift so that the areas affected by MCSs will change. Forecasting MCSs both in real time and projecting their future occurrence in a changing climate therefore remains a grand challenge for meteorology and climate.

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2174 **Figure Captions**

2175

2176 Figure 1 (a) Surface wind and squall line, and (b) cloud outline and circulation in a
2177 vertical plane in a West African “disturbance line.” From Hamilton and Archbold
2178 (1945)

2179 Figure 2 Time cross section through squall line and cold front, Wilmington, Ohio, 0730-
2180 1135 EST 29 May 1947. Heavy lines show boundaries of squall-front and polar-
2181 front layers; heavy dotted lines, boundaries of subsidence inversion. Light solid
2182 lines isotherms (0°C); light dashed lines, isolines of mixing ratio (g kg^{-1}). Below
2183 cross section is the time of radiosonde observation before or after squall-line
2184 passage; distance scale is in miles. From Newton (1950).

2185 Figure 3 Schematic section through a squall-line. Adapted from Fujita (1955).

2186 Figure 4 Schematic structure of surface precipitation features seen in early
2187 meteorological radar data. Adapted from Ligda (1956).

2188 Figure 5 Examples of radar data from the rain areas of midlatitude MCSs. Low elevation
2189 reflectivity patterns from the National Severe Storms Laboratory radar located at
2190 Norman (NOR), Oklahoma, are indicated by shading levels corresponding to 20–
2191 24 dBZ (light gray), 25–34 dBZ (dark gray), 35–44 dBZ (black), 45–54 dBZ
2192 (white), 55–64 dBZ (light gray), and >65 dBZ (dark gray). Range rings are at 20,
2193 200, and 240 km. Registration marks on outermost ring are at 90 azimuth
2194 intervals (north toward top of page). From Houze et al. (1990).

2195 Figure 6 Infrared image showing temperature ranges corresponding to the various grey
2196 shades. From Maddox (1980).

2197 Figure 7 Schematic cross section through a squall line system observed over the eastern
2198 tropical Atlantic Ocean. Streamlines show flow relative to squall line. Thin
2199 dashed streamlines show convective updraft circulation. Thin solid Streamlines
2200 show convective-scale downdraft circulation associated with mature squall line
2201 element, and wide arrows show mesoscale downdraft below the base of the anvil
2202 cloud. Wide, dashed arrows show mesoscale ascent in the anvil. Dark shading
2203 shows strong radar echo in the melting band and in the heavy precipitation zone
2204 of the mature squall line element. Light shading shows weaker radar echoes.
2205 Scalloped line indicates visible cloud boundaries. Adapted from Houze (1977) by
2206 Houze and Betts (1981).

2207 Figure 8 Total rain integrated over the convective (circled points) and stratiform regions
2208 of a squall line MCS located over the eastern tropical Atlantic Ocean. The data
2209 were obtained by three shipborne radars. The symbols •, +, and x indicate
2210 different methods used for combining the information from the three radars. From
2211 (Houze 1977).

2212 Figure 9 Stratiform rain fraction obtained from TRMM precipitation radar data. From
2213 Schumacher and Houze (2003).

2214 Figure 10 Conceptual model of a tropical oceanic squall line with trailing stratiform
2215 precipitation. All flow is relative to the squall line, which was moving from right
2216 to left. Numbers in ellipses are typical values of equivalent potential temperature
2217 (K). Adapted from Zipser (1977).

2218 Figure 11 Airflow pattern inferred by multiple-Doppler radar synthesis in the convective
2219 region of a tropical squall-line system observed by dual-Doppler radar in Ivory

2220 Coast, West Africa, on 23 June 1981. System is moving from right to left.

2221 Vertical arrow indicates scale of airflow vectors. Horizontal arrow C shows

2222 velocity of individual convective cells. Airflow vectors are computed relative to

2223 the cells. Contours show radar reflectivity (dBZ). From Roux (1988).

2224 Figure 12 Conceptual model of the kinematic, microphysical, and radar-echo structure of

2225 a convective line with trailing stratiform precipitation viewed in a vertical cross

2226 section oriented perpendicular to the convective line (and generally parallel to its

2227 motion. Medium and dark shading indicate intermediate and strong radar

2228 reflectivities. Adapted from Houze et al. (1989).

2229 Figure 13 (a) System relative winds at the 4-km level derived from airborne Doppler

2230 radar data within a bow echo on 10 June 2003. Aircraft tracks are superimposed.

2231 Reflectivity in dBZ is in color. (b) Vertical cross section along the white line in

2232 (a) of Doppler derived storm relative flow in the plane of the cross section.

2233 Negative velocities (yellow, green, and blue colors) recede from the convective

2234 line while positive velocities (brown, red, and magenta colors) approach the line.

2235 The vector scale [shown in the upper right of (b)] is vertically stretched to match

2236 the aspect ratio of the plot. Panel (a) is from Davis et al. (2004). Panel (b) is from

2237 Jorgensen et al. (2004).

2238 Figure 14 Schematic of airflow in the stratiform regions of an MCS over the western

2239 tropical Pacific as observed by airborne Doppler radar in TOGA COARE. The

2240 numbers indicate the observed ranges of values of the horizontal relative wind

2241 velocity and the horizontal scale of the midlevel inflow. Based on figures and

2242 tables of Kingsmill and Houze (1999).

2243 Figure 15 Idealization of a horizontal map of radar reflectivity (b) divided into convective
2244 and stratiform regions (a). Light gray represents the lowest reflectivity. Arrows in
2245 (a) show possible midlevel inflow directions. Arrows in (b) show possible low-
2246 level inflow directions. Adapted from Houze (1997).

2247 Figure 16 Schematic of airflow in the convective regions of an MCS over the western
2248 tropical Pacific as observed by airborne Doppler radar in TOGA COARE. The
2249 numbers (from bottom to top) indicate the observed ranges of values of the depth
2250 of the inflow layer, horizontal relative velocity of inflow and outflow air currents,
2251 the slope of the updraft (angle measured relative to the ocean surface), horizontal
2252 relative outflow velocities, and the width of the divergent region aloft. The
2253 horizontal directional differences of the low-level updraft inflow and midlevel
2254 downdraft inflow were often significantly different from 180°. Based on figures
2255 and tables of Kingsmill and Houze (1999).

2256 Figure 17 Heuristic diagram indicating horizontal vorticity (+ and -) in the vicinity of a
2257 long-lived line of multicell storms. Profile of horizontal wind component normal
2258 to the line is shown on the right of each panel. Frontal symbol marks outflow
2259 boundary. (a) Case in which there is no wind shear normal to the line in the
2260 environment. (b) Early stage of line development in case of low-level shear ahead
2261 of the gust front. (c) Late stage of line development in case of low-level shear
2262 ahead of the gust front. Adapted from Rotunno et al. (1988), Weisman (1992),
2263 and Weisman and Rotunno (2004).

2264 Figure 18 Vertical profiles of layer-mean storm-relative pre-MCS winds for MCSs
2265 containing lines of convection. Wind vectors depicted as line-parallel ($\otimes$) and

2266 line-perpendicular ($\rightarrow$) components in m s^{-1} . Layers depicted are 0–1, 2–4, 5–8,
 2267 and 9–10 km. Typical base scan radar reflectivity patterns (shading) and
 2268 hypothetical cloud outlines are drawn schematically for reference. MCS leading
 2269 edges are to the right. From Parker and Johnson (2000).

2270 Figure 19 Data obtained during GATE from an acoustic sounder on board the ship
 2271 *Oceanographer* on 4 September 1974. Black echoes extending from top to bottom
 2272 of chart are from precipitation. Rectangular echoes at 1150–1200, 1845–1940 and
 2273 2015 GMT are from tethered balloon instruments flown from the *Oceanographer*.
 2274 Spikey echoes emanating from bottom of chart are from turbulent plumes.
 2275 Intermittent echoes between 300 and 600 m levels prior to 1325 are cloud echoes.
 2276 The continuous layer of echo after 1815 is from a stable layer. From Houze
 2277 (1977).

2278 Figure 20 Composite surface sensible heat flux (W m^{-2}) beneath an MCS observed in
 2279 GATE. Dots indicate hourly positions of ships *Gilliss* (G), *Meteor* (M), *Dallas*
 2280 (D), *Researcher* (R), and *Oceanographer* (O). Composite stability atop the mixed
 2281 layer inversion (m s^{-2}). From Johnson and Nicholls (1983).

2282 Figure 21 Schematic of the structure of the various features of an elevated MCS. Arrows
 2283 indicate system-relative flows. Adapted from Marsham et al. (2010).

2284 Figure 22 MCS in successive stages of development. Adapted from Houze (1982).

2285 Figure 23 Schematic vertical cross section of an idealized MCS with convective region
 2286 (CONV.), associated stratiform precipitation region and non-precipitating
 2287 cirriform anvil. LW and SW are long and shortwave radiation. Symbols are
 2288 defined in text. Adapted from Houze et al. (1980).

2289 Figure 24 (a) Idealized stratiform, deep convective, and shallow convective latent heating
 2290 profiles. The x-axis is meant to be nondimensional until a precipitation amount is
 2291 specified. (b) Total latent heating profiles for 0%, 40%, and 70% stratiform rain
 2292 fractions, assuming 3.5 m yr^{-1} rain accumulation. Adapted from Schumacher et al.
 2293 (2004).

2294 Figure 25 Conceptual diagram of the structure and development mechanism of a midlevel
 2295 PV maximum associated with an MCS—based on a case study. Thin arrows along
 2296 the ordinate indicate the vertical profile of the environmental wind. Frontal
 2297 symbols indicate outflow boundaries. Dashed lines are potential temperature (5 K
 2298 intervals) and solid lines are potential vorticity (PV, in intervals of $2 \times 10^{-7} \text{ m}^2 \text{ s}^{-1}$
 2299 K kg^{-1}). The system is propagating left to right at about $5\text{--}8 \text{ m s}^{-1}$ and is being
 2300 overtaken by air of high equivalent potential temperature in the low-level jet. Air
 2301 overtaking the vortex ascends isentropic surfaces, reaches its level of free
 2302 convection (LFC), and thereby initiates deep convection. Shading indicates cloud.
 2303 Adapted from Fritsch et al. (1994).

2304 Figure 26 Schematic vertical cross section and vertical profile of radar reflectivity (along
 2305 dashed line A-A' in the cross section) in horizontally uniform precipitation
 2306 associated with an anvil cloud. The anvil cloud occurs to the rear of intense
 2307 convective cells moving from right to left in the figure. The dark solid line is the
 2308 contour of minimum detectable radar echo, lighter solid lines and shading indicate
 2309 contours of higher reflectivity, and the scalloped line indicates the cloud
 2310 boundary. From Leary and Houze (1979a).

2311 Figure 27 Ice-particle data obtained on aircraft flights through nimbostratus in tropical

2312 mesoscale convective systems over the Bay of Bengal. Plots show relative
 2313 frequency of observation of ice particles of a particular type per minute of in-
 2314 cloud flight time as a function of flight-level temperature. From Houze and
 2315 Churchill (1987).

2316 Figure 28 Melting and freezing rates retrieved from dual-Doppler radar data for a squall
 2317 line MCS with its leading edge at about 100 km on the horizontal axis. From
 2318 Braun and Houze (1995).

2319 Figure 29 Schematics showing the location of hydrometeor types in an MCS over the
 2320 tropical ocean relative to: (a) the layered airflow crossing the stratiform region;
 2321 (b) the layer of upward slantwise motion entering in the convective region. The
 2322 percentages in the color bar indicate the average areal coverage of each particle
 2323 type. From Barnes and Houze (2014).

2324 Figure 30 Conceptual model of the development of nimbostratus associated with deep
 2325 convection. Panels (a), (c), and (e) show horizontal radar echo pattern at Earth's
 2326 surface with two levels of intensity at three successive times, t_o , $t_o + \Delta t$, and $t_o + 2\Delta t$.
 2327 Panels (b), (d), and (f) show corresponding vertical cross sections. A sketch of the
 2328 visible cloud boundary has been added to the vertical cross sections. Asterisks
 2329 trace the fallout of three ice particles. From Houze (2014).

2330 Figure 31 Conceptual model of a buoyant updraft element. Dots indicate hydrometeors
 2331 suspended by updraft, downward pointing arrows indicate particles heavy enough
 2332 to fall through updraft, and horizontal arrows indicate lateral spreading of bubble.
 2333 Open arrows represent the vector field of the buoyancy pressure-gradient force.
 2334 From Yuter and Houze (1995).

Figure 32 Conceptual model of an ensemble of particle fountains in a multicellular MCS.

Shaded area represents radar reflectivity along a cross section normal to the convective region. Cloud boundary is indicated by the scalloped outline. Inset shows approximate scales and arrangement of the largest particle fountains relative to the radar echo. From Yuter and Houze (1995).

Figure 33. Results of a 3D simulation of a squall line MCS initiated with a line thermal

(with random perturbations) and periodic along-line boundaries. The initial sounding is moderately sheared, and the microphysics scheme is that of Thompson et al. (2008) microphysics. The domain is 800 km (cross-line) by 80 km (along line) x 20 km deep. Horizontal resolution is 2 km. The images are averaged over a 3.5 h period (between simulation hours 4.5 and 8), and the x - z images are also averaged along the line. (a) Radar reflectivity (color) and wind component (m s^{-1} , negative right-to-left) in the plane of the cross section relative to the storm. (b) potential temperature (color) and same relative wind component as in (a). Provided to the author by Professor Robert Fovell.

Figure 34 Schematic diagram showing the airflow relative to a two-dimensional, steady

state mesoscale convective system in a large-scale environment of given wind shear and potentially instability. The streamlines are those required by conservation of mass, momentum, entropy, and vorticity. Adapted from Moncrieff (1992).

Figure 35 Time averaged numerical model simulation of a squall line with trailing

stratiform precipitation. (a) Simulated radar reflectivity (in intervals of 5 dBZ). (b) Streamlines of system relative airflow. (c) Equivalent potential temperature

2358 (intervals of 3 K). Bold solid contour outlines cold pool (region of negative
 2359 potential temperature perturbation). Adapted from Fovell and Ogura (1988).
 2360 Figure 36 Streamlines for an MCS simulated by making a Wave-CISK assumption to
 2361 represent the latent heating. The dashed and dotted contours outline regions of
 2362 updraft and downdraft, respectively. From Raymond (1984).
 2363 Figure 37 Two-dimensional model simulation results for a leading-line/trailing stratiform
 2364 squall-line MCS. (a) Time-mean thermal forcing due to the leading convective
 2365 line alone. Contour interval is 0.001 K s^{-1} . (b) Horizontal velocity at time = 6 h
 2366 generated by the thermal forcing in (a) Horizontal velocity contours are at
 2367 intervals of 4 m s^{-1} . Arrows indicate direction of the horizontal flow. Cold pool
 2368 forward boundary is at $x=0$. Bold contour and shading emphasize layer inflow
 2369 constituting the layer ascent of air originating ahead of the storm and rising
 2370 through it. Adapted from Pandya and Durran (1996).
 2371 Figure 38 Depiction of three of cloud populations, made up of shallow convection, deep
 2372 convection, and stratiform elements. In (a) the fraction of shallow convection is
 2373 highest in the left-hand population. In (b) the fraction of deep convective elements
 2374 is highest. In (c) the fraction of stratiform elements is greatest. Adapted from
 2375 Mapes et al. (2006).
 2376 Figure 39 (a) Composite time–height sections of divergence (shading, 10^{-6} s^{-1}) and
 2377 anomalies of vertical velocity (contours, $10^{-3} \text{ hPa s}^{-1}$; solid lines indicate positive
 2378 values) calculated from reanalysis data composited around the time of maximum
 2379 in rain accumulation seen by the National Center for Atmospheric Research S-
 2380 PolKa radar located on Addu Atoll on the equator in the central Indian Ocean

2381 during DYNAMO. (b) Composite of the frequency of occurrence of different
2382 types (color-coded) of radar echo structure before (negative time) and after
2383 (positive time). The right y-axis is for the colored-coded frequency curves. The
2384 rainfall accumulation amounts shown on the left y-axis is the composite computed
2385 by centering each of the 11 rain episodes of DYNAMO on the time of the
2386 maximum of its running-mean curve. From Zuluaga and Houze (2013).

2387 Figure 40 Vertical structure of the MJO system. Primary airflow branches are sketched.
2388 All quantities are 20-day averages and relative to the traveling system. From
2389 Moncrieff (2004).

2390 Figure 41 Frequency of occurrence of different types of radar echoes seen by the TRMM
2391 Precipitation Radar over the Central Indian Ocean: Total frequency of (a) isolated
2392 shallow echoes and (b) broad stratiform regions The frequency is defined as the
2393 number of TRMM Precipitation Radar pixels in the central Indian Ocean that
2394 contain an isolated shallow echo normalized by the total number of TRMM PR
2395 pixels detected in the central Indian Ocean, which include both echo-covered and
2396 echo-free pixels. The frequency is reported as a percent. The blue lines show the
2397 frequency-phase series for the 20 realizations, the black line is the mean of the
2398 realizations, and the dashed red lines are 99% confidence intervals of the mean
2399 and are calculated using the Student's t-statistic. In the central Indian Ocean the
2400 indices for the active and suppressed phases of the MJO are 3 and 6, respectively.
2401 From Barnes and Houze (2013).

2402 Figure 42 Net heating from isolated shallow convective elements (red), deep convective
2403 cores (cyan), wide convective cores (dark blue), and broad stratiform regions

2404 (green) as seen by the TRMM radar during different stages of the MJO over the
 2405 central Indian Ocean. Net heating is normalized by the total number of pixels
 2406 sensed by the TRMM Precipitation Radar. From Barnes et al. (2015).

2407 Figure 43 Analysis of the 500 hPa level prior to MCC development. Heights (m) are
 2408 heavy solid contours, isotherms ($^{\circ}\text{C}$) are dashed, and mixing ratio (g kg^{-1}) is
 2409 indicated by light solid contours. Winds (full barb = 5 m s^{-1}) are plotted at every
 2410 other grid point and dark arrow shows axis of maximum winds. The crosshatched
 2411 region indicates terrain elevations above the 500 hPa level. From Maddox (1983).

2412 Figure 44 Results of a cloud-resolving model run for two warm seasons. The fields are
 2413 composites of the mean geopotential height (in meters, contour lines), wind
 2414 (vector), and equivalent potential temperature (in K, color filled) at 500 hPa at 0-3
 2415 h before MCS initiation. The dots indicate the locations of MCS initiation. The
 2416 box is the region in which MCS initiation was sought. From Yang et al. (2017).

2417 Figure 45 Sequence of 300-hPa analyses with geopotential height contours (solid)
 2418 analyzed in 12 dam intervals with the 35 and 45 m s^{-1} isotachs (dashed) at (a)
 2419 0600 CST 3 June, (b) 1800 CST 3 June, and (c) 0600 CST 4 June 1985. The bold
 2420 arrow schematically illustrates the approximate streamline of the low-level jet in
 2421 (b) and (c). From Trier and Parsons (1993).

2422 Figure 46 Frequency of occurrence of (a) “small separated,” (b) “large separated,” and (c)
 2423 “connected” tropical MCSs during the months December, January, and February,
 2424 as shown by satellite infrared and passive microwave sensors on the NASA Aqua
 2425 satellite. Small and large MCSs have anvil plus raining area $<10^4 \text{ km}^2$ and
 2426 $>2.25 \times 10^4 \text{ km}^2$, respectively. Connected MCSs have separate infrared cloudtop

2427 signatures but share a rain area. From Yuan and Houze (2010).

2428 Figure 47 Vertical cross section of radar reflectivity in dBZ seen by the TRMM PR along
 2429 the red line superimposed on the plan view at 4 shown in the inset. The vertical
 2430 cross section runs from left to right along the red line. Latitudes -2 to -6 are in the
 2431 Southern Hemisphere. Longitudes are in the Eastern Hemisphere. From Houze et
 2432 al. (2015).

2433 Figure 48 (a) Distribution of the population of large convective echoes (area $> 1000 \text{ km}^2$)
 2434 and with a near-circular shape (minor/major axis ratio > 0.6). (b) Distribution of
 2435 the population of large convective lines (convective echoes with minor/major axis
 2436 ratio < 0.2 and area $> 1000 \text{ km}^2$). From Liu and Zipser (2013).

2437 Figure 49 Locations of precipitation features (PFs) seen by the GPM satellite and
 2438 categorized by size. From Liu and Zipser (2015).

2439 Figure 50 Geographical distribution of the probability of finding various categories of
 2440 echoes seen by the TRMM Precipitation Radar during December-February: (a)
 2441 isolated shallow echoes; (b) strong threshold deep convective cores; (c) moderate
 2442 threshold deep convective cores; and (d) moderate threshold wide convective
 2443 cores. The black contour inside the continental regions represents the 700 m
 2444 elevation. The probability is on a scale of 0 to 1 and is computed as the number of
 2445 pixels identified as belonging to an echo category divided by the total number of
 2446 pixels sampled by the TRMM PR over the given time period within a 0.5° by 0.5°
 2447 grid element. From Houze et al. (2015).

2448 Figure 51 The 400-hPa latent heating (shaded) and the resulting 250-hPa streamfunction
 2449 anomalies (contours) computed from a simple climate model for the resting basic

2450 state and latent heating derived from the PR annually averaged precipitation field
2451 and geographically uniform stratiform rain fractions of (a) 0% and (b) 40%. The
2452 streamfunction contour interval is $2 \times 10^6 \text{ m}^2 \text{ s}^{-1}$. Negative contours are dashed.
2453 From Schumacher et al. (2004).

2454 Figure 52 Overlay of a convective cloud population and superimposed layered
2455 overturning. Adapted from Houze et al. (1980) and Moncrieff et al. (2017).

2456 Figure 53 Idealized distribution of convection in a low-pressure system (L) in the process
2457 of developing into a tropical cyclone. The distribution contains isolated
2458 convective cells and MCSs in various stages of development. The convective
2459 updrafts stretch the environmental vorticity and advect positive vorticity upward.
2460 These updrafts with concentrated positive vorticity are referred to as vortical hot
2461 towers (VHTs). The larger and longer-lived MCS contain VHTs in early MCS
2462 stages. The MCSs also contain stratiform regions with midlevel mesoscale
2463 convective vortex (MCVs). Both the VHTs and MCVs are thought to contribute
2464 to cyclogenesis. From Houze et al. (2009).

2465 Figure 54 Schematic of an MCS under high concentrations of (a) lower-tropospheric and
2466 (b) midtropospheric aerosol particles. The gray area represents cloudy regions,
2467 while the green shading represents rainfall with darker shading depicting heavier
2468 rain. The blue frontal symbols represent the leading cold pool boundary, and the
2469 white arrows represent the primary front-to-rear ascending flow. The three boxes
2470 represent three different precipitation regions of the MCS where different
2471 microphysical processes are governing the precipitation. Particle types are
2472 specified in the legends, with the amounts and sizes of hydrometeors

2473 representative of the relative concentrations and mean diameters of particles
2474 within those regions. From Marinescu et al. (2017).
2475

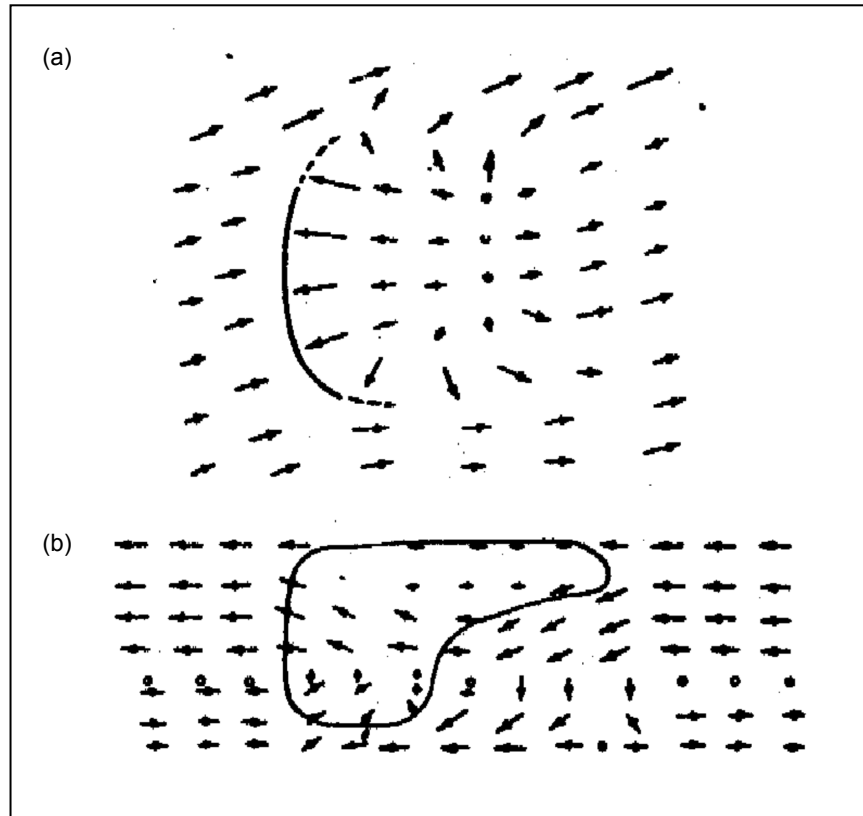


Figure 1. (a) Surface wind and squall line, and (b) cloud outline and circulation in a vertical plane in a West African "disturbance line." From Hamilton and Archbold (1945).

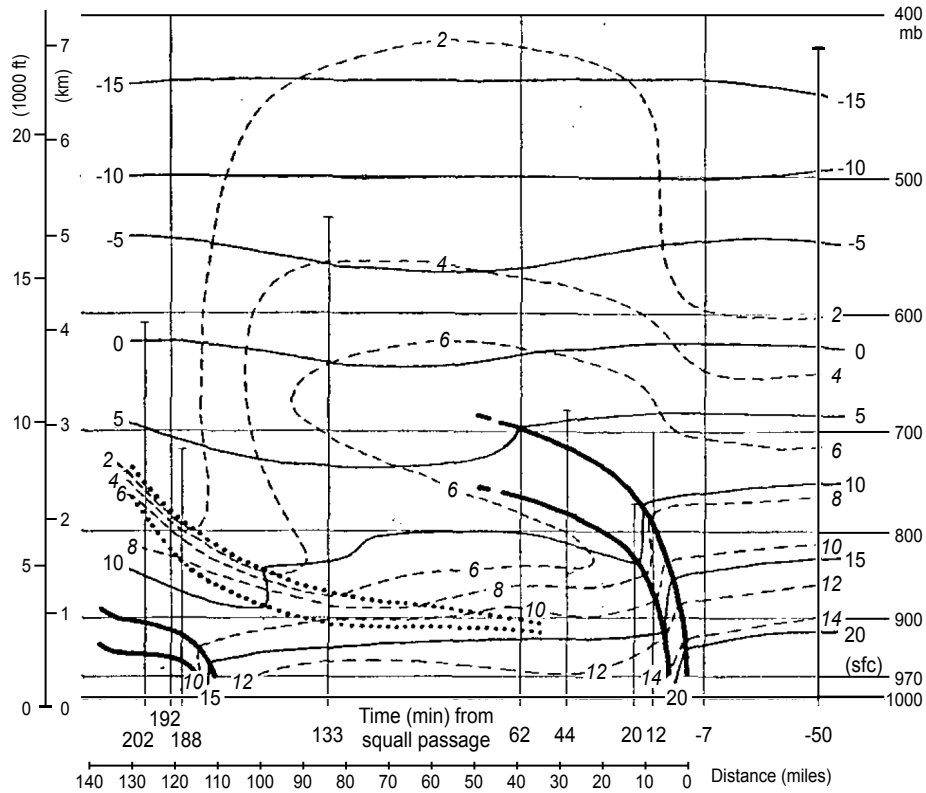


Figure 2. Time cross section through squall line and cold front, Wilmington, Ohio, 0730-1135 EST 29 May 1947. Heavy lines show boundaries of squall-front and polar-front layers; heavy dotted lines, boundaries of subsidence inversion. Light solid lines isotherms (0°C); light dashed lines, isolines of mixing ratio (g kg^{-1}). Below cross section is the time of radiosonde observation before or after squall-line passage; distance scale is in miles. From Newton (1950).

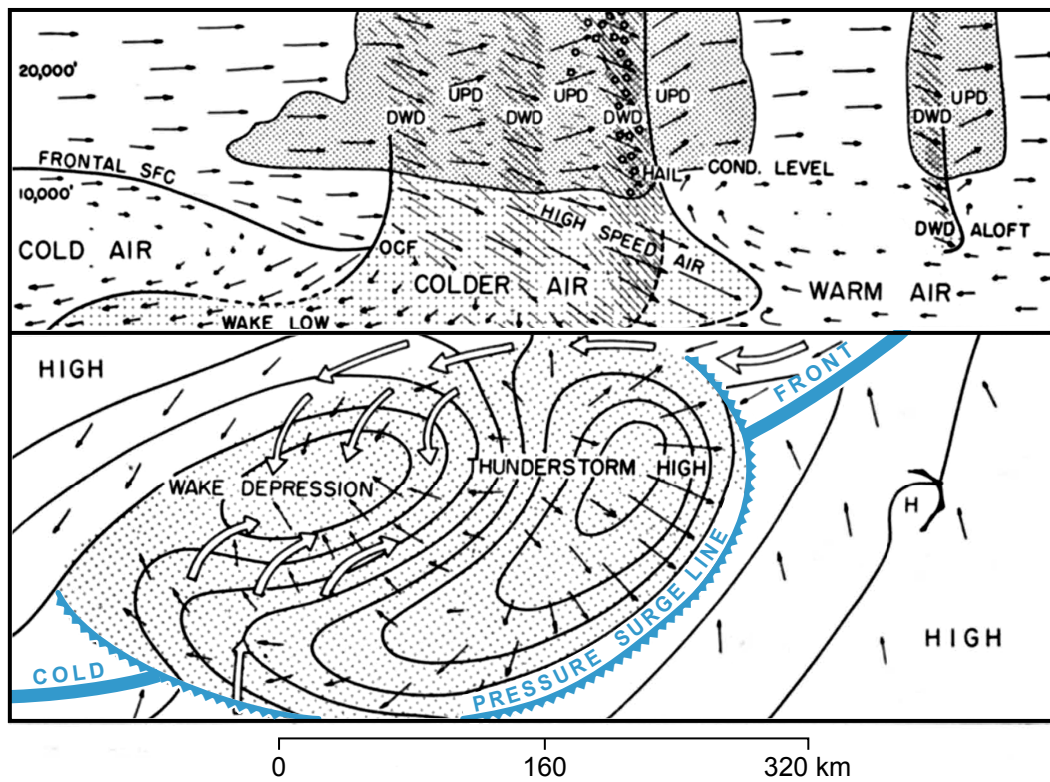


Figure 3. Schematic section through a squall line. Adapted from Fujita (1955).

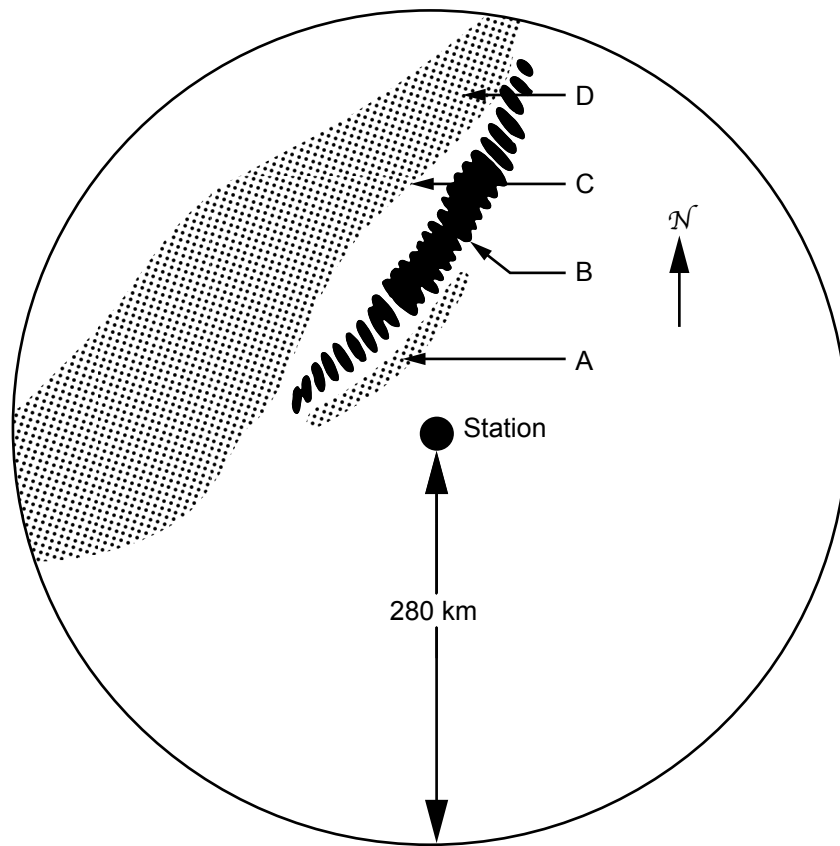


Figure 4. Schematic structure of surface precipitation features seen in early meteorological radar data. Adapted from Ligda (1956).

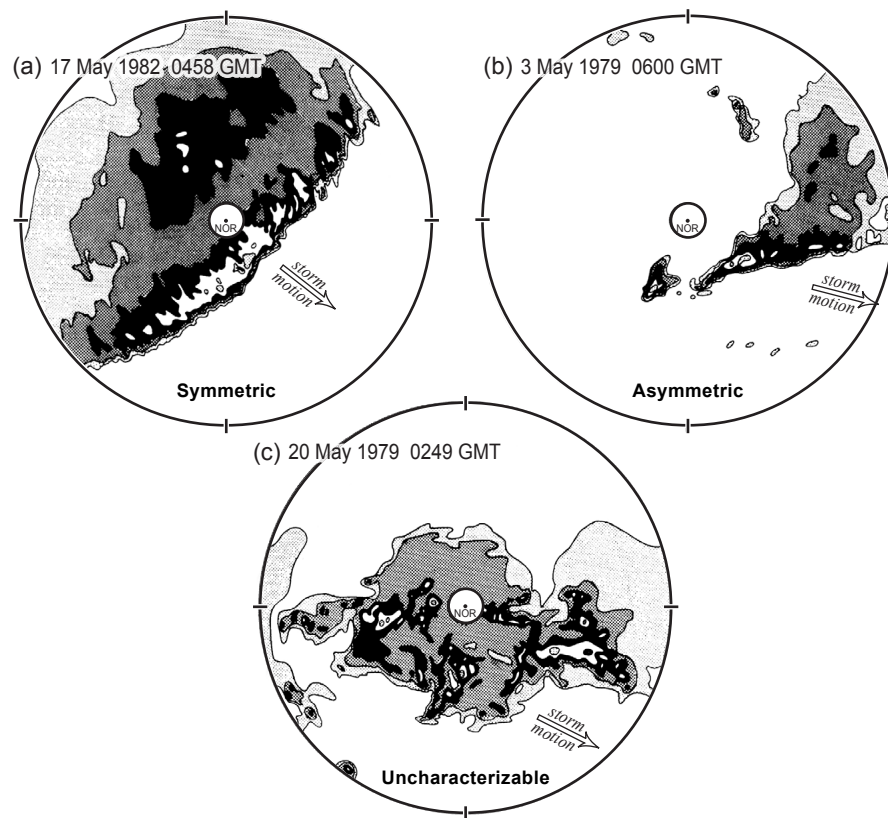


Figure 5. Examples of radar data from the rain areas of midlatitude MCSs. Low elevation reflectivity patterns from the National Severe Storms Laboratory radar located at Norman (NOR), Oklahoma, are indicated by shading levels corresponding to 20–24 dBZ (light gray), 25–34 dBZ (dark gray), 35–44 dBZ (black), 45–54 dBZ (white), 55–64 dBZ (light gray), and >65 dBZ (dark gray). Range rings are at 20, 200, and 240 km. Registration marks on outer ring are at 90 azimuth intervals (north toward top of page). From Houze et al. (1990).

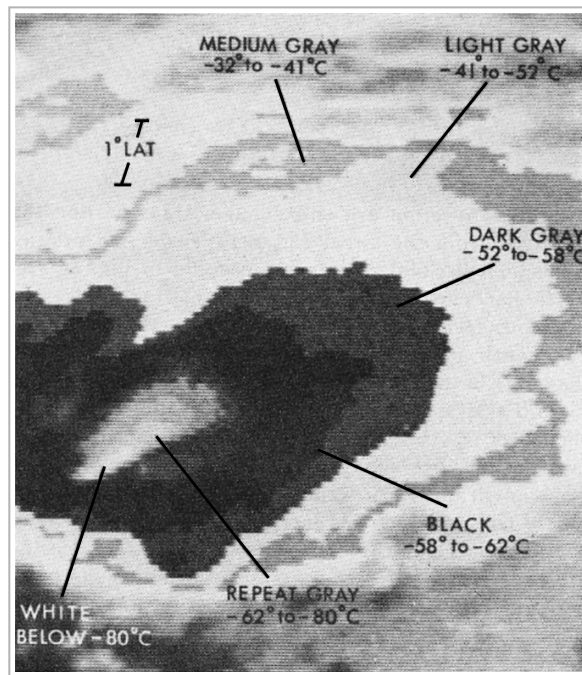


Figure 6. Infrared image showing temperature ranges corresponding to the various gray shades. From Maddox (1980).

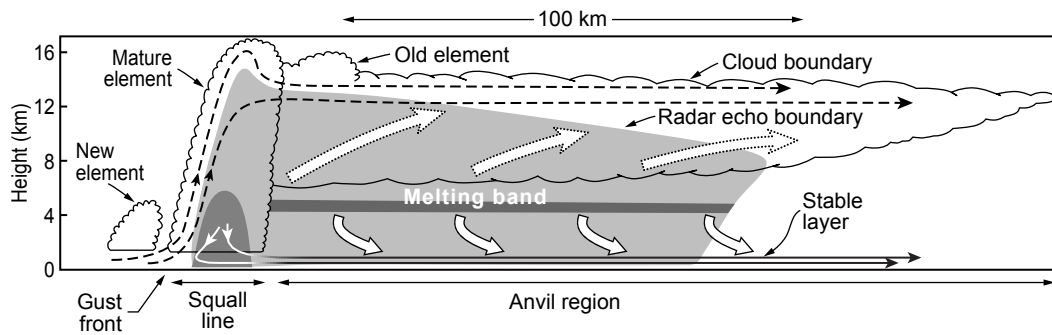


Figure 7. Schematic cross section through a squall line system observed over the eastern tropical Atlantic Ocean. Streamlines show flow relative to squall line. Thin dashed streamlines show convective updraft circulation. Thin solid streamlines show convective-scale downdraft circulation associated with mature squall line element, and wide arrows show mesoscale downdraft below the base of the anvil cloud. Wide, dashed arrows show mesoscale ascent in the anvil. Dark shading shows strong radar echo in the melting band and in the heavy precipitation zone of the mature squall line element. Light shading shows weaker radar echoes. Scalloped line indicates visible cloud boundaries. Adapted from Houze (1977) by Houze and Betts (1981).

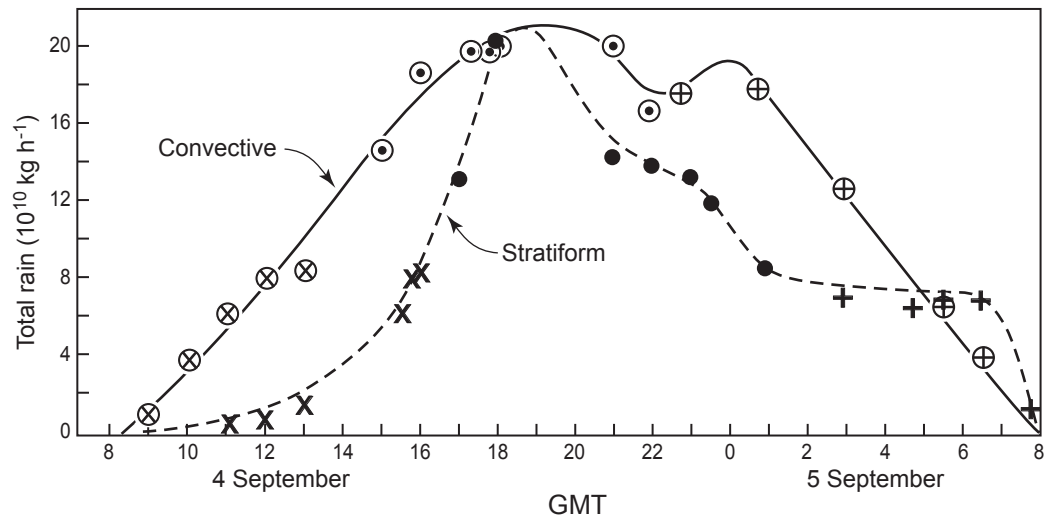


Figure 8. Total rain integrated over the convective (circled points) and stratiform regions of a squall line MCS located over the eastern tropical Atlantic Ocean. The data were obtained by three shipborne radars. The symbols •, +, and x indicate different methods used for combining the information from the three radars. From (Houze 1977).

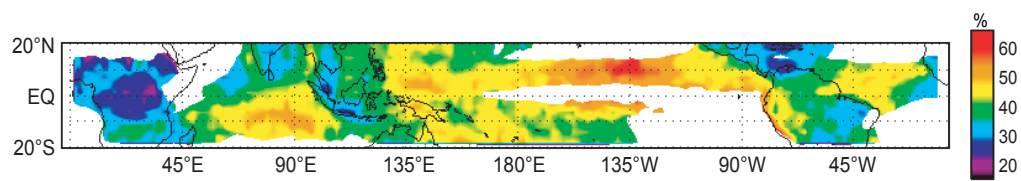


Figure 9. Stratiform rain fraction obtained from TRMM precipitation radar data. From Schumacher and Houze (2003).

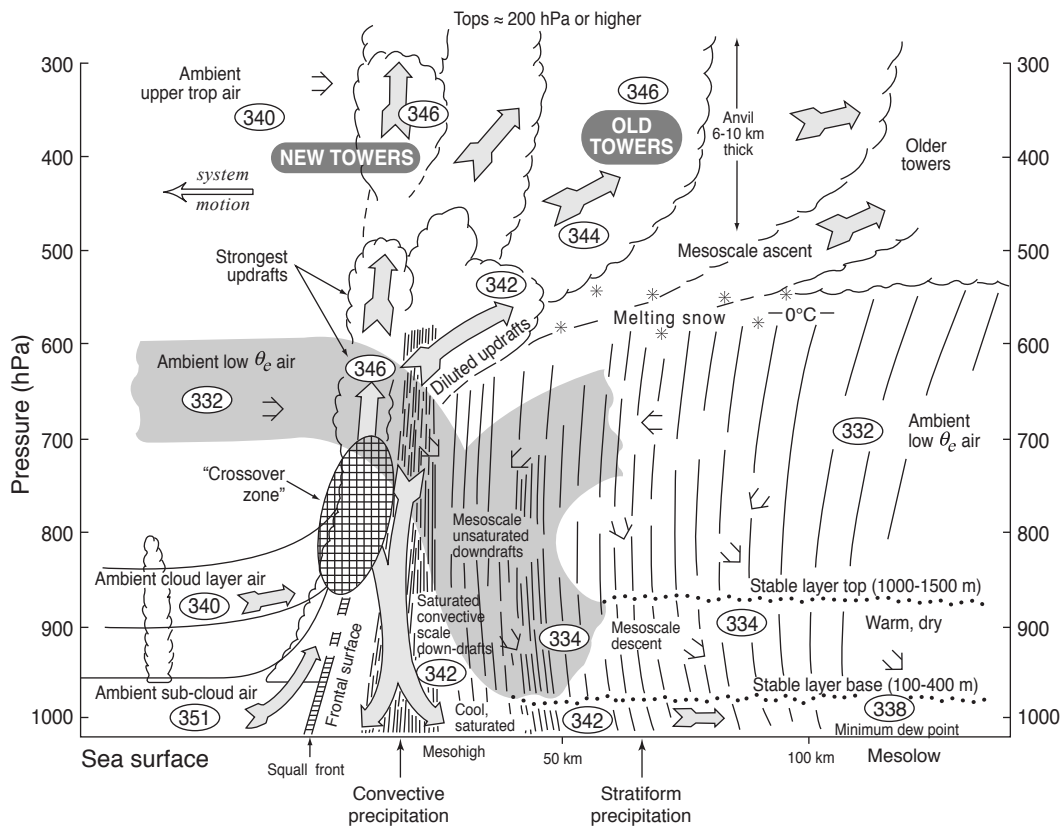


Figure 10. Conceptual model of a tropical oceanic squall line with trailing stratiform precipitation. All flow is relative to the squall line, which was moving from right to left. Numbers in ellipses are typical values of equivalent potential temperature (K). Adapted from Zipser (1977).

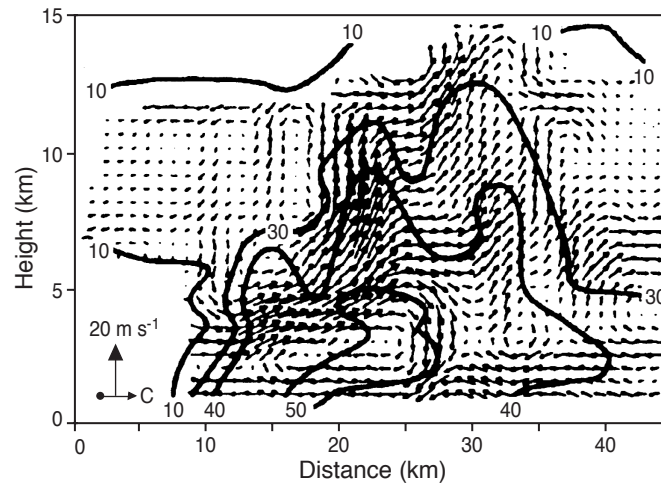


Figure 11. Airflow pattern inferred by multiple-Doppler radar synthesis in the convective region of a tropical squall-line system observed by dual-Doppler radar in Ivory Coast, West Africa, on 23 June 1981. System is moving from right to left. Vertical arrow indicates scale of airflow vectors. Horizontal arrow C shows velocity of individual convective cells. Airflow vectors are computed relative to the cells. Contours show radar reflectivity (dBZ). From Roux (1988).

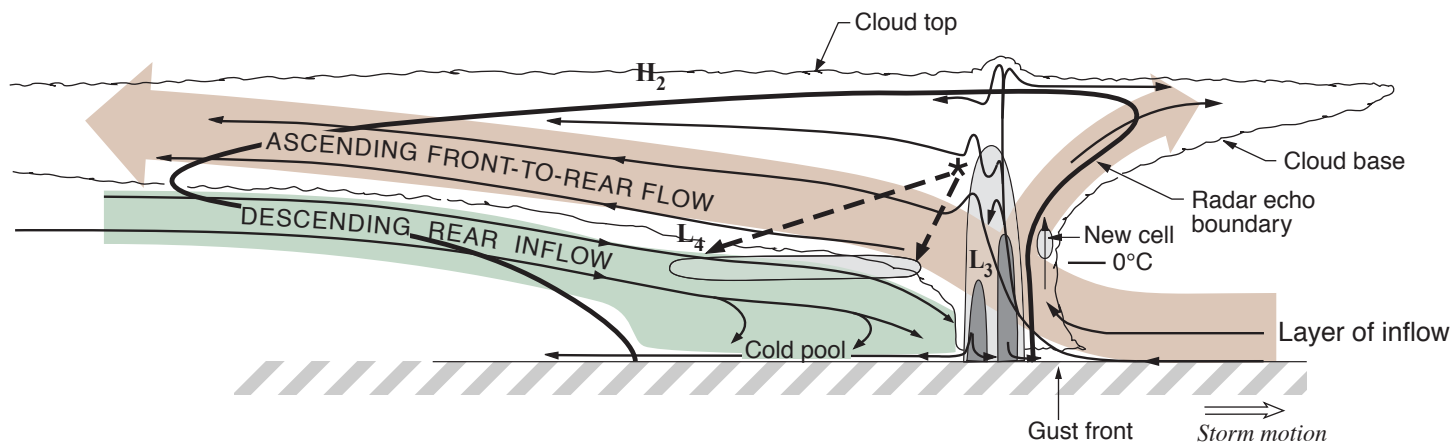


Figure 12. Conceptual model of the kinematic, microphysical, and radar-echo structure of a convective line with trailing stratiform precipitation viewed in a vertical cross section oriented perpendicular to the convective line (and generally parallel to its motion). Medium and dark shading indicate intermediate and strong radar reflectivities. Adapted from Houze et al. (1989).

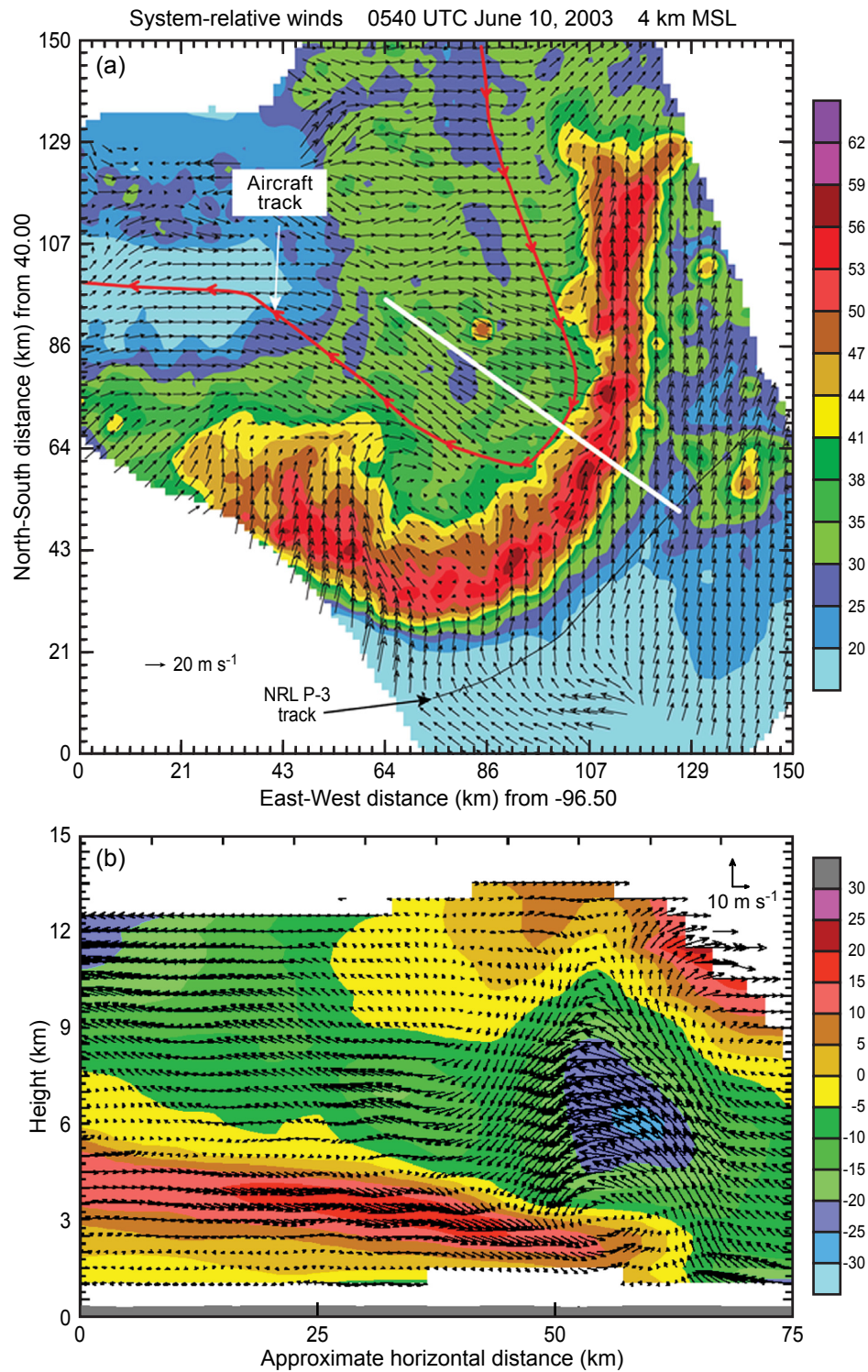


Figure 13. (a) System relative winds at the 4-km level derived from airborne Doppler radar data within a bow echo on 10 June 2003. Aircraft tracks are superimposed. Reflectivity in dBZ is in color. (b) Vertical cross section along the white line in (a) of Doppler derived storm relative flow in the plane of the cross section. Negative velocities (yellow, green, and blue colors) recede from the convective line while positive velocities (brown, red, and magenta colors) approach the line. The vector scale [shown in the upper right of (b)] is vertically stretched to match the aspect ratio of the plot. Panel (a) is from Davis et al. (2004). Panel (b) is from Jorgensen et al. (2004).

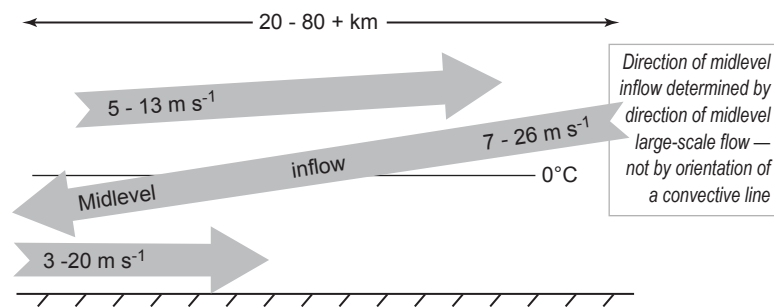


Figure 14. Schematic of airflow in the stratiform regions of an MCS over the western tropical Pacific as observed by airborne Doppler radar in TOGA COARE. The numbers indicate the observed ranges of values of the horizontal relative wind velocity and the horizontal scale of the midlevel inflow. Based on figures and tables of Kingsmill and Houze (1999).

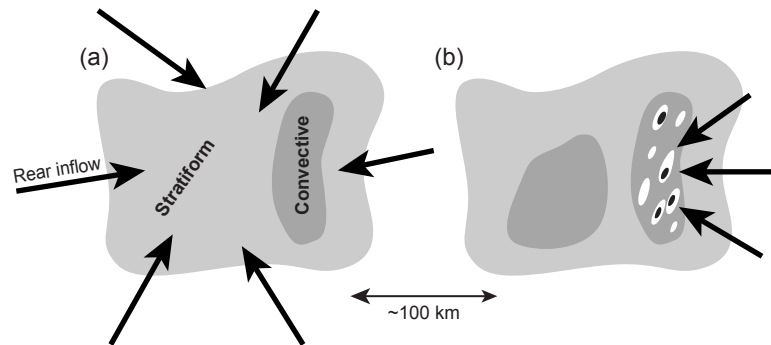


Figure 15. Idealization of a horizontal map of radar reflectivity (b) divided into convective and stratiform regions (a). Light gray represents the lowest reflectivity. Arrows in (a) show possible midlevel inflow directions. Arrows in (b) show possible low-level inflow directions. Adapted from Houze (1997).

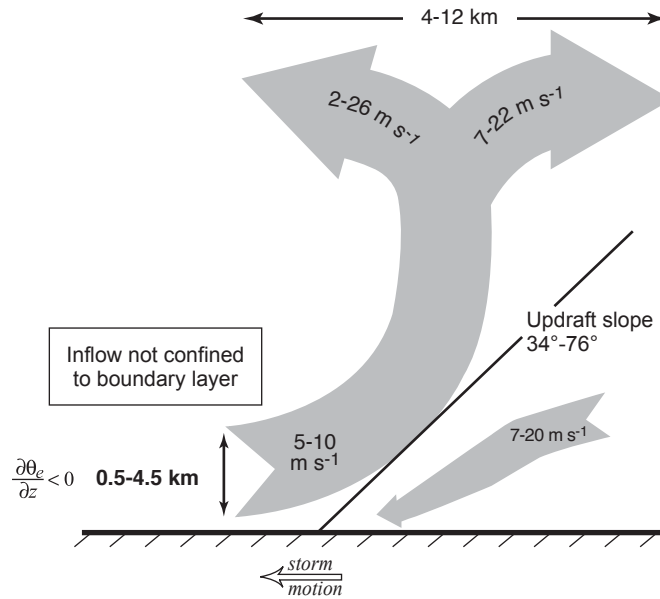


Figure 16. Schematic of airflow in the convective regions of an MCS over the western tropical Pacific as observed by airborne Doppler radar in TOGA COARE. Numeric labeling (from bottom to top) indicates the observed ranges of values of the depth of the inflow layer, horizontal relative velocity of inflow and outflow air currents, the slope of the updraft (angle measured relative to the ocean surface), horizontal relative outflow velocities, and the width of the divergent region aloft. The horizontal directional differences of the low-level updraft inflow and midlevel downdraft inflow were often significantly different from 180°. Based on figures and tables of Kingsmill and Houze (1999).

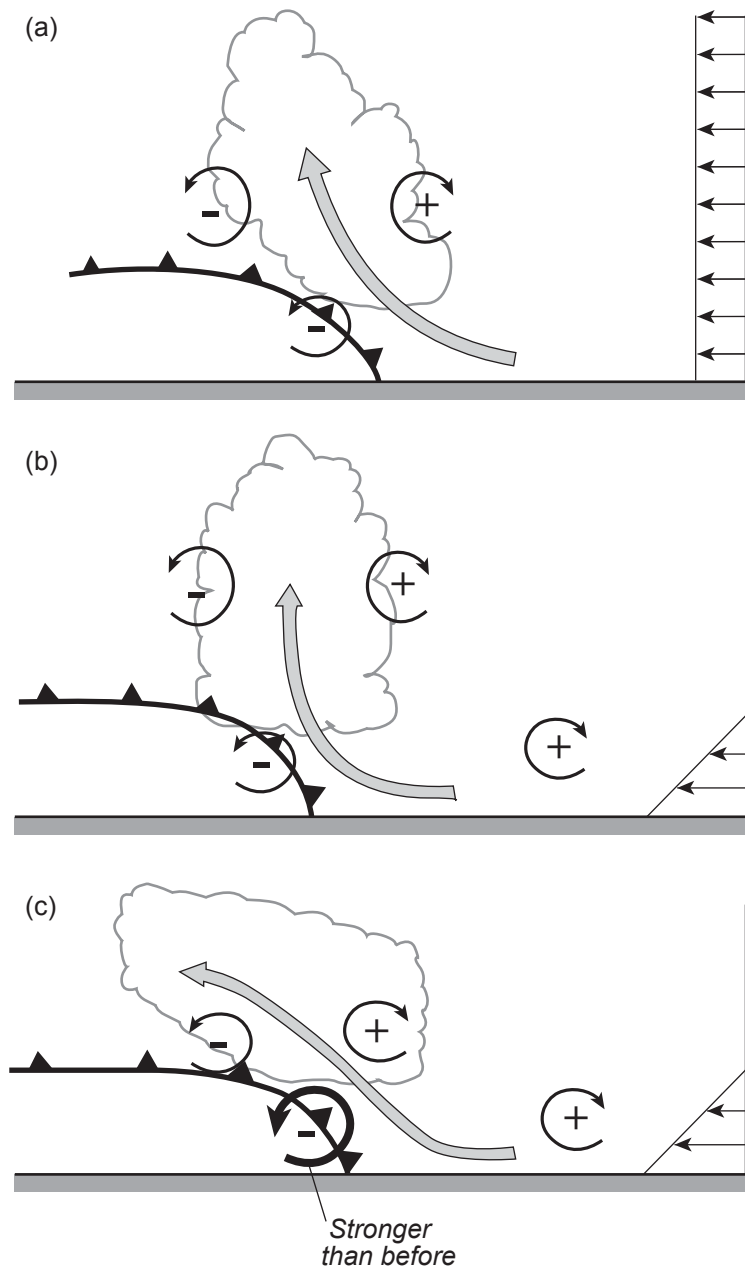


Figure 17. Heuristic diagram indicating horizontal vorticity (+ and -) in the vicinity of a long-lived line of multicell storms. Profile of horizontal wind component normal to the line is shown on the right of each panel. Frontal symbol marks outflow boundary. (a) Case in which there is no wind shear normal to the line in the environment. (b) Early stage of line development in case of low-level shear ahead of the gust front. (c) Late stage of line development in case of low-level shear ahead of the gust front. Adapted from Rotunno et al. (1988), Weisman (1992), and Weisman and Rotunno (2004).

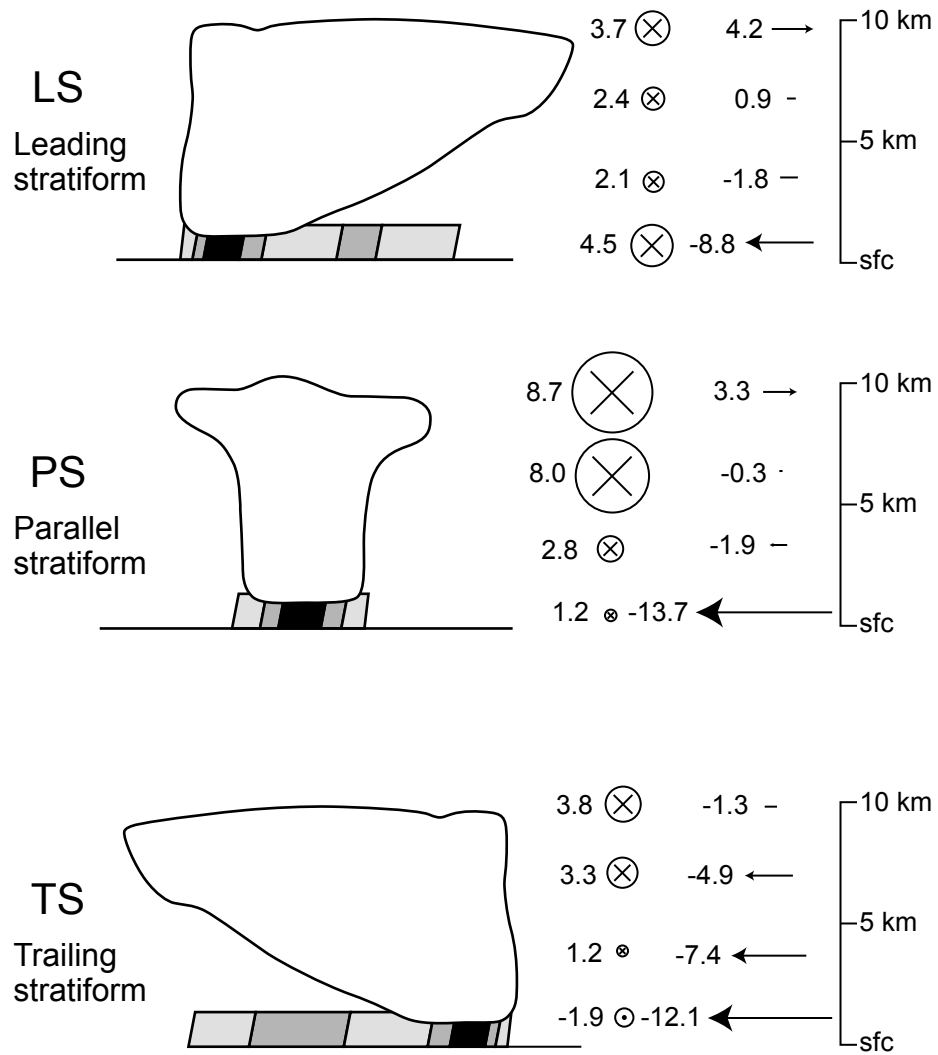


Figure 18. Vertical profiles of layer-mean storm-relative pre-MCS winds for MCSs containing lines of convection. Wind vectors depicted as line-parallel (⊗) and line-perpendicular (→) components in $m s^{-1}$. Layers depicted are 0–1, 2–4, 5–8, and 9–10 km. Typical base scan radar reflectivity patterns (shading) and hypothetical cloud outlines are drawn schematically for reference. MCS leading edges are to the right. From Parker and Johnson (2000).

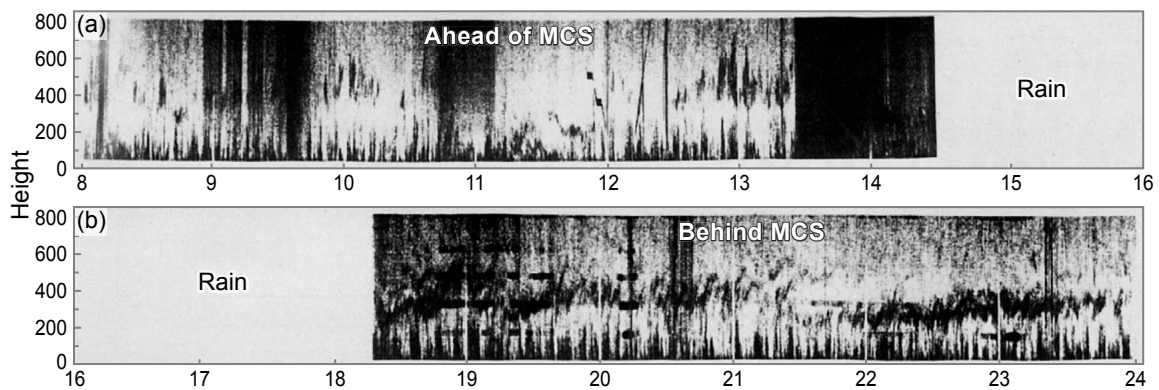


Figure 19. Data obtained during GATE from an acoustic sounder on board the ship *Oceanographer* on 4 September 1974. Black echoes extending from top to bottom of chart are from precipitation. Rectangular echoes at 1150-1200, 1845-1940 and 2015 GMT are from tethered balloon instruments flown from the *Oceanographer*. Spikey echoes emanating from bottom of chart are from turbulent plumes. Intermittent echoes between 300 and 600 m levels prior to 1325 are cloud echoes. The continuous layer of echo after 1815 is from a stable layer. From Houze (1977).

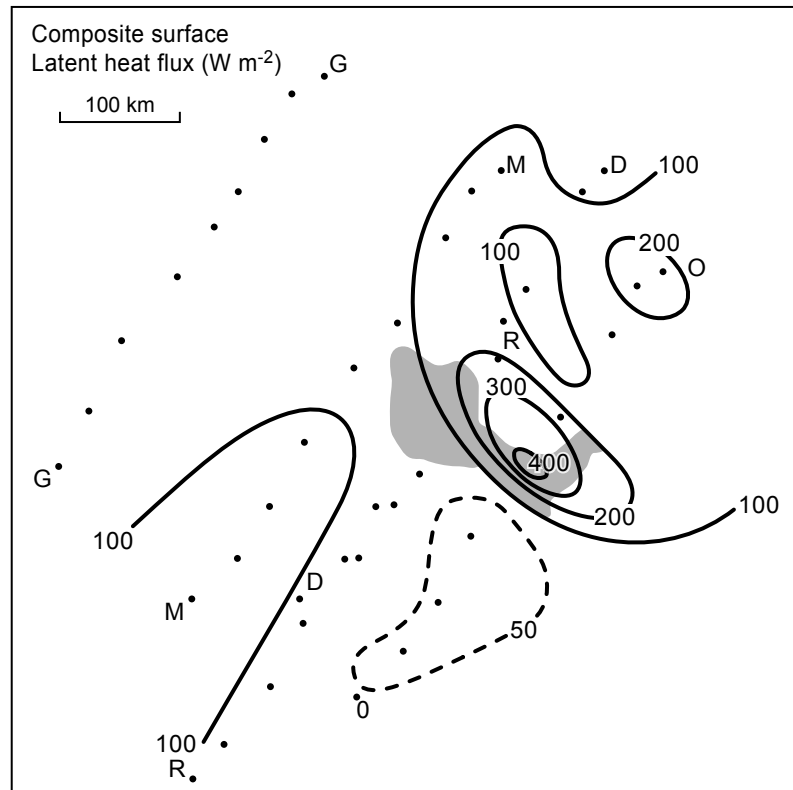


Figure 20. Composite surface sensible heat flux (W m^{-2}) beneath an MCS observed in GATE. Dots indicate hourly positions of ships *Gilliss* (G), *Meteor* (M), *Dallas* (D), *Researcher* (R), and *Oceanographer* (O). Composite stability atop the mixed layer inversion (m s^{-2}). From Johnson and Nicholls (1983).

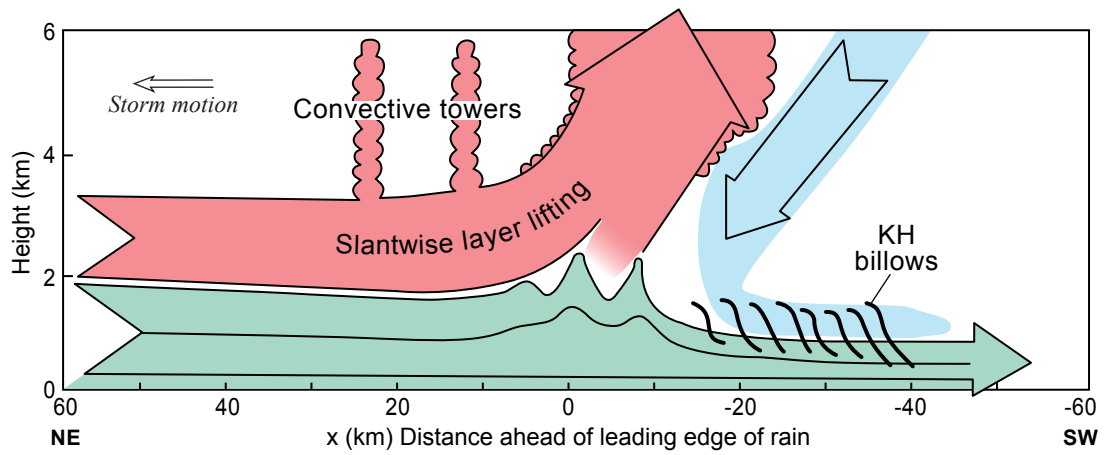


Figure 21. Schematic of the structure of the various features of an elevated MCS. Arrows indicate system-relative flows. Adapted from Marsham et al. (2010).

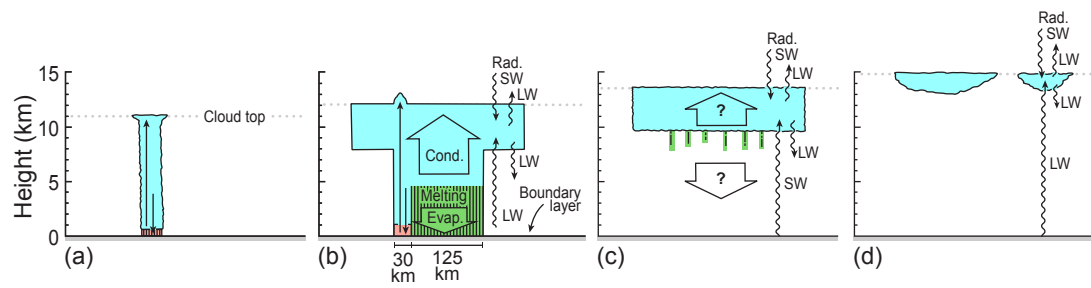
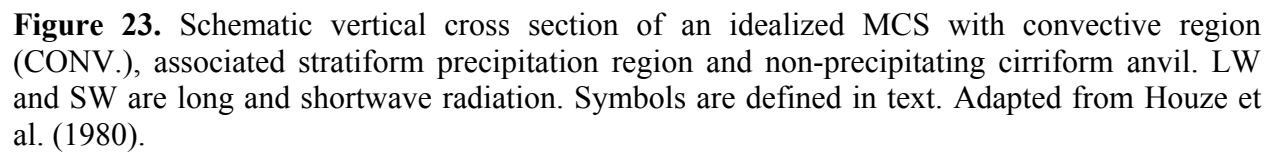


Figure 22. MCS in successive stages of development. Adapted from Houze (1982).



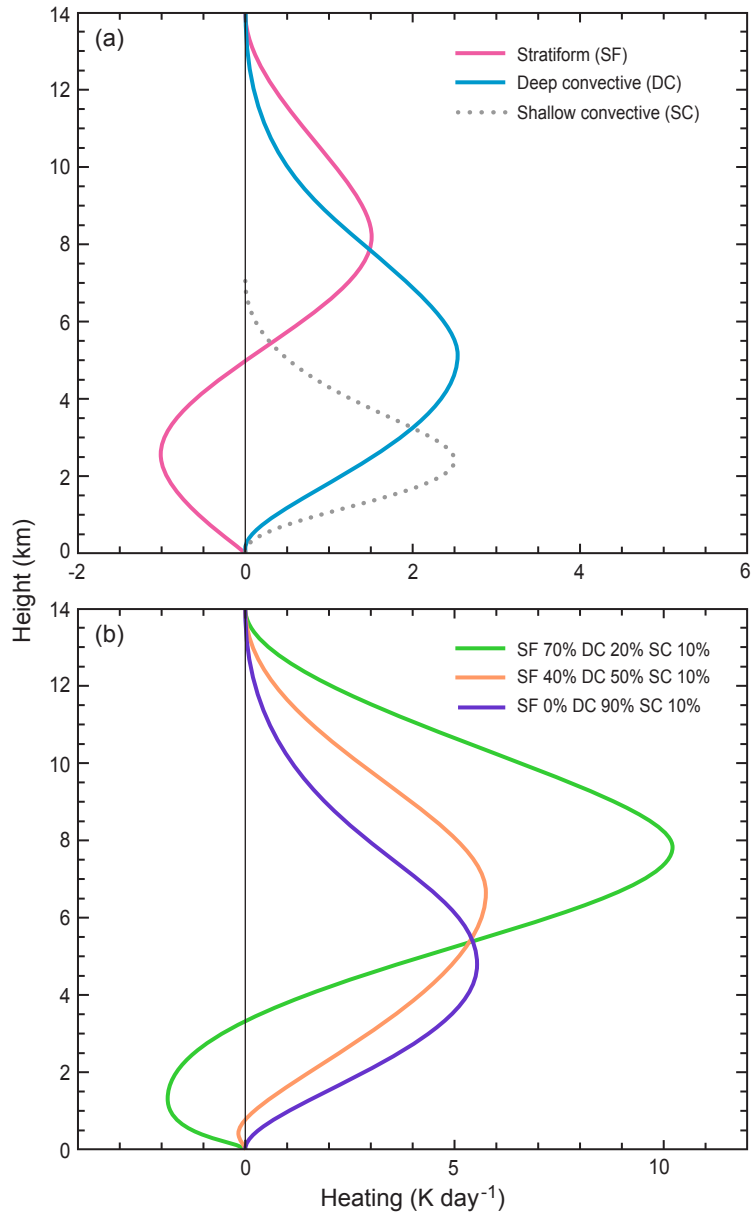


Figure 24. (a) Idealized stratiform, deep convective, and shallow convective latent heating profiles. The x -axis is meant to be nondimensional until a precipitation amount is specified. (b) Total latent heating profiles for 0%, 40%, and 70% stratiform rain fractions, assuming 3.5 m yr^{-1} rain accumulation. Adapted from Schumacher et al. (2004).

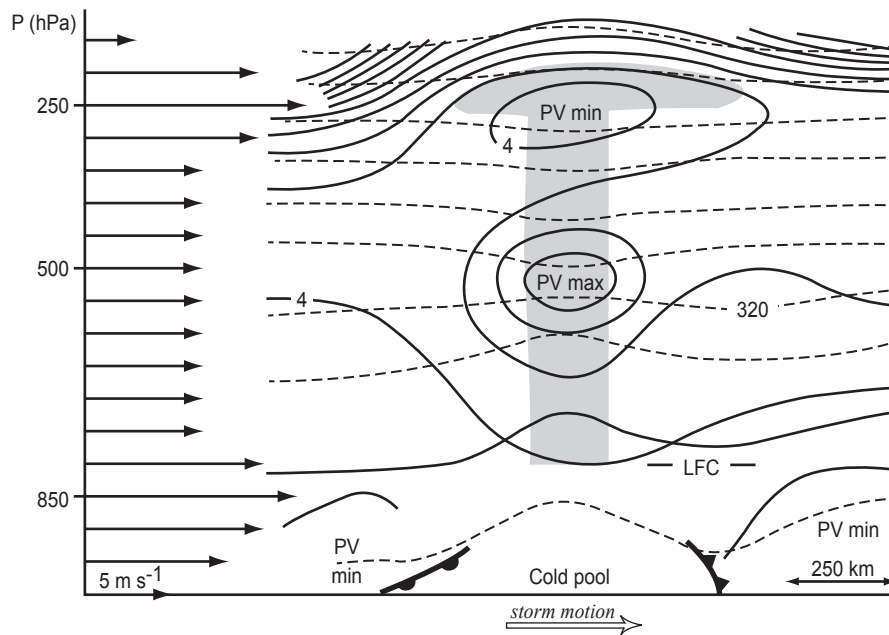


Figure 25. Conceptual diagram of the structure and development mechanism of a midlevel PV maximum associated with an MCS—based on a case study. Thin arrows along the ordinate indicate the vertical profile of the environmental wind. Frontal symbols indicate outflow boundaries. Dashed lines are potential temperature (5 K intervals) and solid lines are potential vorticity (PV, in intervals of $2 \times 10^{-7} \text{ m}^2 \text{ s}^{-1} \text{ K kg}^{-1}$). The system is propagating left to right at about $5\text{--}8 \text{ m s}^{-1}$ and is being overtaken by air of high equivalent potential temperature in the low-level jet. Air overtaking the vortex ascends isentropic surfaces, reaches its level of free convection (LFC), and thereby initiates deep convection. Shading indicates cloud. Adapted from Fritsch et al. (1994).

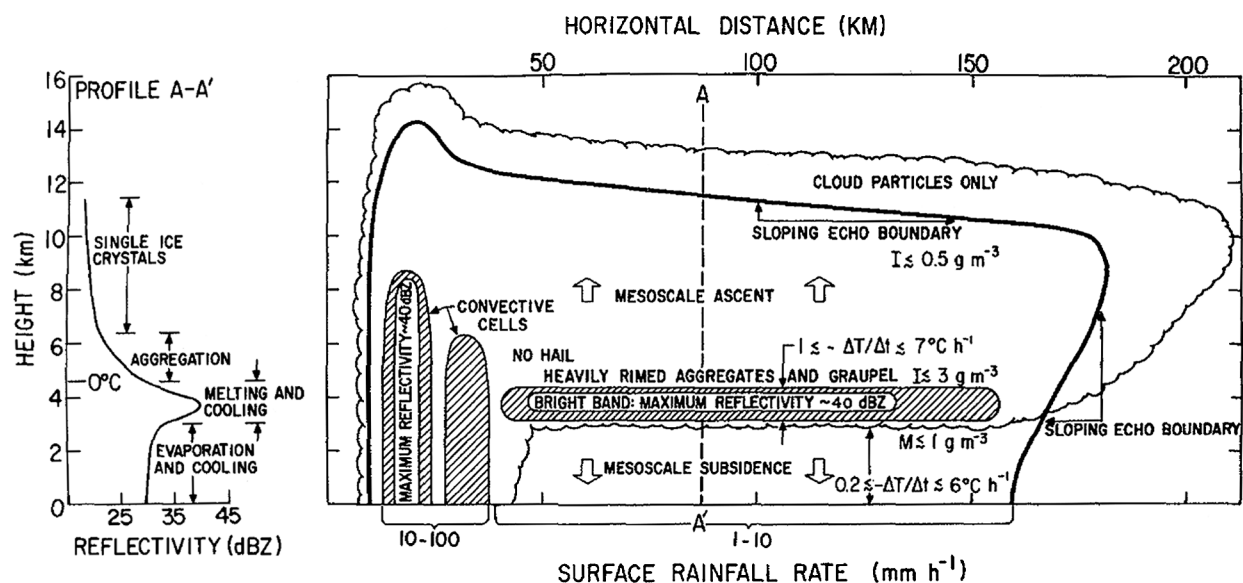


Figure 26. Schematic vertical cross section and vertical profile of radar reflectivity (along dashed line A-A' in the cross section) in horizontally uniform precipitation associated with an anvil cloud. The anvil cloud occurs to the rear of intense convective cells moving from right to left in the figure. The dark solid line is the contour of minimum detectable radar echo, lighter solid lines and shading indicate contours of higher reflectivity, and the scalloped line indicates the cloud boundary. From Leary and Houze (1979a).

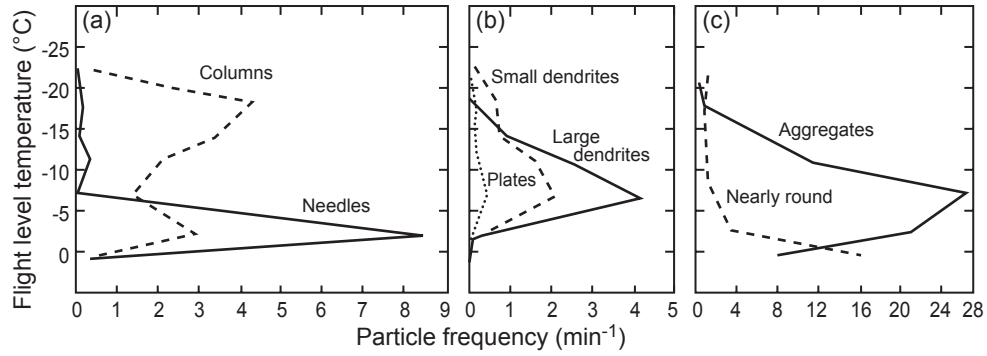


Figure 27. Ice-particle data obtained on aircraft flights through nimbostratus in tropical mesoscale convective systems over the Bay of Bengal. Plots show relative frequency of observation of ice particles of a particular type per minute of in-cloud flight time as a function of flight-level temperature. From Houze and Churchill (1987).

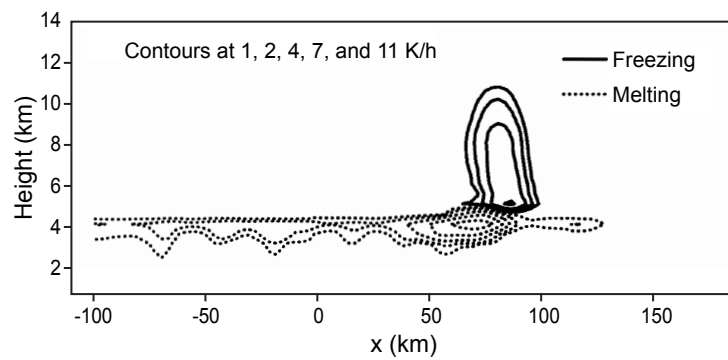


Figure 28. Melting and freezing rates retrieved from dual-Doppler radar data for a squall line MCS with its leading edge at about 100 km on the horizontal axis. From Braun and Houze (1995).

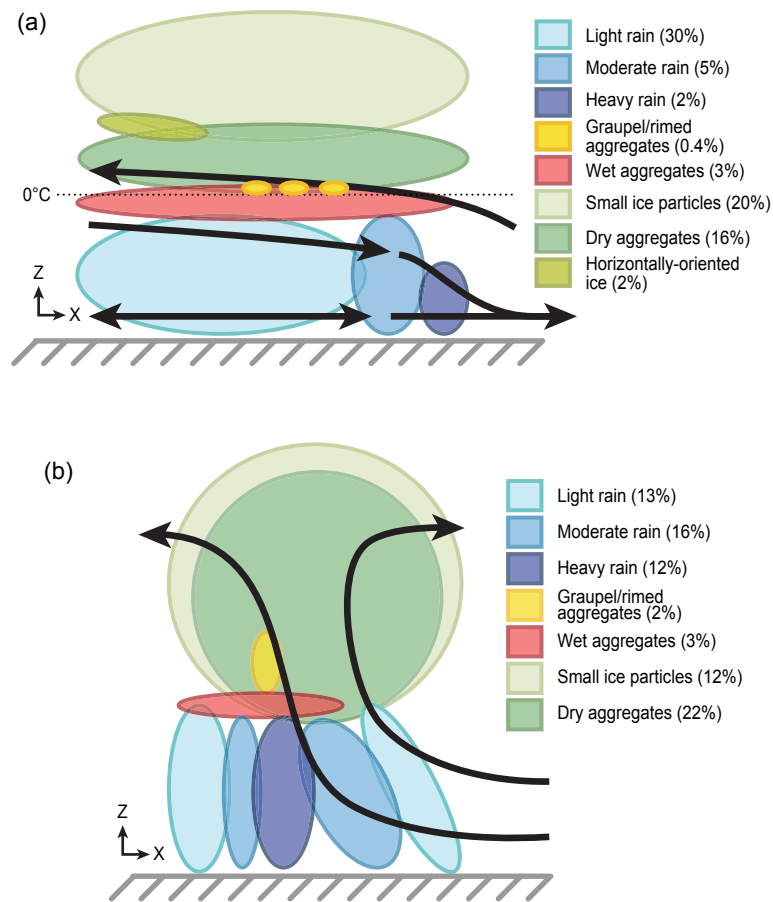


Figure 29. Schematics showing the location of hydrometeor types in an MCS over the tropical ocean relative to: (a) the layered airflow crossing the stratiform region; (b) the layer of upward slantwise motion entering in the convective region. The percentages in the color bar indicate the average areal coverage of each particle type. From Barnes and Houze (2014).

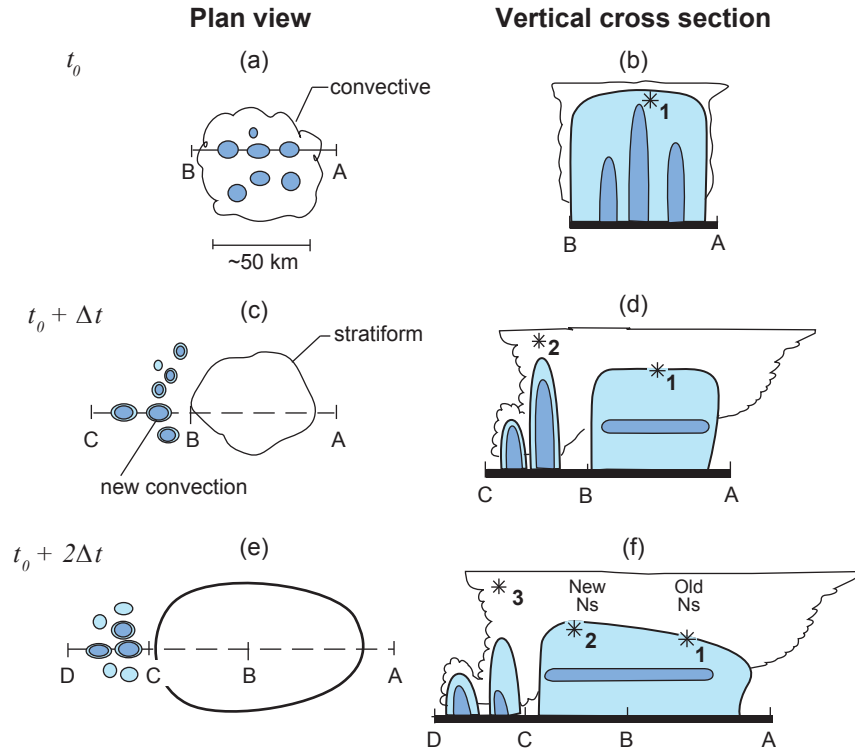


Figure 30. Conceptual model of the development of nimbostratus associated with deep convection. Panels (a), (c), and (e) show horizontal radar echo pattern at Earth's surface with two levels of intensity at three successive times, t_0 , $t_0 + \Delta t$, and $t_0 + 2\Delta t$. Panels (b), (d), and (f) show corresponding vertical cross sections. A sketch of the visible cloud boundary has been added to the vertical cross sections. Asterisks trace the fallout of three ice particles. From Houze (2014).

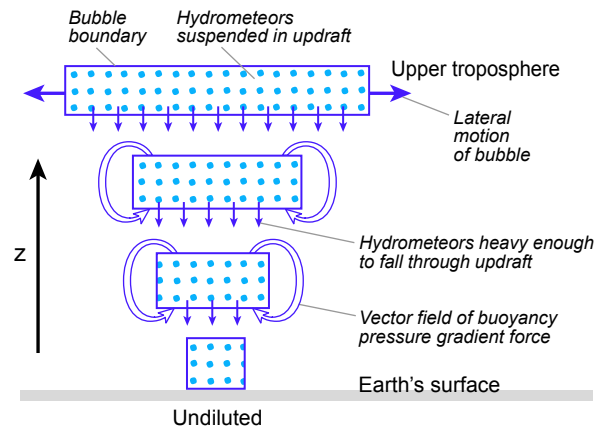


Figure 31. Conceptual model of a buoyant updraft element. Dots indicate hydrometeors suspended by updraft, downward pointing arrows indicate particles heavy enough to fall through updraft, and horizontal arrows indicate lateral spreading of bubble. Open arrows represent the vector field of the buoyancy pressure-gradient force. From Yuter and Houze (1995).

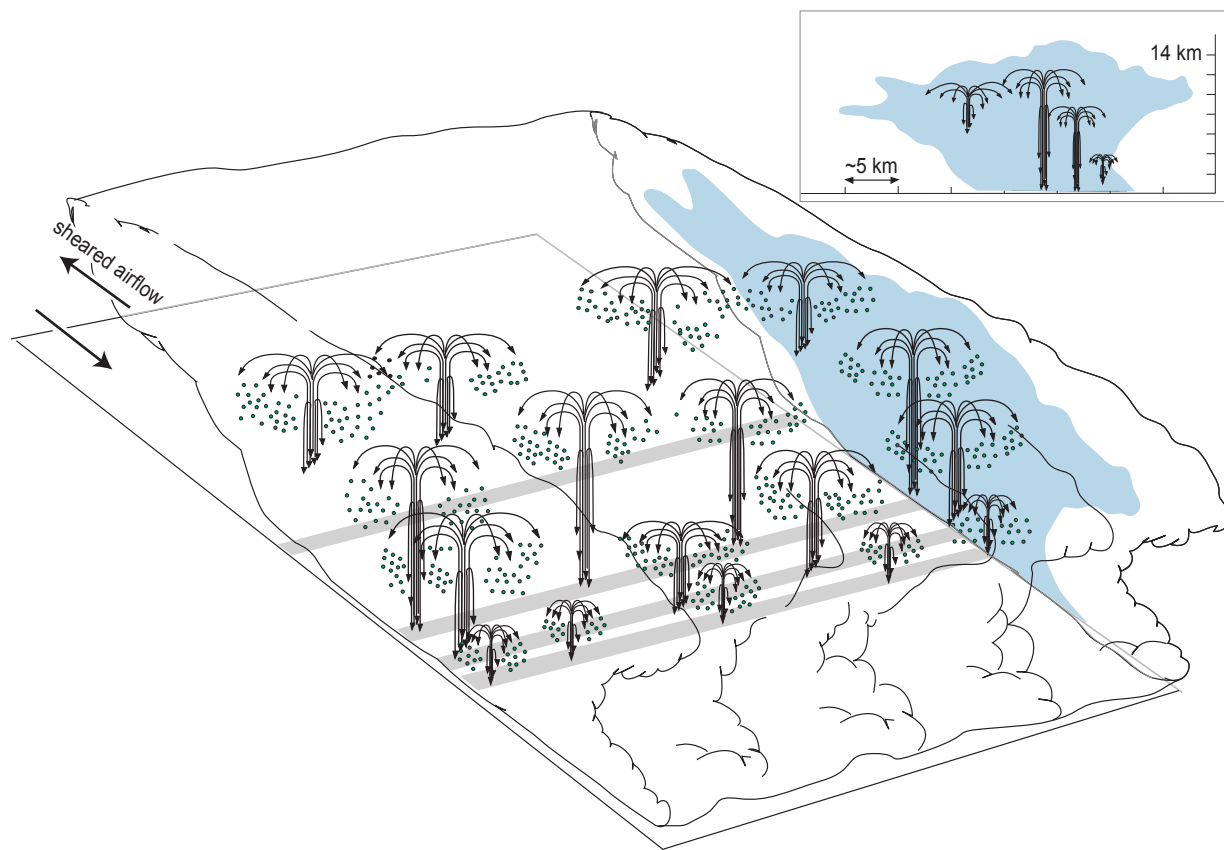


Figure 32. Conceptual model of an ensemble of particle fountains in a multicellular MCS. Shaded area represents radar reflectivity along a cross section normal to the convective region. Cloud boundary is indicated by the scalloped outline. Inset shows approximate scales and arrangement of the largest particle fountains relative to the radar echo. From Yuter and Houze (1995).

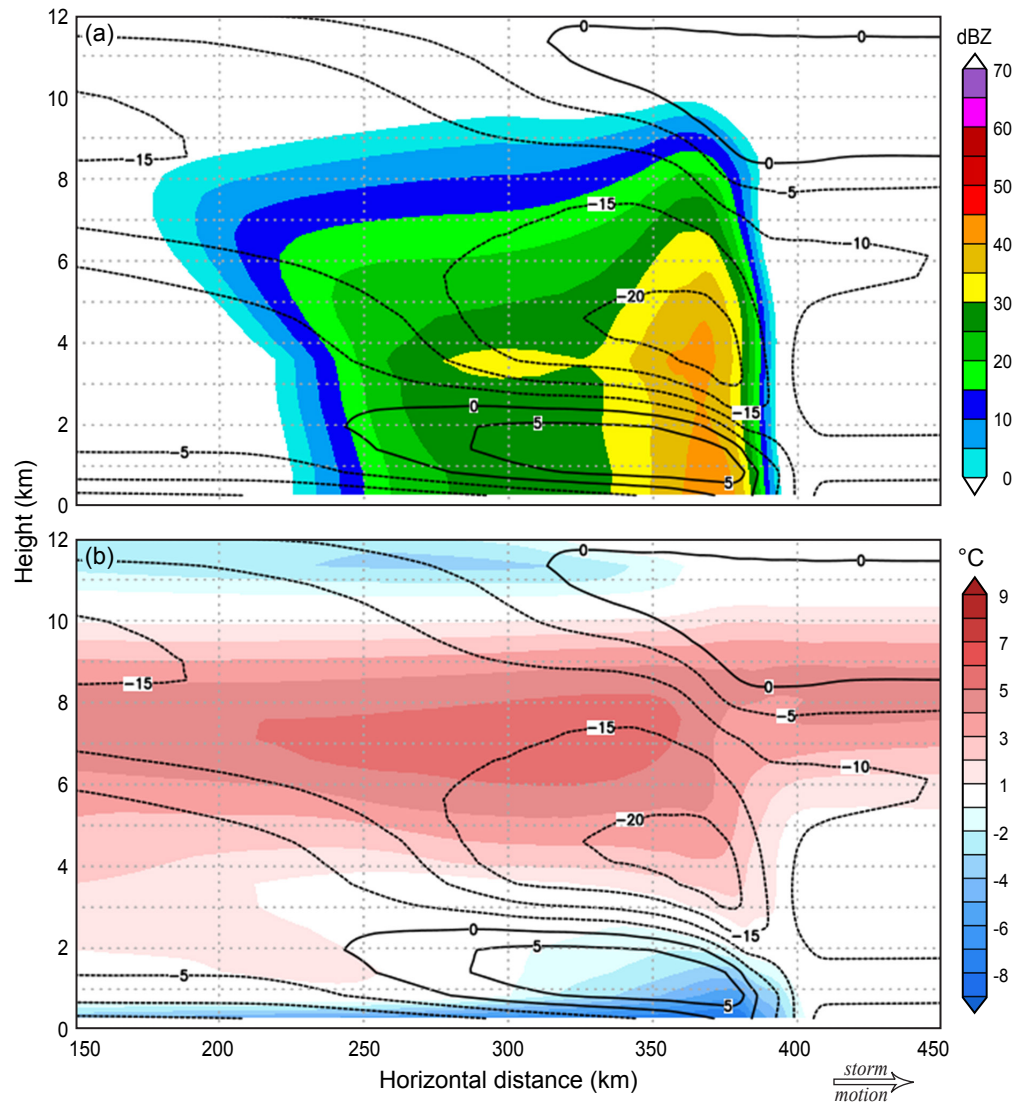


Figure 33. Results of a 3D simulation of a squall line MCS initiated with a line thermal (with random perturbations) and periodic along-line boundaries. The initial sounding is moderately sheared, and the microphysics scheme is that of Thompson et al. (2008) microphysics. The domain is 800 km (cross-line) by 80 km (along line) x 20 km deep. Horizontal resolution is 2 km. The images are averaged over a 3.5 h period (between simulation hours 4.5 and 8), and the x - z images are also averaged along the line. (a) Radar reflectivity (color) and wind component (m s^{-1} , negative right-to-left) in the plane of the cross section relative to the storm. (b) potential temperature (color) and same relative wind component as in (a). Provided to the author by Professor Robert Fovell.

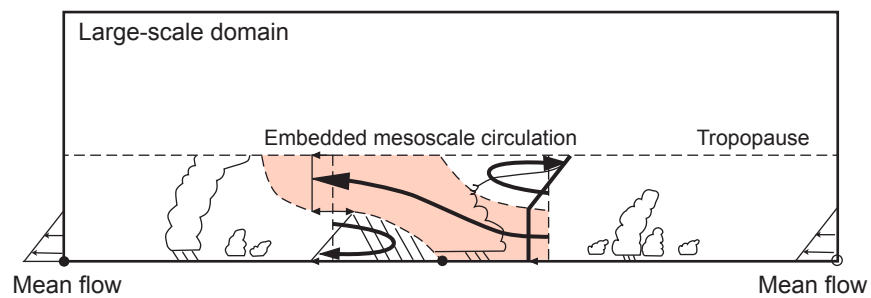


Figure 34. Schematic diagram showing the airflow relative to a two-dimensional, steady state mesoscale convective system in a large-scale environment of given wind shear and potentially instability. The streamlines are those required by conservation of mass, momentum, entropy, and vorticity. Adapted from Moncrieff (1992).

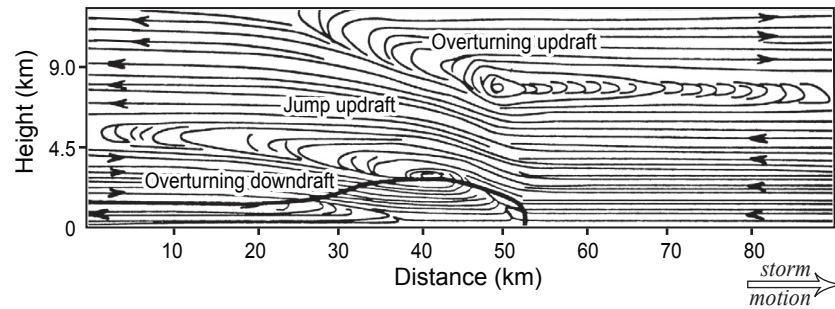


Figure 35. Time-averaged numerical model simulation of a squall line with trailing stratiform precipitation. (a) Simulated radar reflectivity (in intervals of 5 dBZ). (b) Streamlines of system relative airflow. (c) Equivalent potential temperature (intervals of 3 K). Bold solid contour outlines cold pool (region of negative potential temperature perturbation). Adapted from Fovell and Ogura (1988).

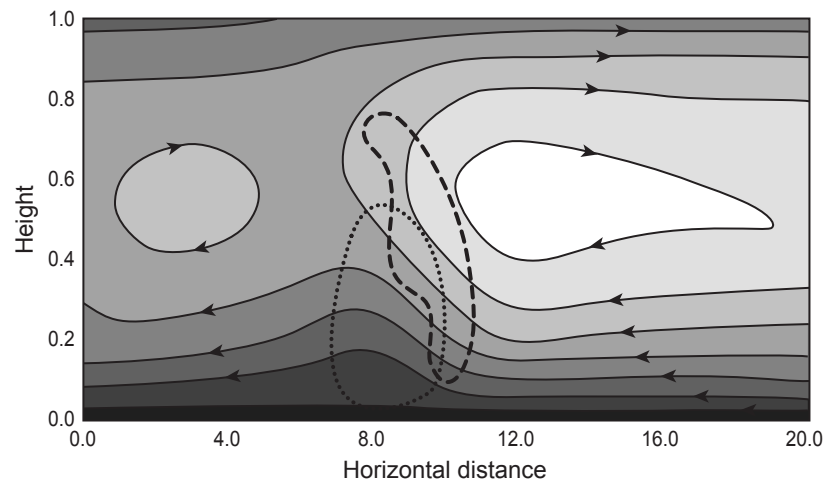


Figure 36. Streamlines for an MCS simulated by making a Wave-CISK assumption to represent the latent heating. The dashed and dotted contours outline regions of updraft and downdraft, respectively. From Raymond (1984).

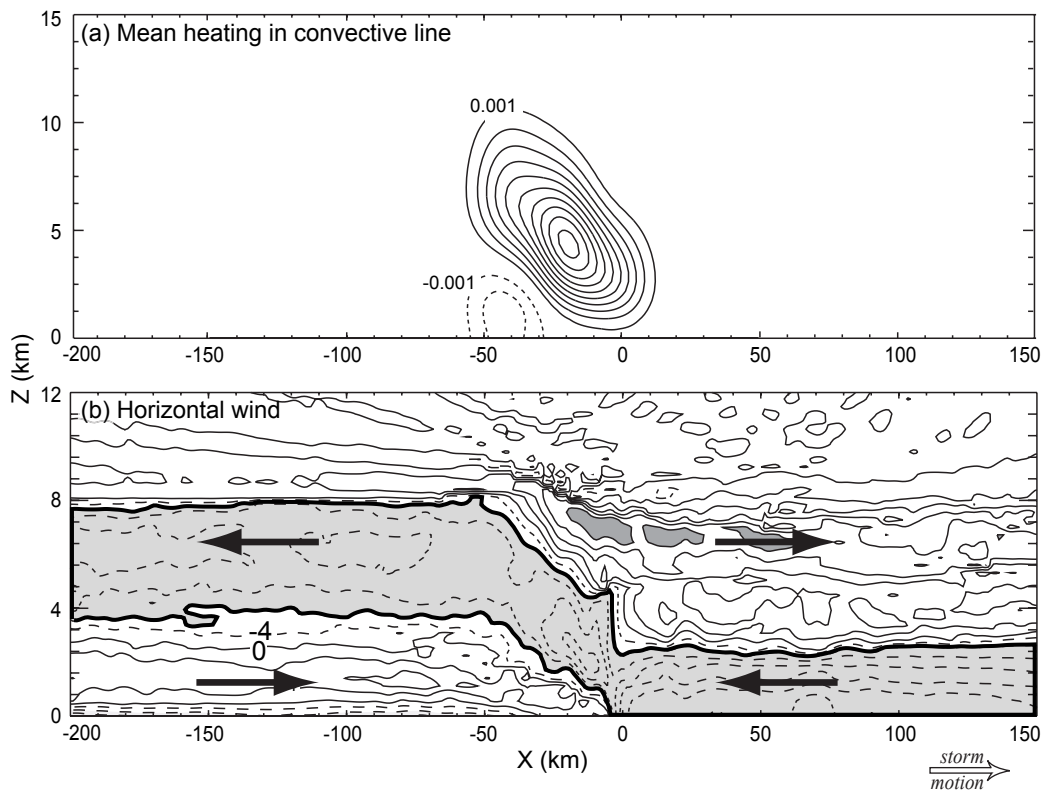


Figure 37. Two-dimensional model simulation results for a leading-line/trailing stratiform squall-line MCS. (a) Time-mean thermal forcing due to the leading convective line alone. Contour interval is 0.001 K s^{-1} . (b) Horizontal velocity at time = 6 h generated by the thermal forcing in (a). Horizontal velocity contours are at intervals of 4 m s^{-1} . Arrows indicate direction of the horizontal flow. Cold pool forward boundary is at $X=0$. Bold contour and shading emphasize layer inflow constituting the layer ascent of air originating ahead of the storm and rising through it. Adapted from Pandya and Durran (1996).

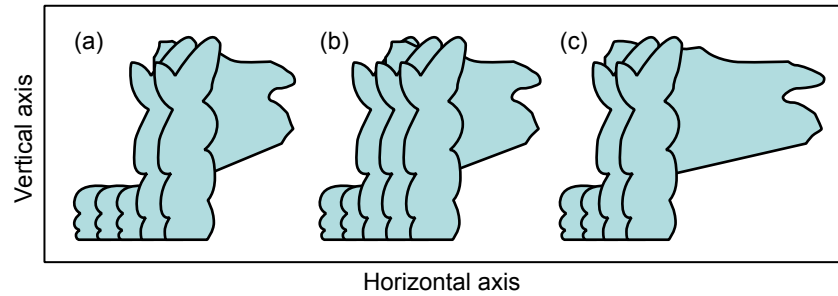


Figure 38. Depiction of three of cloud populations, made up of shallow convection, deep convection, and stratiform elements. In (a) the fraction of shallow convection is highest in the left-hand population. In (b) the fraction of deep convective elements is highest. In (c) the fraction of stratiform elements is greatest. Adapted from Mapes et al. (2006).

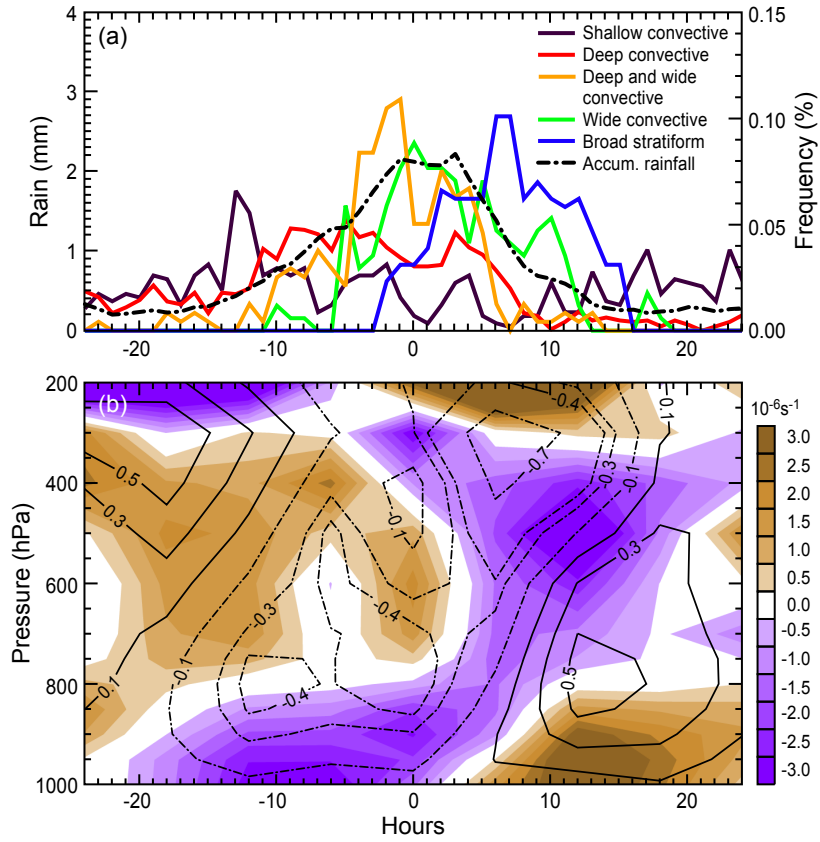


Figure 39. (a) Composite time–height sections of divergence (shading, 10^{-6} s^{-1}) and anomalies of vertical velocity (contours, $10^{-3} \text{ hPa s}^{-1}$; solid lines indicate positive values) calculated from reanalysis data composited around the time of maximum in rain accumulation seen by the National Center for Atmospheric Research S-PolKa radar located on Addu Atoll on the equator in the central Indian Ocean during DYNAMO. (b) Composite of the frequency of occurrence of different types (color-coded) of radar echo structure before (negative time) and after (positive time). The right y-axis is for the colored-coded frequency curves. The rainfall accumulation amounts shown on the left y-axis is the composite computed by centering each of the 11 rain episodes of DYNAMO on the time of the maximum of its running-mean curve. From Zuluaga and Houze (2013).

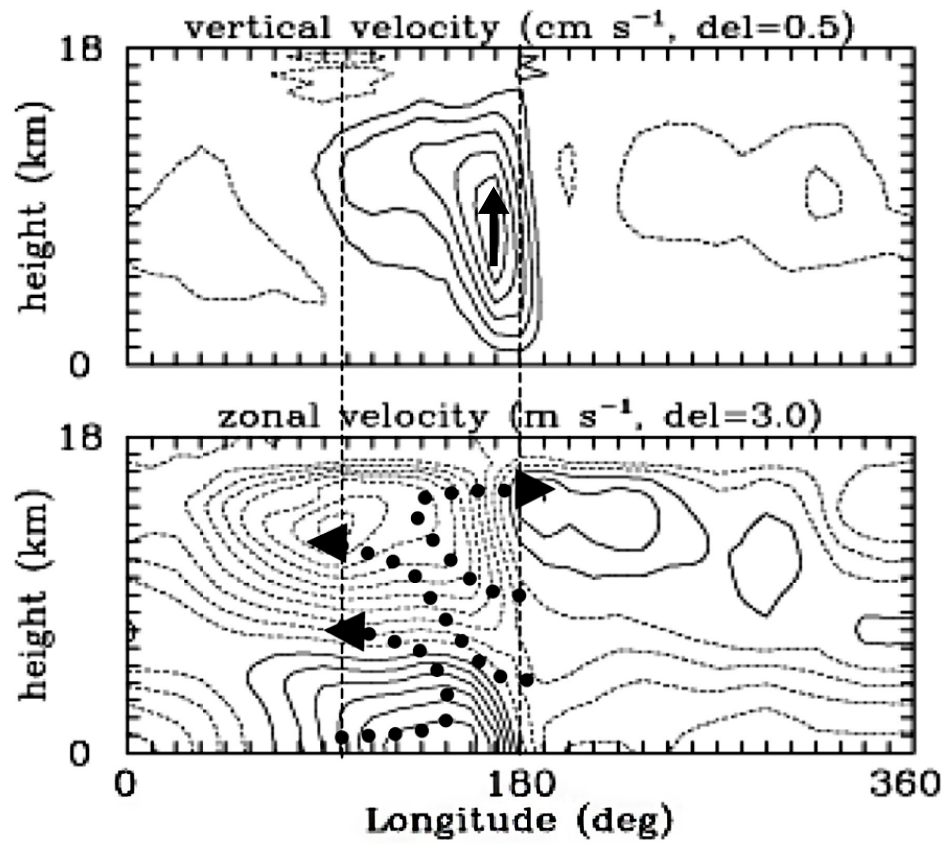


Figure 40. Vertical structure of the MJO system. Primary airflow branches are sketched. All quantities are 20-day averages and relative to the traveling system. From Moncrieff (2004).

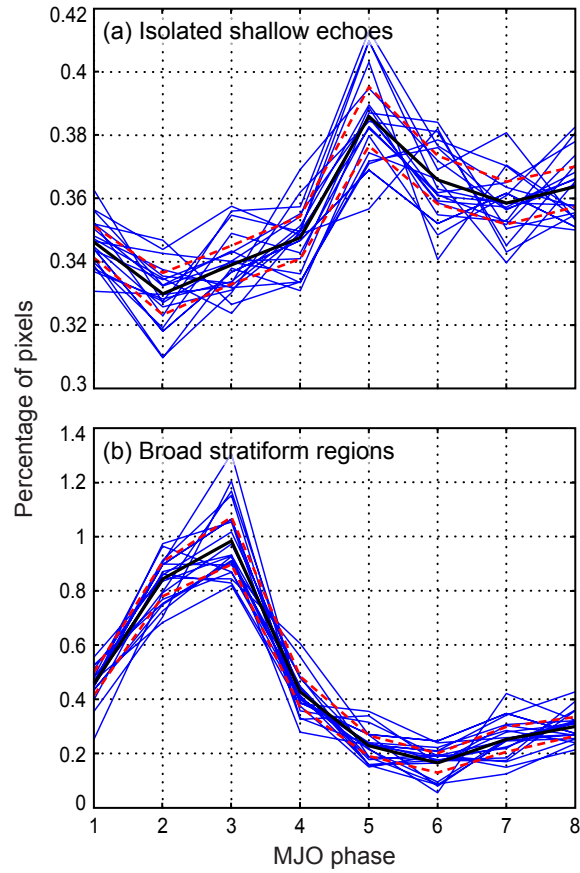


Figure 41. Frequency of occurrence of different types of radar echoes seen by the TRMM Precipitation Radar over the Central Indian Ocean: Total frequency of (a) isolated shallow echoes and (b) broad stratiform regions. The frequency is defined as the number of TRMM Precipitation Radar pixels in the central Indian Ocean that contain an isolated shallow echo normalized by the total number of TRMM PR pixels detected in the central Indian Ocean, which include both echo-covered and echo-free pixels. The frequency is reported as a percent. The blue lines show the frequency-phase series for the 20 realizations, the black line is the mean of the realizations, and the dashed red lines are 99% confidence intervals of the mean and are calculated using the Student's t-statistic. In the central Indian Ocean the indices for the active and suppressed phases of the MJO are 3 and 6, respectively. From Barnes and Houze (2013).

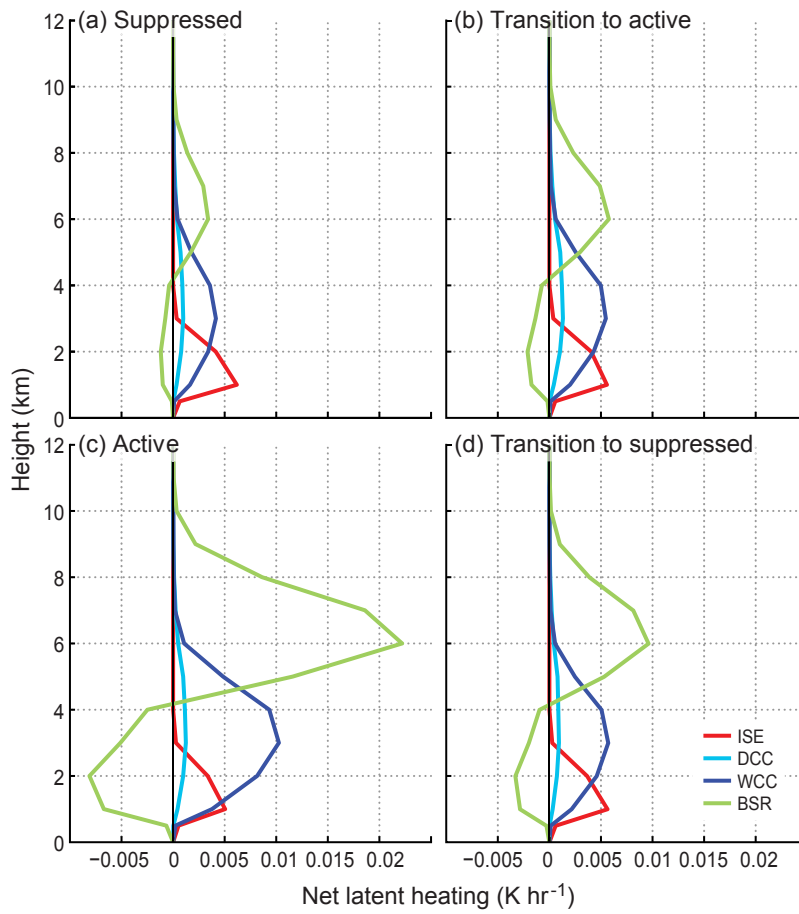


Figure 42. Net heating from isolated shallow convective elements (red), deep convective cores (cyan), wide convective cores (dark blue), and broad stratiform regions (green) as seen by the TRMM radar during different stages of the MJO over the central Indian Ocean. Net heating is normalized by the total number of pixels sensed by the TRMM Precipitation Radar. From Barnes et al. (2015).

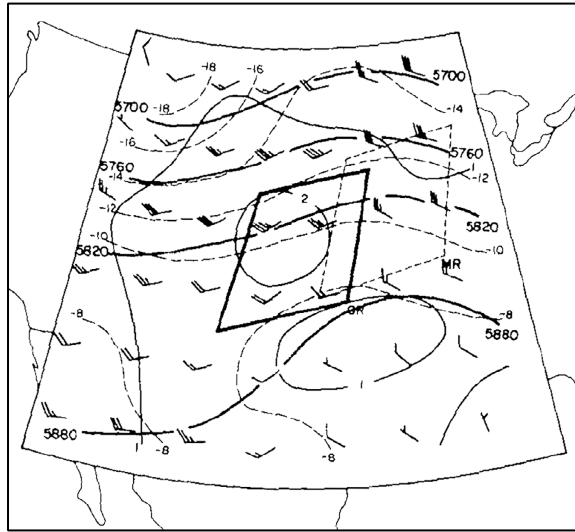


Figure 43. Analysis of the 500 hPa level prior to MCC development. Heights (m) are heavy solid contours, isotherms ($^{\circ}\text{C}$) are dashed, and mixing ratio (g kg^{-1}) is indicated by light solid contours. Winds (full barb = 5 m s^{-1}) are plotted at every other grid point and dark arrow shows axis of maximum winds. The crosshatched region indicates terrain elevations above the 500 hPa level. From Maddox (1983).

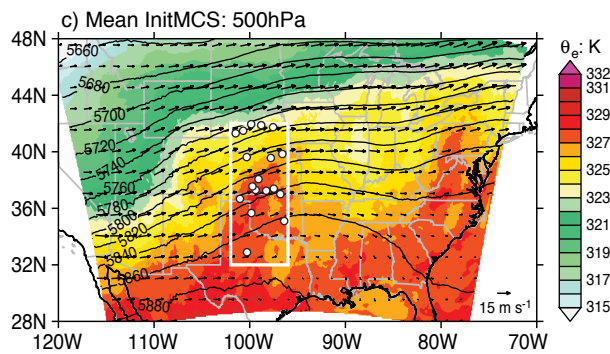


Figure 44. Results of a cloud-resolving model run for two warm seasons. The fields are composites of the mean geopotential height (in meters, contour lines), wind (vector), and equivalent potential temperature (in K, color filled) at 500 hPa at 0-3 h before MCS initiation. The dots indicate the locations of MCS initiation. The box is the region in which MCS initiation was sought. From Yang et al. (2017).

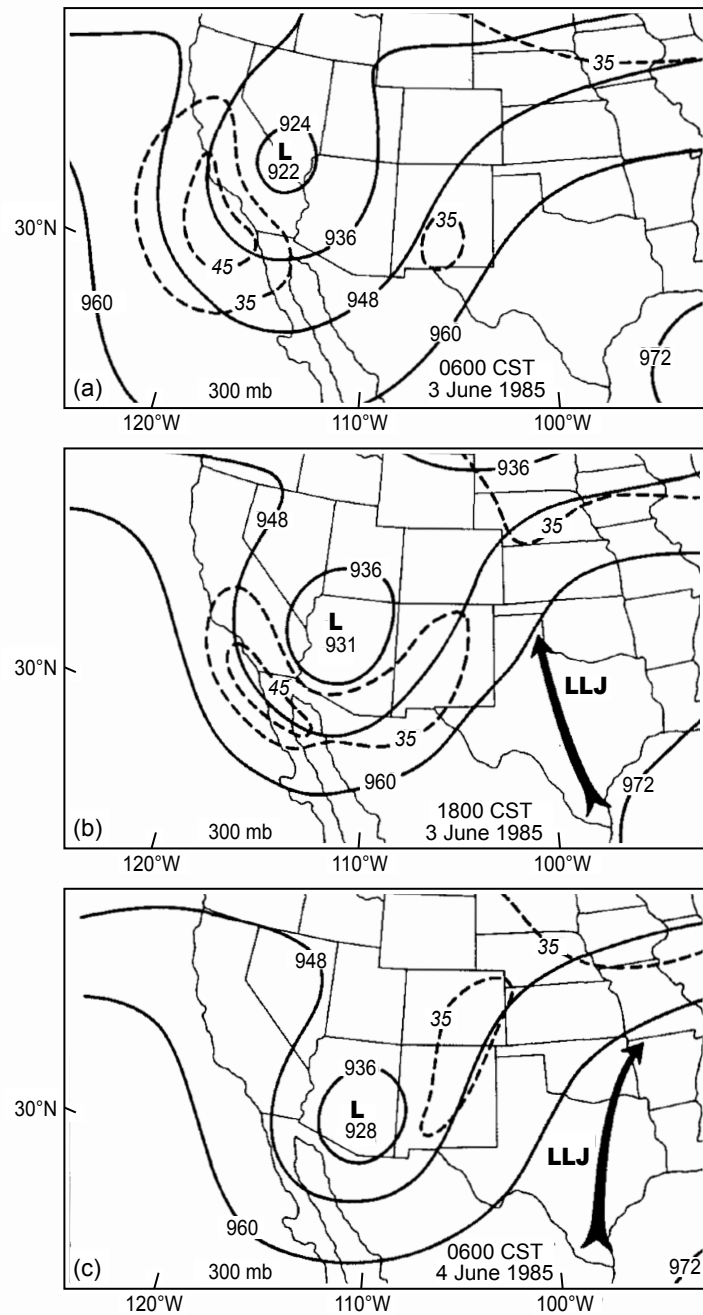


Figure 45. Sequence of 300-hPa analyses with geopotential height contours (solid) analyzed in 12 dam intervals with the 35 and 45 m s⁻¹ isotachs (dashed) at (a) 0600 CST 3 June, (b) 1800 CST 3 June, and (c) 0600 CST 4 June 1985. The bold arrow schematically illustrates the approximate streamline of the low-level jet in (b) and (c). From Trier and Parsons (1993).

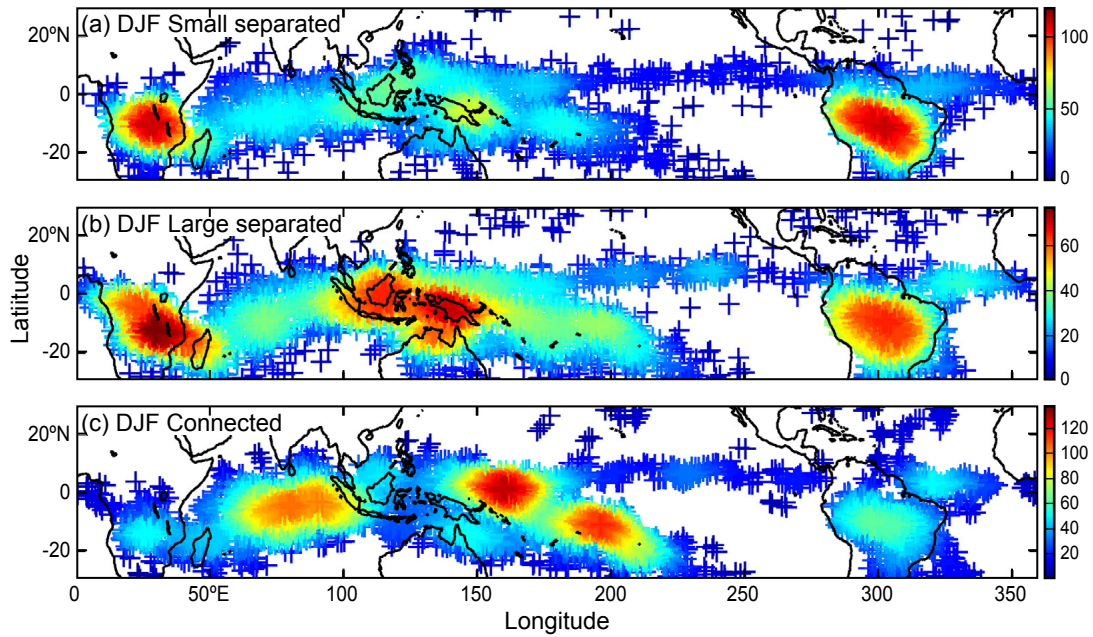


Figure 46. Frequency of occurrence of (a) “small separated,” (b) “large separated,” and (c) “connected” tropical MCSs during the months December, January, and February, as shown by satellite infrared and passive microwave sensors on the NASA Aqua satellite. Small and large MCSs have anvil plus raining area $<10^4$ km² and $>2.25 \times 10^4$ km², respectively. Connected MCSs have separate infrared cloudtop signatures but share a rain area. From Yuan and Houze (2010).

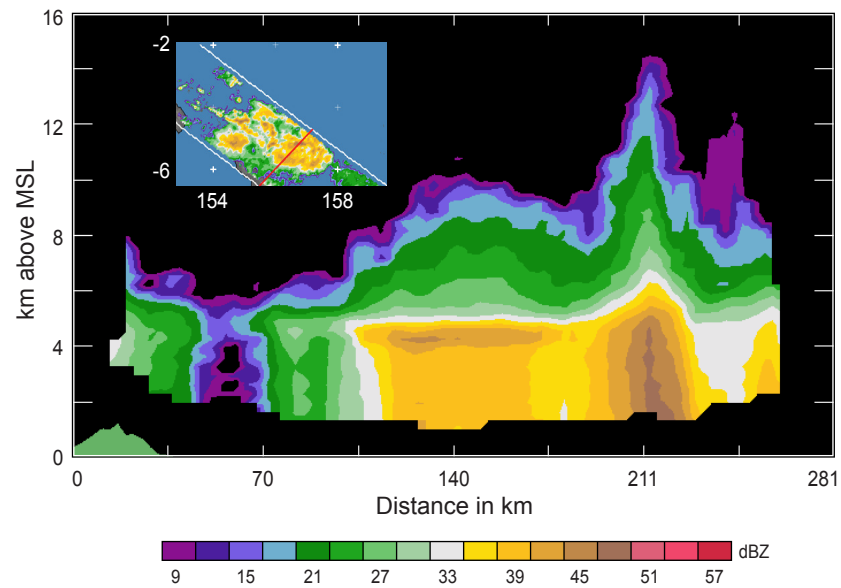


Figure 47. Vertical cross section of radar reflectivity in dBZ seen by the TRMM PR along the red line superimposed on the plan view at 4 shown in the inset. The vertical cross section runs from left to right along the red line. Latitudes -2 to -6 are in the Southern Hemisphere. Longitudes are in the Eastern Hemisphere. From Houze et al. (2015).

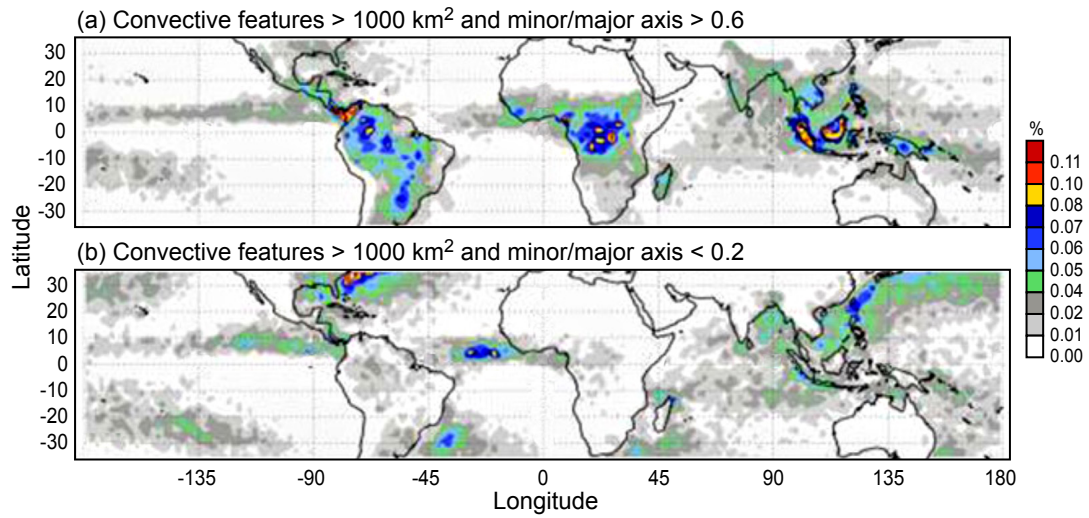


Figure 48. (a) Distribution of the population of large convective echoes (area $> 1000 \text{ km}^2$) and with a near-circular shape (minor/major axis ratio > 0.6). (b) Distribution of the population of large convective lines (convective echoes with minor/major axis ratio < 0.2 and area $> 1000 \text{ km}^2$). From Liu and Zipser (2013).

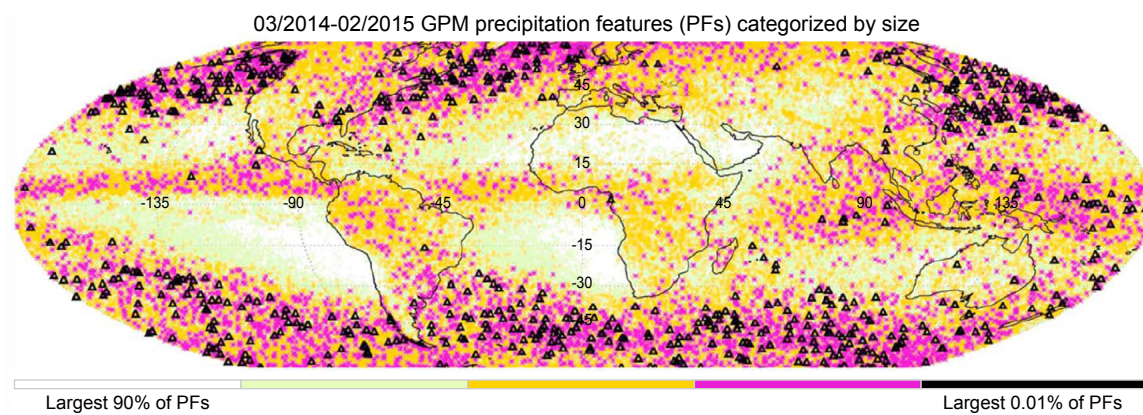


Figure 49. Locations of precipitation features (PFs) seen by the GPM satellite and categorized by size. From Liu and Zipser (2015).

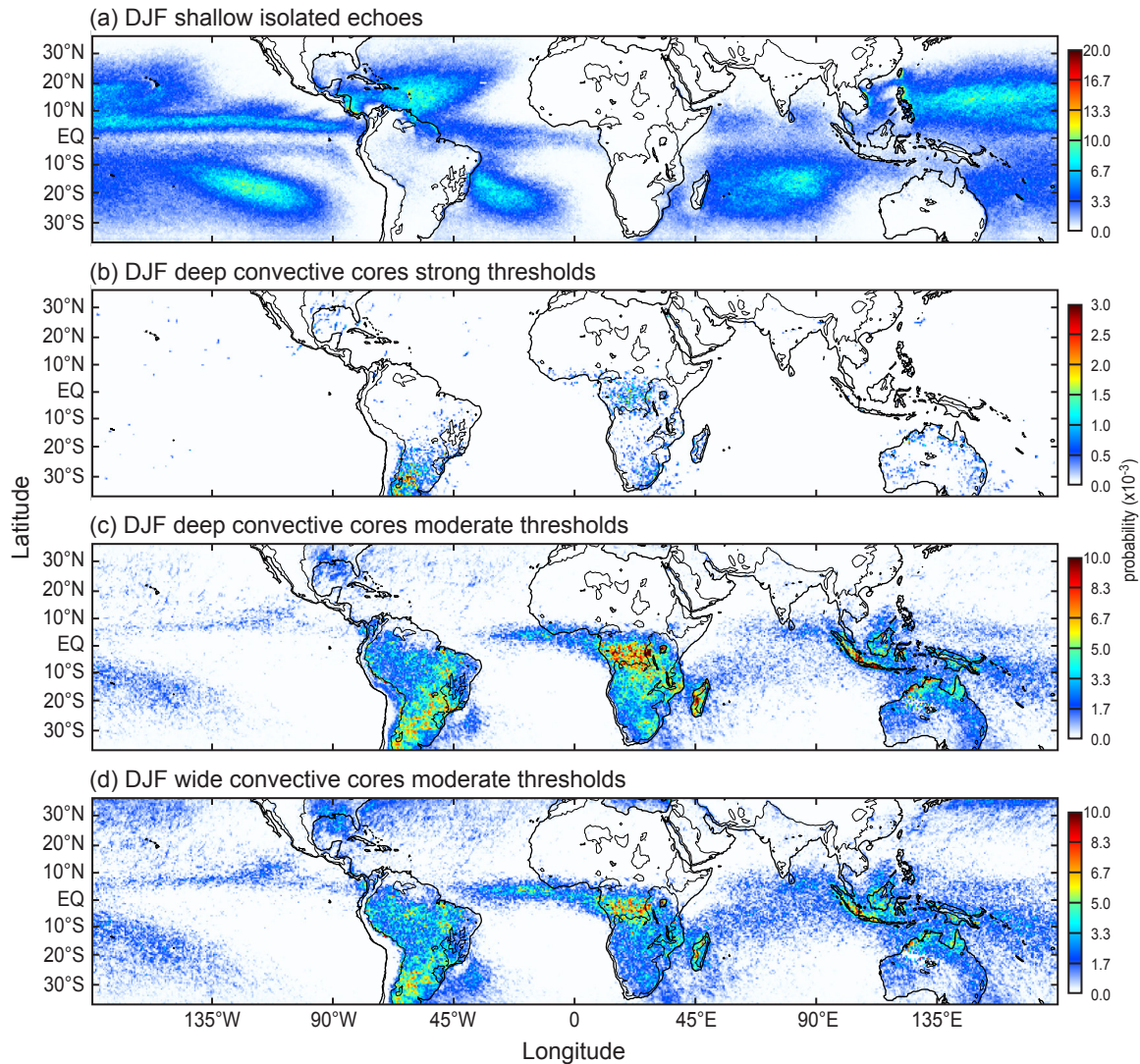


Figure 50. Geographical distribution of the probability of finding various categories of echoes seen by the TRMM Precipitation Radar during December-February: (a) isolated shallow echoes; (b) strong threshold deep convective cores; (c) moderate threshold deep convective cores; and (d) moderate threshold wide convective cores. The black contour inside the continental regions represents the 700 m elevation. The probability is on a scale of 0 to 1 and is computed as the number of pixels identified as belonging to an echo category divided by the total number of pixels sampled by the TRMM PR over the given time period within a 0.5° by 0.5° grid element. From Houze et al. (2015).

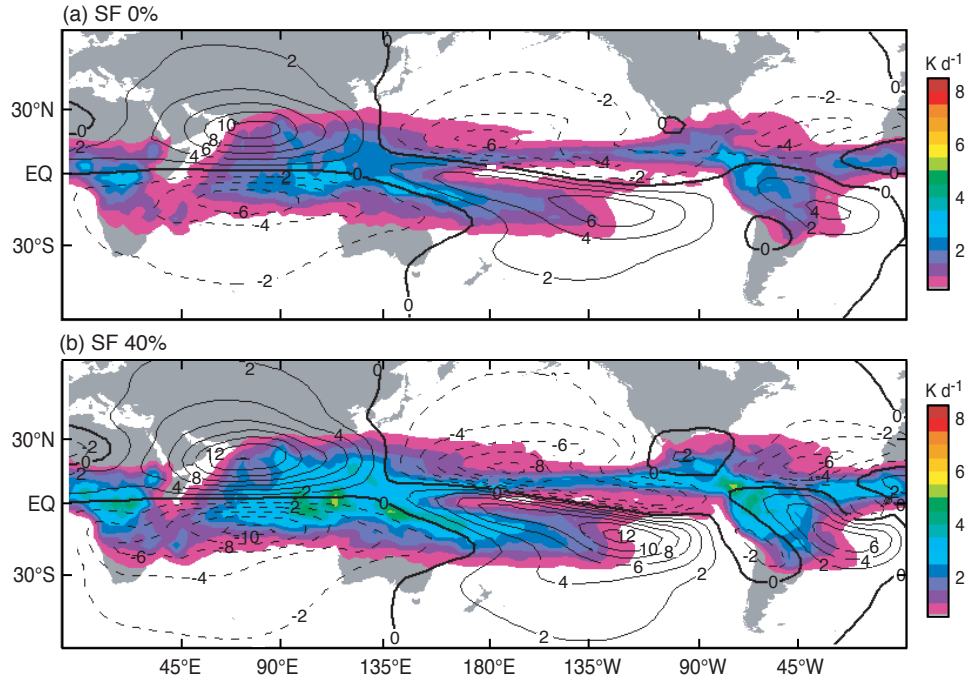


Figure 51. The 400-hPa latent heating (shaded) and the resulting 250-hPa streamfunction anomalies (contours) computed from a simple climate model for the resting basic state and latent heating derived from the PR annually averaged precipitation field and geographically uniform stratiform rain fractions of (a) 0% and (b) 40%. The streamfunction contour interval is $2 \times 10^6 \text{ m}^2 \text{ s}^{-1}$. Negative contours are dashed. From Schumacher et al. (2004).

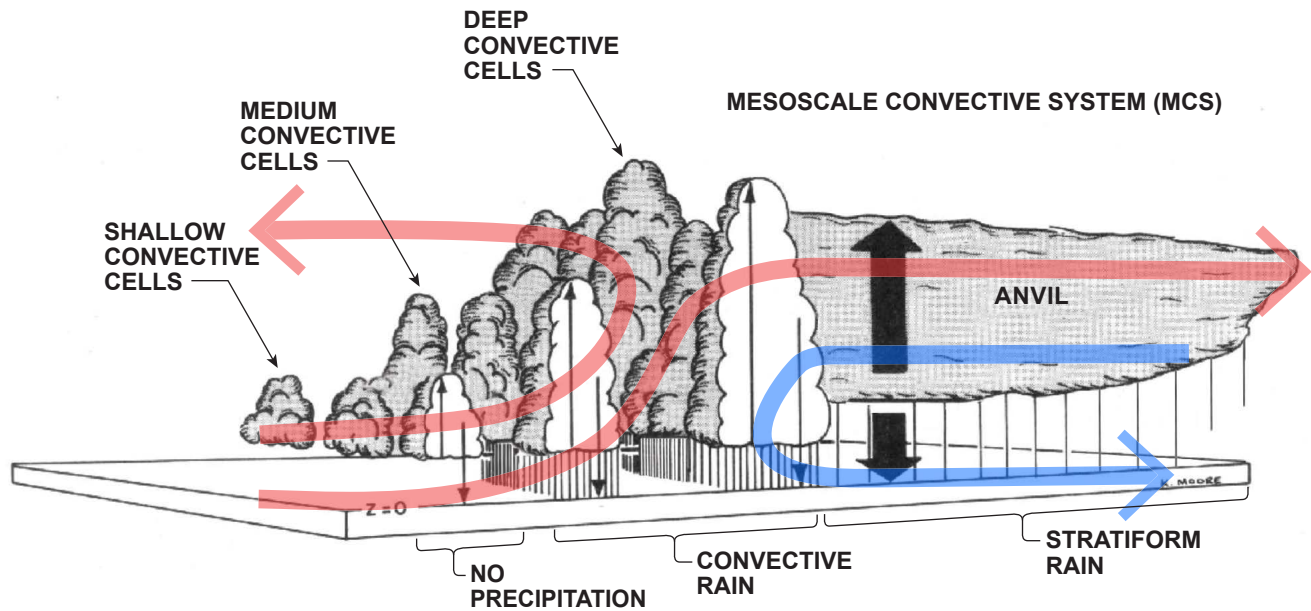


Figure 52. Overlay of a convective cloud population and superimposed layered overturning. Adapted from Houze et al. (1980) and Moncrieff et al. (2017).

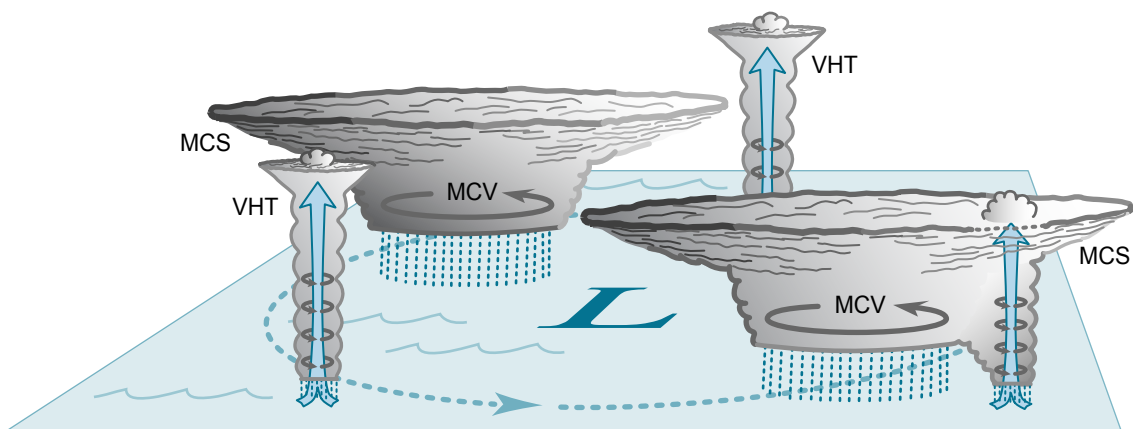
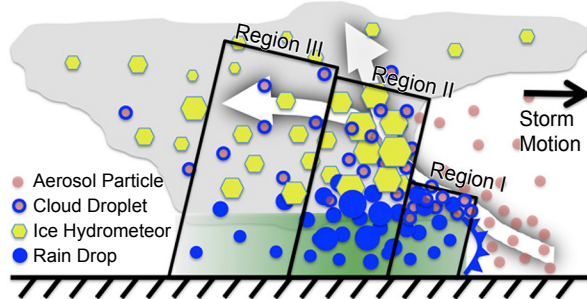


Figure 53. Idealized distribution of convection in a low-pressure system (L) in the process of developing into a tropical cyclone. The distribution contains isolated convective cells and MCSs in various stages of development. The convective updrafts stretch the environmental vorticity and advect positive vorticity upward. These updrafts with concentrated positive vorticity are referred to as vortical hot towers (VHTs). The larger and longer-lived MCS contain VHTs in early MCS stages. The MCSs also contain stratiform regions with midlevel mesoscale convective vortex (MCVs). Both the VHTs and MCVs are thought to contribute to cyclogenesis. From Houze et al. (2009).

(a) High concentrations of lower tropospheric aerosol particles



(b) High concentrations of middle tropospheric aerosol particles

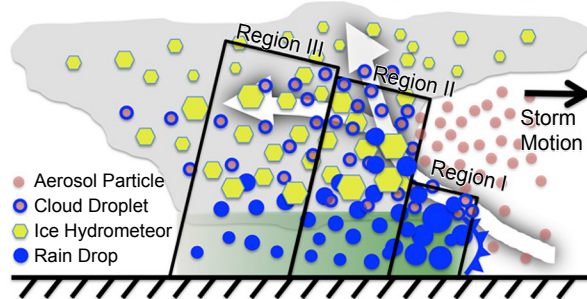


Figure 54. Schematic of an MCS under high concentrations of (a) lower-tropospheric and (b) midtropospheric aerosol particles. The gray area represents cloudy regions, while the green shading represents rainfall with darker shading depicting heavier rain. The blue frontal symbols represent the leading cold pool boundary, and the white arrows represent the primary front-to-rear ascending flow. The three boxes represent three different precipitation regions of the MCS where different microphysical processes are governing the precipitation. Particle types are specified in the legends, with the amounts and sizes of hydrometeors representative of the relative concentrations and mean diameters of particles within those regions. From Marinescu et al. (2017).