8

Mesoscale processes

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8.1 INTRODUCTION

Within Asia and other monsoon regions of the World, most significant weather events are localized or *mesoscale* in nature. The term mesoscale generally refers to horizontal scales between ten and several hundreds of kilometers, which lie between the scale height of the atmosphere and the Rossby radius of deformation (Ooyama, 1982). The latter scale can become quite large in equatorial regions owing to the weakness in background rotation, hence mesoscale processes in the tropics can occur over a broad range of horizontal scales. Although local weather is influenced by processes ranging from the largest scales to the smallest, those on the mesoscale have the most direct impact.

In the Asian monsoon region, there are a multitude of mesoscale processes that influence the weather. Convection is arguably the most important, contributing through latent heat release to the energetics of the large-scale monsoon circulation. Its far-reaching effects impact short-term weather, the diurnal cycle, as well as intraseasonal, seasonal, and interannual variability of the monsoon. Convection responds to and modulates its environment through a wide range of processes that can be classified as local, advective, or dynamical (Johnson and Mapes, 2001). These processes are associated with convective preconditioning and triggering, as well as the feedback of convection onto its environment and larger scales of motion.

In addition to convection, a wide range of other mesoscale processes influence the Asian monsoon. Notable among these are topographically forced local circulations (including flow blocking, sea and land breezes, and mountain and valley circulations), jets, surface-atmosphere interactions, gravity currents and gravity waves, coastally trapped disturbances, and mesoscale instabilities. Most of these phenomena are significantly modulated by the diurnal cycle. In this chapter we review mesoscale atmospheric processes and provide some examples that occur in

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the Asian monsoon region, many of which are relevant to other monsoon regions of the World.

8.2 CONVECTION

Atmospheric convection is one of nature's most complex and multifaceted phenomena (e.g., Ludlam, 1980; Cotton and Anthes, 1991; Emanuel, 1993; and Houze, 1993). Considerable insight into tropical and monsoon convection has been gained from field experiments such as the GATE (GARP Atlantic Tropical Experiment), MONEX (Monsoon Experiment), and TOGA/COARE (Tropical Ocean–Global Atmosphere/Coupled Ocean–Atmosphere Response Experiment) (Houze and Betts, 1981; Johnson and Houze, 1987; Godfrey *et al.*, 1998); however, many aspects of convection are still not well understood.

A map of the global distribution of precipitation based on data from the Tropical Rainfall Measuring Mission (TRMM) is presented in Figure 8.1 (color section). Much of the World's heaviest rainfall occurs in the regions of the Asian–Australian monsoon. The largest annual totals occur in proximity to coastlines, suggesting possible roles of sea and land breezes and topographic effects in the precipitation mechanisms. To understand this distribution or other aspects of monsoon rainfall, we need to first examine the precipitation characteristics and structural properties of moist convection.

8.2.1 Distribution, organization, and structure of tropical convection

There is growing evidence from field experiments over the past three decades that convective systems in the various monsoon and tropical regions of the World bear a close resemblance to each other. In particular, deep convection tends to organize on the mesoscale and undergo an evolution characterized by a dominance of convective precipitation (localized heavy rainfall) early in the life cycle followed by an upscale growth and development of stratiform precipitation (lighter rainfall) on a timescale of 2–4 hours and longer (Zipser, 1977; Houze, 1977; Leary and Houze, 1979). The stratiform precipitation is partly a result of the transfer of hydrometeors from the convective region and partly a result of *in situ* condensation and deposition in the stratiform region. The net result is a mesoscale convective system or MCS, defined by Houze (1993) as a cloud system that occurs in connection with a cluster of showers and produces a contiguous precipitation area $\sim$100 km or more in horizontal scale in at least one direction.

Global climatologies of MCSs in the monsoon regions were first carried out using satellite studies of mesoscale convective complexes or MCCs (Maddox, 1980), the largest and longest lived of MCS populations. Laing and Fritsch (1997) present a map of MCC locations using satellite data (Figure 8.2). They found that MCCs are (1) mostly continental, (2) tend to occur in gradient zones between OLR maxima and minima (i.e., they are normally not in the most frequently raining areas), and (3) tend to occur in the lee (relative to the prevailing mid-level flow) of

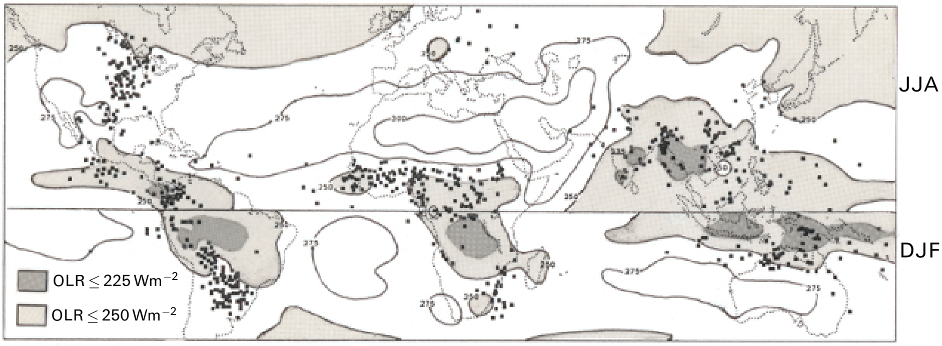

Figure 8.2. MCC locations based on 1980s satellite data for JJA in the northern hemisphere and DJF in the southern hemisphere. Outgoing long-wave radiation (OLR) values are shaded. From Laing and Fritsch (1997) Royal Meteorological Society.

elevated terrain. Figure 8.2 shows that MCCs are common not only over the region of the Asian–Australian monsoon (China, India, Bangladesh, and northern Australia) but also over South America to the lee of the Andes and West Africa. These findings have recently been confirmed and extended by TRMM microwave measurements (Nesbitt *et al.*, 2000).

The structure and dynamics of MCSs have been the subject of intensive study for the past 30 years (Houze, 1993). A recent investigation of the evolution of nearly 100 MCSs over the central USA has revealed new characteristics of such systems (Parker and Johnson, 2000). Three main patterns of MCS organization have been identified (Figure 8.3). The three modes are convective lines with trailing (TS), leading (LS), and parallel (PS) stratiform precipitation. TS systems were the most common, accounting for ~60% of the cases, with the LS and PS each accounting for about 20%. TS systems have received considerable attention (e.g., Houze *et al.*, 1990), but the occurrence of LS and PS systems is not insignificant, and there is evidence they are important in monsoon regions. For example, MCSs in the Baiu front appear to have these precipitation structures (Ninomiya and Muarkami, 1987). Wang (2004) recently found LS organization of MCSs to be commonplace over the northern South China Sea during the 1998 South China Sea Monsoon Experiment (SCSMEX). TS systems normally propagate rapidly ($\sim$10–15 m s^{-1}) and, as such, produce brief, heavy rainfall but usually not flash floods. LS and PS systems, on the other hand, move more slowly and are often implicated in flash flooding as a result of slow-moving, 'training', and/or back-building cells. For example, heavy rainfall over Taiwan has been attributed to back-building cells associated with the Mei-yu front (Li *et al.*, 1997).

It has long been known that the organization of tropical convection is influenced predominantly by the vertical shear and convective available potential energy or CAPE (Moncrieff and Green, 1972). Various observational studies in the eastern Atlantic and northern Australia have confirmed the strong influence of environmental winds on the structure, orientation, and propagation of convective bands

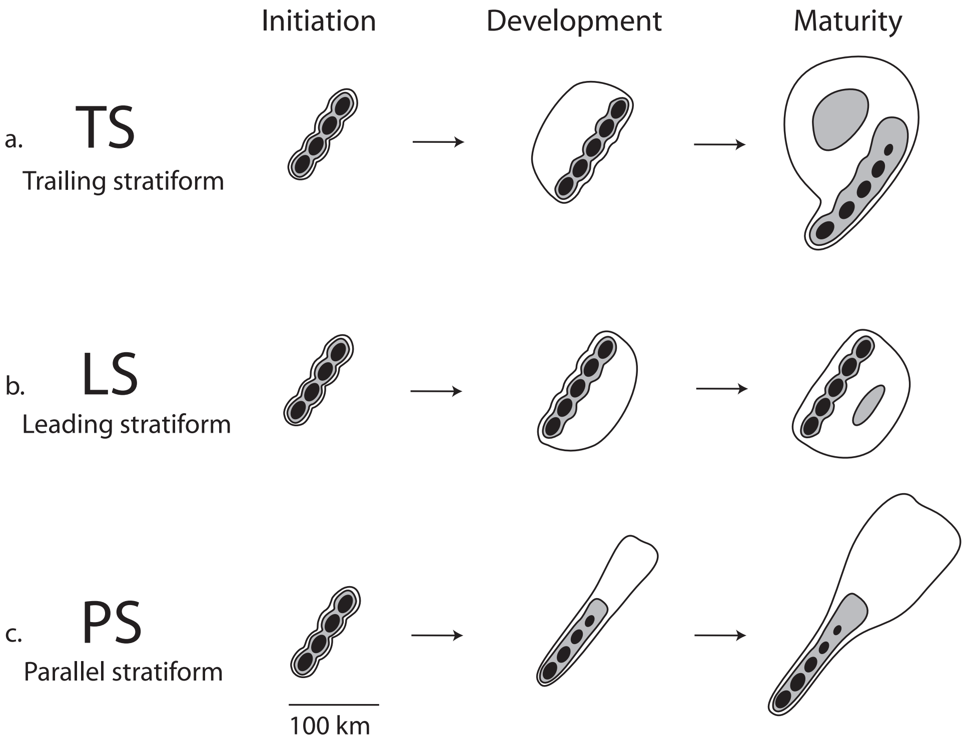

Figure 8.3. Schematic reflectivity drawing of idealized life cycles for three linear MCS archetypes: (a) TS, (b) LS, and (c) PS. Approximate time intervals between phases: for TS 3–4 h; for LS 2–3 h; for PS 2–3 h. Levels of shading roughly correspond to, 20, 40, and 50 dBZ. From Parker and Johnson (2000).

(e.g., Barnes and Sieckman, 1984; Alexander and Young, 1992; Keenan and Carbone, 1992). LeMone *et al.* (1998) investigated the organization of convection over the western Pacific warm pool using aircraft data from the TOGA/COARE. In agreement with Alexander and Young (1992) they found that vertical shear in the low to mid-troposphere is a key factor in determining the orientation of convective bands, while CAPE influences their depth and longevity. Their results have been recently supported by numerical simulations of convection in shear by Robe and Emanuel (2001).

The results of LeMone *et al.* (1998) have been recently extended to the Asian summer monsoon by Johnson *et al.* (2005). The modes of organization of convection over the northern South China Sea during the onset of the 1998 east Asian summer monsoon have been determined using the BMRC C-POL radar located on Dongsha Island (Figure 8.4). This figure, adapted from LeMone *et al.* (1998), is a summary of the findings from SCSMEX. In general, the organizational modes for SCSMEX were consistent with those determined by LeMone *et al.* (1998) for the western Pacific warm pool. It is found that when the shear in the lowest 200 hPa exceeds $4\,\mathrm{m\,s^{-1}}$ and

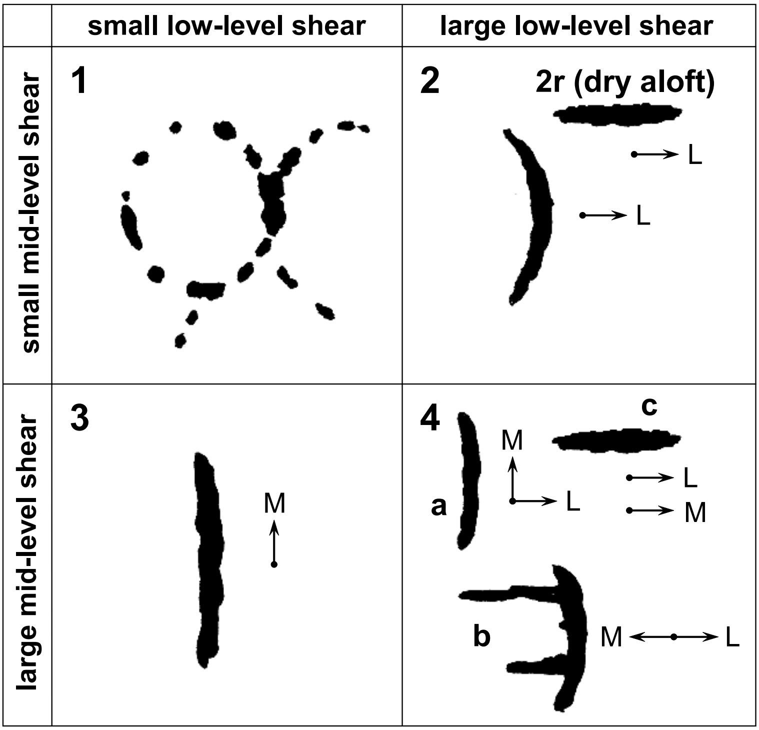

Figure 8.4. Schematic depiction adapted from LeMone *et al.* (1998) of four main categories of convective structures for given vertical shears in the lower troposphere (1,000–800 hPa) and at middle levels (800–400 hPa) based on COARE observations, but modified to include results from SCSMEX (modes 2r and 4c added). Length of schematic convective bands is ~100–300 km; line segments in upper left frame are up to 50 km in length. Cutoff between 'strong' and 'weak' shear for the lower layer (1,000–800 hPa) is $4\,m\,s^{-1}$ and for the middle layer (800–400 hPa) is $5\,m\,s^{-1}$. Arrows marked L and M are shear vectors for lower and middle layers, respectively. See text for description of convective modes.
From Johnson *et al.* (2005).

the shear from 800–400 hPa is less than $5\,m\,s^{-1}$, the orientation of the primary convective band in MCSs is perpendicular to the low-level shear (type 2 in Figure 8.4). Secondary lines parallel to the low-level shear are found in some cases ahead of these bands. In the absence of strong low-level shear, lines form parallel to the 800–400 hPa shear when its magnitude exceeds $5\,m\,s^{-1}$ (3 in the lower left frame of Figure 8.4). When the vertical shear exceeds the thresholds in both layers and the shear vectors are not in the same direction, the primary band is normal to the low-level shear (4a or 4b in the lower right frame of Figure 8.4). Trailing secondary bands parallel to the mid-level shear occur if the mid-level shear is opposite the low-level shear (4b). When the shear in both layers is weak, convection develops in arcs along outflow boundaries (1). Two additional modes of convection have been identified from analysis of SCSMEX C-POL radar data (Figure 8.4): shear-parallel bands (2r) for strong low-level shear and weak mid-level shear when the air is dry aloft, and shear-parallel bands (4c) for strong shears in both layers when the shear vectors are in the same direction. The latter two modes are related to the passage of mid-latitude troughs (Johnson *et al.*, 2005).

8.2.2 Latent heating profiles and the monsoon

MCSs produce areas of convective and stratiform precipitation that vary over the life cycle of the storms. These two precipitation features have strongly contrasting latent heating profiles, as illustrated in Figure 8.5 for the mature stage of a westward-propagating tropical squall line. The convective region is characterized by a single, condensational heating peak in the low to mid-troposphere. The stratiform region has a peak in the upper troposphere associated with condensation or deposition aloft, and a cooling peak in the lower troposphere associated with melting and evaporation (Houze, 1982; Johnson, 1984). The dynamical effects of latent heating can be assessed through the potential vorticity equation for frictionless flow:

$$\frac{dP}{dt} = -(\varsigma_\theta + f)\frac{\partial\dot{\theta}}{\partial p} - \left[\mathbf{k} \times \nabla_\theta\dot{\theta} \times \frac{\partial\mathbf{v}}{\partial\theta}\right]\frac{\partial\theta}{\partial p} \tag{8.1}$$

where $\dot{\theta}$ is the diabatic heating rate, $\varsigma_\theta + f$ is the absolute vorticity on an isentropic surface, and $P \equiv -(\varsigma_\theta + f)\partial\theta/\partial p$ is the potential vorticity. From the first term on the RHS of equation (8.1), the increase with height of diabatic heating at mid-levels within the stratiform region of an MCS (Figure 8.5) produces a positive potential vorticity anomaly in the mid-troposphere. This heating distribution often leads to the generation of a mid-level mesoscale convective vortex or MCV (Zhang and

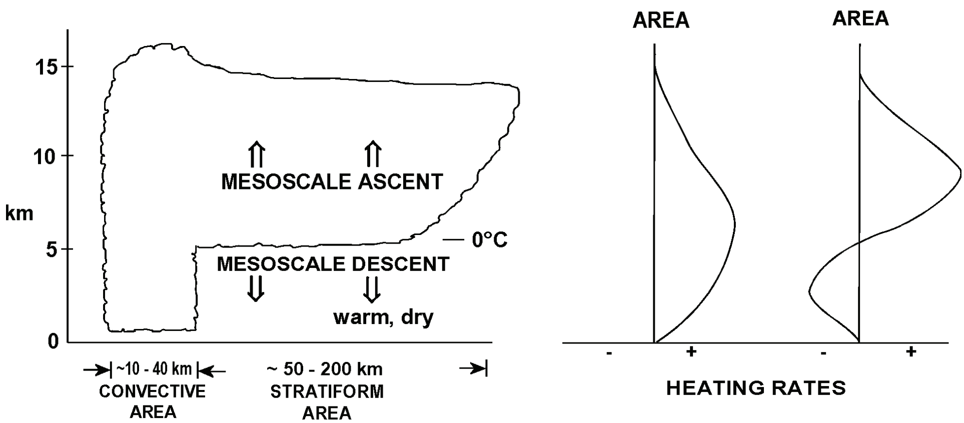

Figure 8.5. Idealized mature stage of MCS illustrating convective and stratiform precipitation areas along with associated heating profiles.
From Johnson (1986) Meteorological Society of Japan.

Fritsch, 1987; Raymond and Jiang, 1991; Hertenstein and Schubert, 1991). MCVs appear to be common over China in the Yangtze valley during the Mei-yu season, potentially contributing to long-lived precipitation systems, heavy rainfall, and flash floods (e.g., Akiyama, 1984a,b). They have also been observed in the Australian summer monsoon (Keenan and Rutledge, 1993). In addition, MCVs are thought to be potential precursors or contributors to tropical cyclogenesis (Velasco and Fritsch, 1987; Montgomery and Enagonio, 1998).

The mesoscale characteristics of convection have an important impact on the temporal and spatial distribution of monsoon heat sources and sinks. Yanai *et al.* (1973) introduced the following notation and terminology in connection with the conservation laws for heat and moisture:

$$Q_1 = \frac{\partial \bar{s}}{\partial t} + \bar{\mathbf{v}} \times \nabla \bar{s} + \bar{\omega}\frac{\partial \bar{s}}{\partial p} = Q_R + L(\bar{c} - \bar{e}) + \frac{\partial}{\partial p}\overline{s'\omega'} \tag{8.2}$$

$$Q_2 = -L\left(\frac{\partial \bar{q}}{\partial t} + \bar{\mathbf{v}} \times \nabla \bar{q} + \bar{\omega}\frac{\partial \bar{q}}{\partial p}\right) = L(\bar{c} - \bar{e}) + L\frac{\partial}{\partial p}\overline{q'\omega'} \tag{8.3}$$

where $s = c_p T + gz$ is the dry static energy, Q_R the radiative heating rate, c the condensation rate, e the evaporation rate, q the specific humidity, and the deviations from horizontal averages (indicated by overbars) are denoted by primes. Q_1 and Q_2 are the residuals of heat and moisture budgets of the 'resolvable' motion, first introduced by Yanai (1961). They are called the 'apparent' heat source and moisture sink, respectively, because they include true sources and sinks (condensation and evaporation) as well as correlation terms resulting from unresolved eddies (Yanai and Johnson, 1993). Comparing equations (8.2) and (8.3), it can be seen that for precipitation systems possessing negligible eddy transports (i.e., stratiform precipitation systems), the profiles of $Q_1 - Q_R$ and Q_2 should closely match (Luo and Yanai, 1984; Arakawa and Chen, 1987). When deep convection is present, the peaks in Q_1 and Q_2 are separated (e.g., as in the tropical western Pacific study of Yanai *et al.* (1973)).

Luo and Yanai (1984) computed heat and moisture budgets over and around the Tibetan Plateau for the period 26 May to 4 July 1979 using sounding data from the First GARP Global Experiment (FGGE). Their results are shown in Figure 8.6 for four regions. Region I, which is centered over the western Plateau near 34°N, 80°E, shows a pattern of mean upward motion with a maximum Q_1 just above the surface and negligible Q_2. This structure is indicative of a strong sensible heat flux in this region with very little precipitation. Recently, Ueda *et al.* (2003a) found a somewhat different result using 1998 GAME (GEWEX Asian Monsoon Experiment) reanalysis data, namely, that heating by deep convection nearly equaled that due to the sensible heat flux over the western Plateau. This finding may indicate important interannual variability in this region, or it may point to the difficulty in using reanalysis data, which are sensitive to convective parameterizations, to assess heating profiles. Over the eastern Plateau (Region II, centered near 34°N, 95°E) the mean vertical motion is still upward, but latent heat release is more important, contributing to nearly half the apparent heat source. Region III, centered over the Yangtze valley

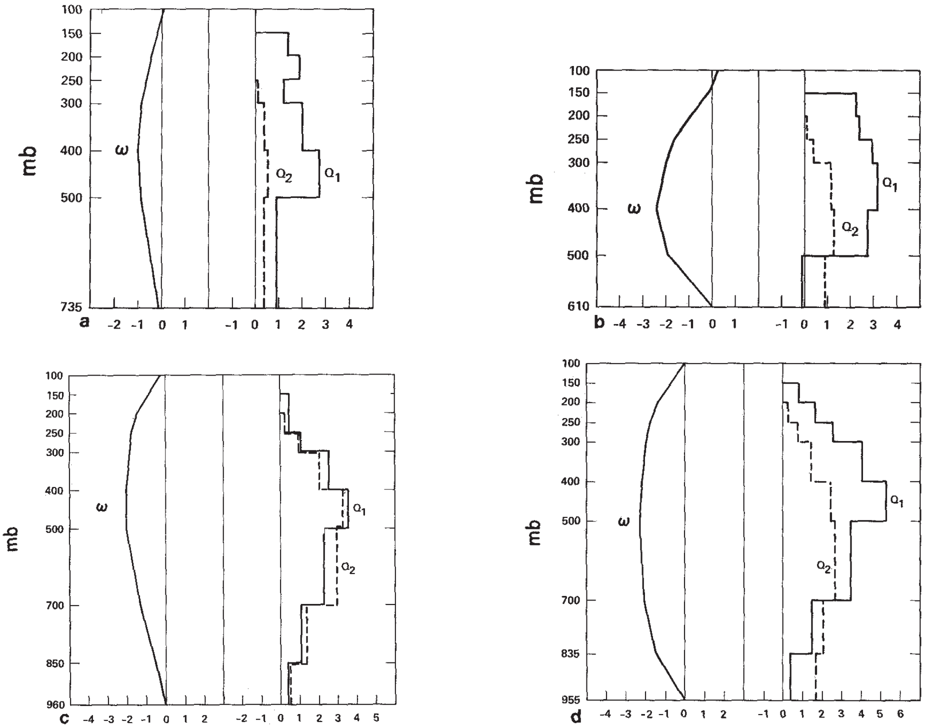

Figure 8.6. Forty-day mean vertical distributions of areal mean vertical p-velocity (hPa h^{-1}), heating rate Q_1, and drying rate Q_2 for (a) Region I, (b) Region II, (c) Region III, and (d) Region IV.
From Luo and Yanai (1984).

(near 28°N, 115°E), exhibits large Q_1 and Q_2 which are virtually coincident in the mid-troposphere, suggestive of a predominance of stratiform precipitation in that region. Finally, Region IV, centered in the Assam–Bengal area near 25°N, 85°E, shows large Q_1 and Q_2 peaks, but displaced from each other, indicative of the prevalence of deep convection.

There have been other studies of the latent heating profiles over China during the summer monsoon for the FGGE year (Nitta, 1983; Kato, 1985) and later years (Ding and Hu, 1988; Ding and Wang, 1988; Johnson *et al.*, 1993), and over Taiwan during the 1987 Taiwan Area Mesoscale Experiment or TAMEX (Johnson and Bresch, 1991). Johnson *et al.* (1993) found that during the pre-Mei-yu and Mei-yu periods of 1987, 1988, and 1989, precipitation type over the Yangtze region was a mixture of convective and stratiform, whereas over southern China deep convection predominated. Inferences from satellite data support these findings (e.g., Ninomiya, 1989), but it has only been in recent years that radar data (including Doppler observations) have been available to document in detail the convective and stratiform properties of MCSs over Tibet and within the Mei-yu and Baiu frontal zones

(Uyeda *et al.*, 2001; Shinoda and Uyeda, 2002; Chen *et al.*, 2003; Yamada *et al.*, 2003).

As noted earlier, the vertical distribution of latent heating determines the generation of potential vorticity anomalies. On the other hand, the generation of available potential energy A_E can be written as:

$$\frac{\partial A_E}{\partial t} = -\frac{[v'T']}{\sigma}\frac{\partial[T]}{\partial y} - \frac{[\omega'T']}{\sigma}\frac{\partial[T]}{\partial p} + \frac{R}{gp}[\omega'T'] + \frac{[Q_1'T']}{c_p\sigma} \tag{8.4}$$

where brackets refer to a zonal average and σ is the static stability. Thus, there is a generation of positive A_E where positive heating anomalies coincide with warm anomalies. Hence, an accurate determination of the vertical and horizontal distribution of Q_1 as well as the temperature field is needed to properly represent both the dynamics and energetics of monsoon circulations and disturbances in prediction models.

Since the heating profiles in the convective and stratiform regions contrast greatly (Figure 8.5), the relative amounts of convective and stratiform precipitation in cloud systems should impact the large-scale response to convection. A climatology of the relative contributions of convective and stratiform precipitation to total rainfall has been carried out by Schumacher and Houze (2003) using the TRMM precipitation radar. Their results (Figure 8.7, color section) show a wide variation in the stratiform rain fraction over the globe, with greatest amounts (50–60%) over the central and western Pacific and Indian Ocean. Low stratiform rain fractions (20–30%) are observed over Africa, parts of the Maritime Continent, and the Caribbean. This variability is not fully understood, but is likely related to the differing instability, humidity, and wind shear profiles in the different locations. Schumacher *et al.* (2004) have recently used the observed stratiform rain fractions in Figure 8.7 in an idealized general circulation model to show that the strong zonal gradient in the stratiform rain fraction across the equatorial Pacific, from ~30% over Indonesia to ~60% over the eastern Pacific, accounts for the observed tilted structure of the east–west Walker circulation.

The impact of convection on the modeled large-scale circulation appears to be critically sensitive to the heating distribution. The theoretical basis for this sensitivity is discussed in Hartmann *et al.* (1984). Early modeling studies of the Madden–Julian Oscillation (MJO) have shown an important sensitivity of the characteristics of this global-scale wave to the vertical heating distribution (e.g., Lau and Peng, 1990). In particular, when the heating peak is in the upper troposphere, the wave appears to move too fast, whereas when the peak is in the lower troposphere, the phase speed is more in line with observations (e.g., Sui and Lau, 1989). However, more recently, Lin and Johnson (1996a), Tung *et al.* (1999), Mapes (2000), and Lin *et al.* (2004) have presented Q_1 profiles for TOGA/COARE MJO events indicating a 'top-heavy' structure to the heating and suggesting (from Figure 8.5) a large contribution from stratiform precipitation. Using TRMM data, Lin *et al.* (2004) confirmed that the MJO for the COARE region has an anomalously large stratiform rain fraction of 60%, much larger than the long-term mean of ~40%. They also found a significant tilt in the heating profile in the MJO over the western Pacific, with heating in stratiform anvils (latent plus radiative) lagging the surface precipitation maximum by a few days. Kiladis *et al.* (2005) also found this tilted structure over the western Pacific (but not the Indian Ocean) and noted that the evolution of the heating field leads to a temperature structure that favors MJO growth. Yanai *et al.* (2000) recently computed the generation of A_E for the MJO observed during TOGA/COARE and found a strong generation by deep convection in the upper troposphere in the active part of this disturbance, while 'convective damping' (a negative value of the last term in equation (8.4)) was found in the lower troposphere.

The latent heating profiles within MCSs also have a profound influence on the mesoscale environment of convection. In particular, convective heating generates gravity waves which propagate away from the convective source (Nicholls *et al.*, 1991; Mapes, 1993). In the work of Nicholls *et al.* (1991) thermally forced gravity waves were considered in the context of a 2-D, linear, incompressible, hydrostatic, and Boussinesq fluid. The thermal forcing was considered to consist of contributions from both convective and stratiform precipitation (as illustrated in Figure 8.5). The sum of the two modes of heating yields positive heating at mid and upper levels and cooling near the surface. This form of heating produces two prominent gravity wave modes, the first internal or $n = 1$ mode and the second internal or $n = 2$ mode. For an atmosphere at rest, the horizontal phase speeds of internal modes are given by $c = NH/n\pi$, where N is the Brunt–Väisälä frequency and H is the fluid depth. Using a typical value of N from 2 to 8 km of $1.1 \times 10^{-2}\,\mathrm{s}^{-1}$ and $H = 12\,\mathrm{km}$, then c for the $n = 1$ and $n = 2$ modes is $\sim$42 and $21\,\mathrm{m\,s}^{-1}$, respectively. Thus, the waves rapidly disperse.

Both modes produce subsidence in the upper troposphere as the waves propagate away from the source, as shown in Figure 8.8 for a superposition of the solutions with a rigid upper lid two hours after the heating has been turned on. Relative to the initial state, the heating produces positive buoyancy in the upper troposphere and negative buoyancy in the lower troposphere. Near the source there is a positive pressure perturbation in the upper troposphere and a minimum in the mid-troposphere. The flow is away from the source in the upper and lower troposphere and toward the source at mid-levels. The latter feature occurs near the melting level since the base of the stratiform anvil is there (Figure 8.5) (i.e., the node in the $n = 2$ profile is near the 0°C level). This mid-level inflow has been referred to by Mapes and Houze (1995) as 'melting convergence.' The solutions are modified somewhat when the rigid lid is removed such that gravity wave energy propagates upwards into the stratosphere and the intensity of the circulations in the troposphere are reduced away from the source, but the overall qualitative results are unchanged (Nicholls *et al.*, 1991; Pandya *et al.*, 1993). Outside the tropics (e.g., in the MCV-forming region of China), the Earth's rotation leads to geostrophic adjustment for a heat source on the scale of MCSs such that the warming is trapped near the source rather than propagating away as shown in Figure 8.8 (Johnson and Mapes, 2001), which could lead to the formation of MCVs. The upward displacement at low levels near the source caused by the $n = 2$ or 'stratiform' mode can reduce convective inhibition and also by cooling,

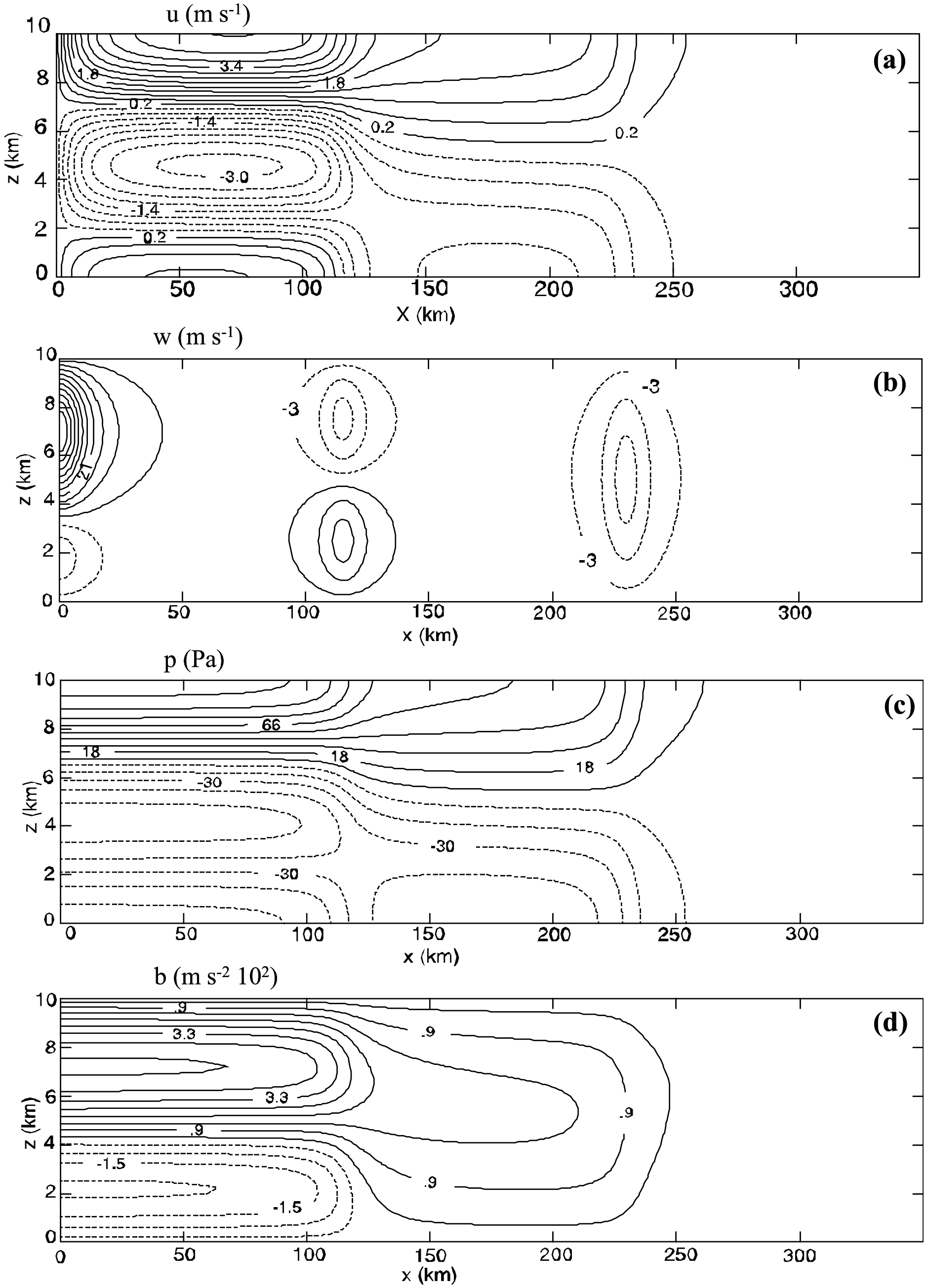

Figure 8.8. Rigid-lid solutions (top to bottom) for perturbations of u, w, p, and buoyancy b for superposition of $n = 1$ and 2 modes. The magnitude of the heat source at the origin is $1.0\,\mathrm{J\,kg^{-1}}$ with a half-width of 10 km, $H = 10$ km, $N = 0.01\,\mathrm{s^{-1}}$, and $t = 2$ h.
From Nicholls *et al.* (1991).

increase the relative humidity. Both effects favor new convection adjacent to old, possibly explaining why convection in the tropics and monsoon regions tends to cluster together or be 'gregarious' (Mapes, 1993). This concept has been generally supported by modeling studies of the clustering of deep convection (Lac *et al.*, 2002), but in a 4-D setting with many MCS heat sources growing and decaying, detailed patterns of vertical displacement are very complex (Mapes, 1998).

It has been pointed out by Mapes (2000) that the top-heaviness of deep-convective heating profiles is inadequate to balance radiative cooling, which is large through the lower troposphere. This finding implies that in the tropics there must be abundant precipitating shallow cumulus and cumulus congestus clouds to provide lower tropospheric heating, essentially to offset widespread cooling beneath stratiform anvils. This finding has been borne out by radar observations of clouds in TOGA/COARE (DeMott and Rutledge, 1998; Johnson *et al.*, 1999; Short and Nakamura, 2000). This topic leads us to a consideration of cloud populations in the tropics and Asian monsoon.

8.2.3 Cloud populations

It has long been known that trade wind cumulus and deep cumulonimbus represent primary components of the broad spectrum of cumulus clouds in the tropics (e.g., Riehl, 1979). Observations from TOGA/COARE have provided new evidence concerning the distribution of tropical clouds (DeMott and Rutledge, 1998; Rickenbach and Rutledge, 1998; Johnson *et al.*, 1999). Johnson *et al.* showed cumulus echo top statistics from the 5-cm radar aboard the R/V *Vickers* indicating that cumulus congestus clouds are the most abundant of all precipitating clouds over the western Pacific warm pool. Specifically, congestus with tops between 4.5 and 9.5 km represent 57% of the precipitating convective clouds in COARE and account for 28% of the total rainfall. Radar studies from GATE reveal a similar result (Houze and Cheng, 1977; Cheng and Houze, 1979). The prevalence of congestus clouds is related to processes that restrict the development of deep convection: entrainment of dry air, precipitation loading, and the existence of a stable layer near the melting layer (Johnson *et al.*, 1996; Mapes and Zuidema, 1996; Redelsperger *et al.*, 2002). The growth of congestus clouds in COARE was often limited to just above the 0°C level, which limited enhancement of cloud buoyancy through glaciation (Zuidema, 1998). The existence of a congestus maximum in the cloud population, along with the well-known maxima of shallow and deep cumulus, has led to the concept of a 'trimodal distribution' of tropical convective clouds (Johnson *et al.*, 1999), which is schematically depicted in Figure 8.9.

While deep convective clouds are the dominant rain producers in the tropics and monsoon regions, shallower clouds serve an important role in moistening the environment and preconditioning it for deep convection. For example, Johnson *et al.* (1999) used shipboard radar observations in TOGA/COARE to document increasing populations of shallow cumulus and cumulus congestus clouds, along with a deepening moist layer in the lower troposphere prior to the active phase of the December 1992 MJO. While such clouds detected by radar

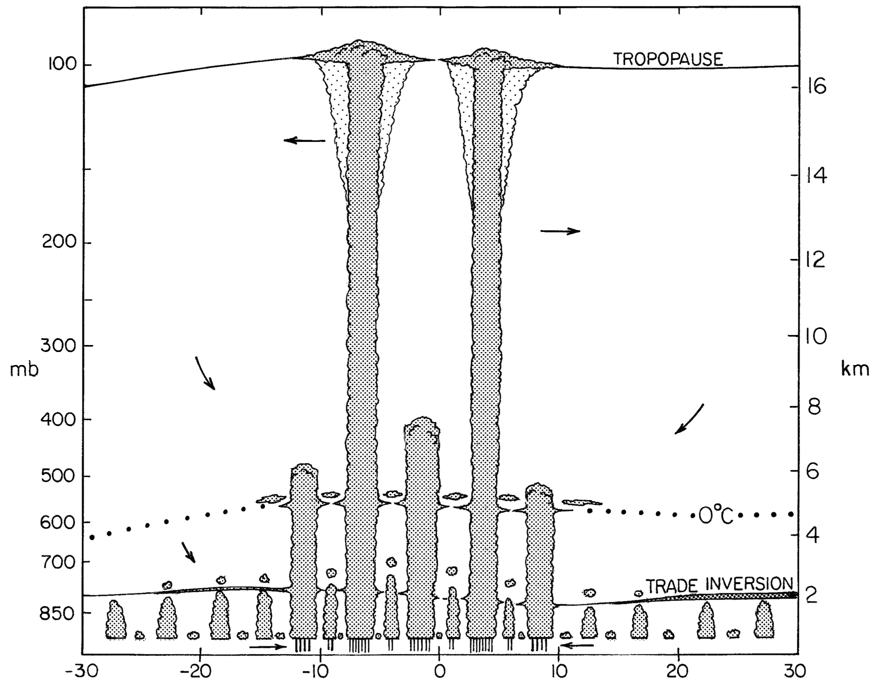

Figure 8.9. Conceptual model of tropical cumulus cloud distributions from 30°N to 30°S based on COARE mean radar data and thermal stratification. Three main cloud types are indicated: shallow cumulus, cumulus congestus, and cumulonimbus. Within the shallow cumulus classification, there are two subdivisions: forced and active cumulus. Three stable layers are indicated: the trade inversion, the 0°C layer, and the tropopause. Shelf clouds and cloud debris near the trade and 0°C stable layers represent detrainment there. Cirrus anvils occur near the tropopause. Considerable overshooting of the trade and 0°C stable layers occurs in the equatorial trough zone. Arrows indicate meridional circulation. Although a double intertropical convergence zone (ITCZ) is indicated, representing Intense Observing Period (IOP)-mean, this structure is transient over the warm pool and a single ITCZ often exists.
From Johnson *et al.* (1999).

produce rain, they do so inefficiently, thereby moistening the low to mid-troposphere (Raymond and Torres, 1998). Moreover, a large amplitude diurnal cycle of the SST (2–3°C) over the western Pacific warm pool during the light-wind phase of the MJO contributes to a deepening of the boundary layer (Johnson *et al.*, 2001) and the development of afternoon showers (Chen and Houze, 1997; Rickenbach and Rutledge, 1997; Sui *et al.*, 1997). Without this diurnal enhancement, precipitation over the warm pool and its associated lower tropospheric moistening would likely be far less during the suppressed phase of the MJO (Webster *et al.*, 1996). This preconditioning likely plays a key role in setting the timescale for the MJO (Bladé and Hartmann, 1993; Hu and Randall, 1994; Kemball-Cook and

Weare, 2001). In a simulation using the Hadley Center HadAM3 model, Inness *et al.* (2001) obtained an improved simulation of the MJO when they increased the vertical resolution in the model. Higher resolution allowed the melting layer to be better resolved, leading to more congestus clouds and an improved representation of the moistening and preconditioning of the lower troposphere in advance of the active phase of the MJO.

There is also evidence from the Asian monsoon of the importance of shallow cumulus clouds in the development of convection along the Mei-yu front. Shinoda and Uyeda (2002) found that in the southern part of the Mei-yu front where there is moist southerly flow, shallow cumulus can develop into deep convection because the clouds do not dissipate through evaporation. In addition, they found that the shallow cumulus serve an important moistening role for the mid-troposphere, transporting water vapor upward from the boundary layer.

8.2.4 Convective momentum transports

In addition to their roles in heating and moistening the atmosphere, deep convection has an important impact on the momentum field. Part of this impact arises through the gravity wave response to convective heating. For example, Pandya and Durran (1996) show that the time-averaged, tilted dipole heating structure in squall lines with trailing stratiform precipitation (a relatively narrow, tilted band of heating followed by a broader, tilted region of cooling in the lower troposphere) can explain the essential circulation features of squall lines: an upward-sloping, front-to-rear flow followed by a descending rear-inflow jet. However, convective clouds also serve to transport momentum vertically. There is evidence to suggest that the momentum transport is a function of convective organization with up-gradient transport in the line-normal direction and down-gradient transport in the line-perpendicular direction (LeMone, 1983; Wu and Yanai, 1994; Tung and Yanai, 2002; Moncrieff, 2004). Over much of the Asian monsoon it is expected that the greatest role for convective momentum transport is simple damping by down-gradient transport.

The MJO, which is a fundamental component of the intraseasonal variability of the Asian monsoon, appears to be influenced significantly by convective momentum transports. By placing aircraft observations from COARE in the context of a Kelvin–Rossby wave paradigm for the MJO, Houze *et al.* (2000) found that very large MCSs, referred to as superconvective systems, located within and to the west of the region of maximum heating have broad mid-level inflows that transport westerly momentum downward, serving to enhance the westerly wind bursts associated with the MJO (see also Moncrieff and Klinker, 1997; Tung and Yanai, 2002). To the east, near the boundary between westerlies and easterlies, the mid-level inflows transport environmental easterly wind downward, thereby enhancing low-level convergence in the MJO. Biello and Majda (2005) developed a multiscale model of the MJO that includes congestus heating in the eastern part of the disturbance and heating by westward-tilted superconvective systems to the west. The superconvective systems drive westerly momentum downward while the congestus heating enhances easterly flow in the leading portion of the MJO envelope, and the model reproduces many of the observed structural features of the MJO. Moncrieff (2004) and Kiladis *et al.* (2005) also found evidence of substantial vertical fluxes of zonal momentum in the MJO, and the latter study provides evidence of similar heating and momentum transports in other propagating convectively coupled equatorial waves.

8.2.5 Diurnal cycle of convection

The diurnal cycle of precipitation is a dominant feature of the Asian monsoon. On the large scale the Tibetan Plateau generates significant diurnally varying circulations, vertical motion, and diabatic heating features (Luo and Yanai, 1983; Nitta, 1983; Krishnamurti and Kishtawal, 2000). On the mesoscale there are local land and sea breezes, mountain/valley circulations, and surface heterogeneities that influence precipitation patterns over the monsoon regions of the World.

There have been numerous studies of the diurnal cycle of convection over Asia (Murakami, 1983; Nitta and Sekine, 1994; Chen and Takahashi, 1995; Ohsawa *et al.*, 2001; Fujinami and Yasunari, 2001; Kurosaki and Kimura, 2002). Over land, many areas exhibit an afternoon maximum of convection, as expected from daytime heating; however, certain regions such as the base of the Himalayas and mountain basins (e.g., the Sichuan Basin) have a late night, early morning maximum (Akiyama, 1989; Johnson *et al.*, 1993; Ohsawa *et al.*, 2001). Over the open ocean, an early morning maximum of precipitation has been observed, which has been attributed to horizontal gradients in radiative cooling between cloud systems and their environment (Gray and Jacobson, 1977), daytime stabilization of the upper troposphere by short-wave heating (Kraus, 1963; Randall *et al.*, 1991), and/or the life cycle effects of MCSs (Chen and Houze, 1997). While an early morning maximum has been found over some ocean areas around Asia, the diurnal cycle there is rather complex. Ohsawa *et al.* (2001) find late night, early morning maxima near the coastlines of south Asia, Thailand, Sumatra, Malaysia, and Borneo which they attribute to an interaction of mountain or land breezes with the prevailing wind.

During the 1978 WMONEX, the diurnal cycle of convection off the north coast of Borneo was studied in detail using radar and sounding data. Houze *et al.* (1981) documented the development of nocturnal MCSs off Borneo, arguing they were a result of low-level convergence of the night-time land breeze with the north-east monsoon flow. The MCSs (discussed in detail in Chapter 3) typically began as a group of convective cells near the coastline and later expanded to a several hundred-km scale dimension with both convective and stratiform components, later dissipating after sunrise as the sea breeze developed.

In a study of convection over Taiwan during the TAMEX, Johnson and Bresch (1991) suggested that the land breeze flow at night was augmented by evaporation of the previous evening's precipitation over the interior elevated terrain. Mapes *et al.* (2003) proposed that the land breeze by itself was inadequate to account for nocturnal convection that regularly occurs offshore Columbia in the Panama Bight. They argued that thermally forced gravity waves (produced by elevated terrain and propagating at about $15\,\mathrm{m\,s^{-1}}$) are an essential part of the process, and that they produce a warm anomaly offshore during the daytime, thereby capping convection, while a cooling is produced at night, thus allowing convection to develop.

Understanding the diurnal cycle of convection in coastal environments is important because so much precipitation occurs there and global models do not properly represent the diurnal cycle of convection (Yang and Slingo, 2001). There are possible global consequences of this deficiency, as the Maritime Continent heat source is too weak in the mean (Neale and Slingo, 2003). In the region of the Asian monsoon, satellite data indicate southward propagation of precipitation systems from India over the Bay of Bengal. This propagation is evident in a time–latitude diagram of brightness temperatures over the Bay of Bengal (Webster *et al.*, 2002, their figure 4). Precipitation systems (inferred from the cold cloud tops) were found to propagate all the way from the India coast near 20°N to the equator. Radar data from the R/V *Ron Brown* in the Bay of Bengal indicate that the convection associated with the diurnal signal has characteristics of TS systems described in Section 8.2.1.

A southward propagation of convective systems over the South China Sea was also observed during the 1998 SCSMEX, similar to that over the Bay of Bengal. The monsoon onset over the northern South China Sea (near 20°N) occurred around mid-May and is characterized by a regular signal of southward propagation of convection (low values of IR brightness temperature) at an approximate speed of $15\,m\,s^{-1}$ (Figure 8.10, color section). In late May the convection shifts southward to the central South China Sea (10–15°N) with a diurnal propagating signal still present, indicating that the diurnal pattern is independent of coastal effects. Then in June the convection shifts back again to China, and diurnal propagation persists.

Propagation tied to the diurnal cycle is also present over land in monsoon regions. Wang *et al.* (2004) documented a diurnal cycle of convection over the eastern Tibetan Plateau, peaking in the late afternoon or early evening then propagating eastward. Kousky (1980) and Molion (1987) presented sequences of satellite images illustrating the afternoon coastal genesis and subsequent inland propagation of squall lines in the Amazon basin. Squall lines over West Africa propagate westward over great distances in association with African easterly waves (see review by Houze and Betts (1981)), and a maximum in convective cloud coverage has been found to occur near or shortly after midnight over West Africa (McGarry and Reed, 1978).

The diurnal cycle of precipitation systems over the ocean has been related to their size by Chen *et al.* (1996). Using satellite data from COARE, they found that the diurnal cycle of accumulated cloudy area of cold cloud tops (clouds with infrared temperatures less than 208 K) is a function of the size of the cloud system (Figure 8.11). The area covered by the smallest cloud clusters (Class 1) had a very small diurnal amplitude, whereas the largest clusters (Class 4) had a strong diurnal variation (nearly 10 : 1 amplitude) with a peak in the early morning hours and a minimum in the afternoon (as also found by Mapes and Houze, 1993). This diurnal cycle reflects the upscale growth of convection to MCS dimensions during the nighttime hours.

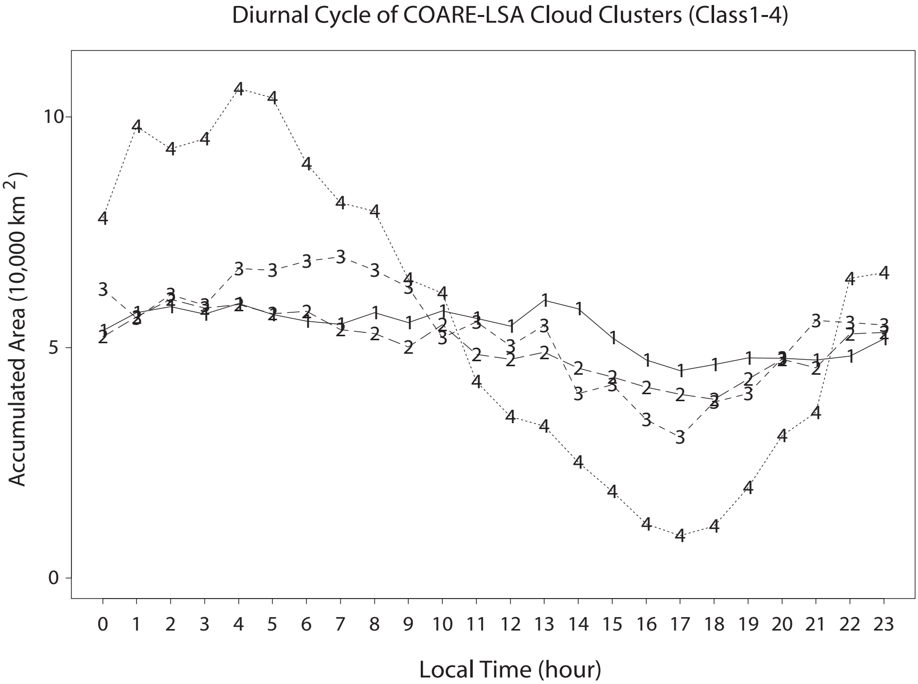

Figure 8.11. Diurnal cycle of accumulated cloudy area covered by the 208 K cloud clusters over the domain of 10°N–10°S, 152°E–180° for each of the four class sizes.
From Chen *et al.* (1996).

8.3 TOPOGRAPHICALLY FORCED LOCAL CIRCULATIONS

Throughout the Asian monsoon region, topography has a significant impact on local weather and precipitation (Riehl, 1954; Ramage, 1971; Ding, 1994). In Figure 8.1, the heavy rainfall at the foot of the Himalayas and adjacent to mountainous coastal regions (e.g., western India, Myanmar, Thailand, and Sumatra) is linked to topographic effects. Although not obvious from Figure 8.1, much of the heavy coastal rainfall occurs just offshore rather than over the windward slopes of the coastal ranges. This behavior has been noted and studied for the heavy rain along the coast of western India upstream of the western Ghats by Krishnamurti *et al.* (1983b) and Grossman and Durran (1984). Grossman and Durran carried out a modeling study explaining this phenomenon in terms of upstream blocking. However, Smith (1985) argued that their study did not include important physics such as the effects of wind shear, air–sea interaction, Coriolis force, and latent heat release. A modeling study of Ogura and Yoshizaki (1988) included those effects and concluded that the positioning of the heaviest rainfall just offshore is dependent on the strong vertical wind shear (low-level westerlies and upper level easterlies) and strong surface fluxes over the ocean. Upper level easterlies advect the cirrus aloft westward over the open oceans (Krishnamurti *et al.*, 1983b), so there is the impression from infrared satellite imagery that the heavy rainfall is spread far offshore; however, most of it is confined near the coast. Similar flow reversals occur during the boreal summer monsoon off Myanmar and the west coast of the Philippines, possibly helping to explain the similar behavior in those regions. Other possible factors contributing to the offshore precipitation are land breeze effects and coastal frictional convergence.

Recently, Chang *et al.* (2005b) studied the relationship of rainfall to the monsoon flow and topography over Indo-China and the Maritime Continent using the TRMM precipitation radar (PR) and QuikSCAT data. A map of topography over this region along with DJF and JJA QuikSCAT winds is shown in Figure 8.12 (color section). Over most of the region there is a marked seasonal reversal of the flow. During boreal winter, there is onshore flow toward coastal mountain ranges in Vietnam, Malaysia, and along the east side of the Philippines. This onshore flow contributes to boreal winter monsoon rainfall maxima in these regions, as seen in Figure 3.9 (DJF minus JJA TRMM PR rainfall and QuikSCAT winds). Positive (negative) anomalies indicate maximum precipitation in boreal winter (summer). In addition to the positive anomalies in boreal winter, there are negative anomalies off the west coasts of Myanmar, Cambodia, and the Philippines indicating maximum rainfall during boreal summer in association with south-westerly monsoon flow (the reverse of that shown in Figure 3.9). These results emphasize the important role of topography on precipitation distributions in the Asian monsoon region.

There are many other mesoscale topographic effects in the monsoon regions of the World affecting local precipitation patterns. For example, flow blocking by Taiwan during the summer monsoon often leads to a low-level jet through the Taiwan Straits north-west of the island and lee vortices downstream, both of which can affect precipitation patterns around Taiwan (e.g., Chen and Yu, 1988; Wang and Chen, 2002). Watanabe and Ogura (1987) found that flow deflection by a mountain range contributed to extreme rainfall within the Baiu front in a 23 July 1983 storm along the west coast of Japan.

A particularly complex topographic effect reported by Xie *et al.* (2003) concerns the impact of the Annam Cordillera (the north–south mountain range on the east coast of Indo-China shown in Figure 8.12 (color section)) on the flow over the South China Sea. During the summer months the south-westerly monsoon flow impinging on the Annam Cordillera creates a strong low-level jet off the south Vietnam coast (Figure 8.13). This jet leads to coastal upwelling of cool water, which is enhanced by Ekman upwelling due to the cyclonic curl of the wind stress on the north side of the jet. Figure 8.13 shows the coolest water displaced just north of the jet axis. The development of this cold filament in midsummer disrupts the summer warming of the South China Sea and causes a pronounced semiannual cycle in the SST. There is considerable interannual variability in this cold filament (e.g., it did not develop during the 1998 SCSMEX year).

Mountain and valley flows and sea and land breezes are important manifestations of the impact of the diurnal cycle of solar heating on terrain features. These

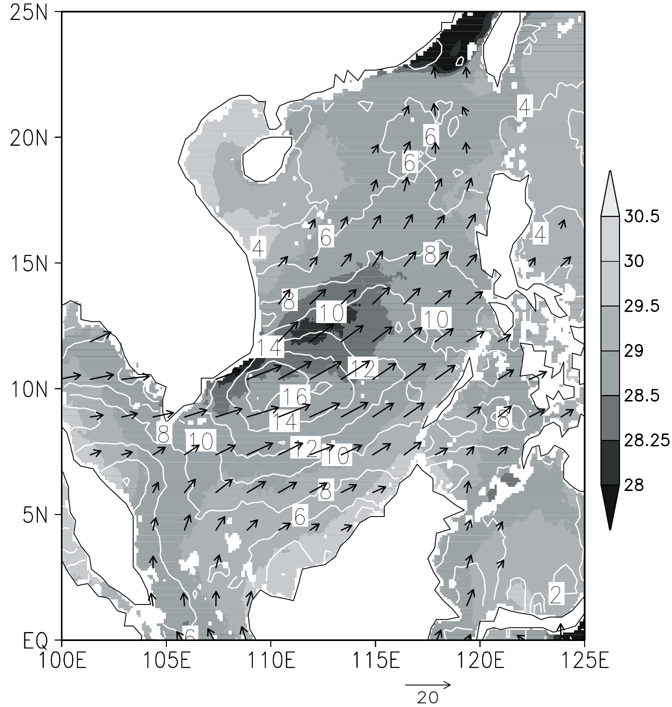

Figure 8.13. 1985–1999 SST climatology (shading in °C) along with QuikSCAT wind vectors and stress magnitude (contours in $10^{-2}\,\mathrm{N\,m^{-2}}$).
From Xie *et al.* (2003).

flows account for the daytime maximum of convection over hills and mountains, and inland of coastlines, as well as the nocturnal maxima in precipitation in mountain basins and just offshore the coastlines of Asia. An interesting example of the impact of mountain and valley flows on convection can be found in a recent study by Fujinami *et al.* (2004). Using geostationary meteorological satellite (GMS) IR data, they found that over the Tibetan Plateau convection is closely tied to two major east–west mountain ranges, indicated by the topographic cross section along 90°E in the left panel of Figure 8.14. A time–latitude plot of cloud cover frequency (right panel) shows clouds developing along these ranges around 09 UTC (15 LT) and then shifting to the valley between them by 13 UTC (19 LT). This shift is presumably a consequence of the development of drainage flow convergence into the valley in the evening augmented by downdraft outflows.

8.4 JETS

As previously noted, the topography of the monsoon regions often contributes to flow deflection or blocking and mesoscale low-level jets. The low-level jet through the

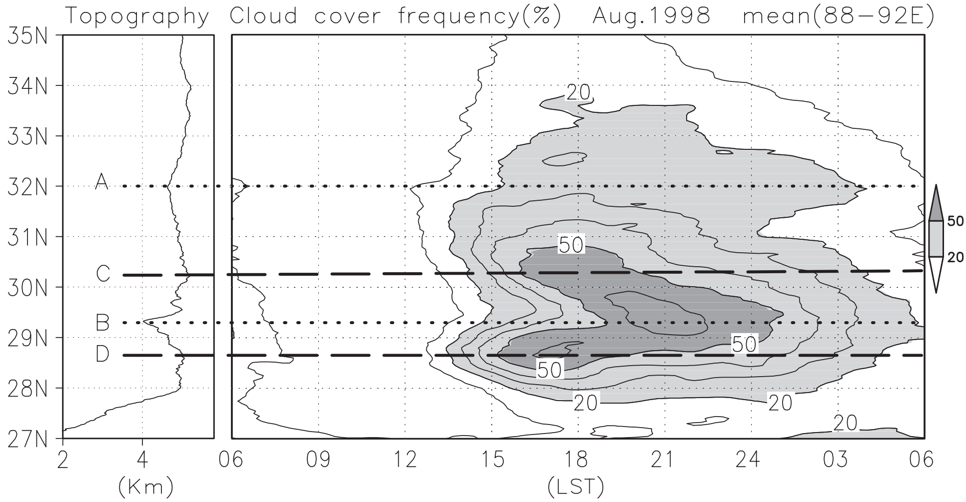

Figure 8.14. Latitude–time section of cloud cover frequency along 90°E for August 1998. Left panel indicates cross section of topography along 90°E.
From Fujinami *et al.* (2004).

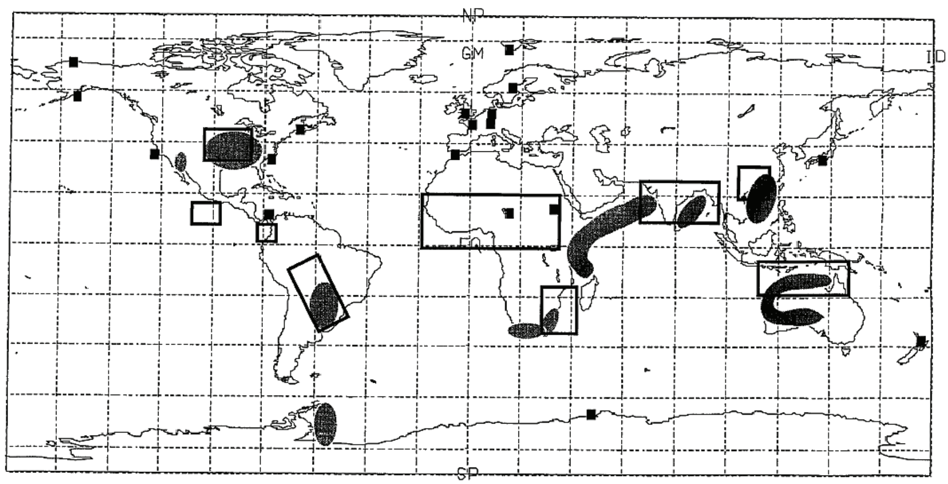

Figure 8.15. Regions where low-level jets are known or suspected to occur with some regularity (shaded) and where mesoscale convective complexes are known to occur frequently during the summer (open boxes). Squares denote locations where low-level jets have been occasionally observed.
From Stensrud (1996).

Taiwan Straits is but one example. However, there are other low-level jets of mesoscale and larger dimensions in the Asian monsoon. A map showing the global distribution of low-level jets is presented in Figure 8.15 (Stensrud, 1996). Within the tropical monsoon regions, low-level jets are observed over the Indian Ocean/Arabian Sea (the Somali jet), the Bay of Bengal, the South China Sea,

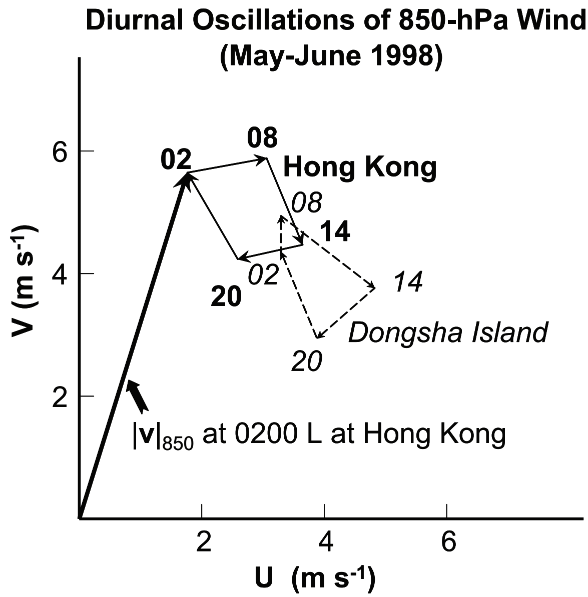

Figure 8.16. Diurnal wind oscillations at Hong Kong and Dongsha Island during SCSMEX.

Australia, and South America. Topography plays an important role in a number of these jets (e.g., the South American low-level jet occurs downstream of the Andes, the Somali jet is influenced by the east African mountains (Krishnamurti *et al.*, 1976)). Many of the areas of significant MCC activity are colocated with low-level jets, indicating the important role these jets play in transporting moisture into the convection thereby promoting large, long-lived systems (Maddox, 1983; Laing and Fritsch, 2000).

Low-level jets can also develop in response to boundary layer nocturnal cooling and an associated inertial oscillation, as observed in the African monsoon region (Blackadar, 1957) and elsewhere. Other mechanisms that can contribute to the formation of low-level jets are the diurnal heating cycle over sloping terrain (producing a diurnal oscillation in the low-level thermal wind), flow blocking by terrain, shallow baroclinic zones due to surface contrasts, and isallobaric forcing in connection with upper level jet streaks (see review by Stensrud (1996) for a detailed discussion of these mechanisms).

To illustrate the nocturnal low-level jet (LLJ), consider the findings from SCSMEX shown in Figure 8.16. An inertial oscillation is present as in Blackadar (1957) characterized by a clockwise turning of the wind, with maximum amplitudes at both Hong Kong and Dongsha Island at 08 LT. The amplitude of the ageostrophic wind oscillation ($\sim 1\,\mathrm{m\,s^{-1}}$), is considerably less than the $\sim 5\,\mathrm{m\,s^{-1}}$ found over the summertime central USA (Whiteman *et al.*, 1997), but is not insignificant. Over the USA the nocturnal LLJ has been linked with a nocturnal precipitation maximum in the Great Plains associated with eastward propagation of convective systems (e.g., Wallace, 1975; Carbone *et al.*, 2002). Similar low-level jets and nocturnal precipitation maxima have been found over South America by Virji (1981) and Velasco and Fritsch (1987); over southern China and Taiwan by Chen and Yu (1988) and Chen and Li (1995); and over Australia by Allen (1981).

In addition to low-level jets, meso-to-synoptic scale processes associated with upper level jets occur in monsoon regions. Keenan and Brody (1988) found that banded cloud structures in the Australian summer monsoon are associated with secondary circulations in the equatorial entrance region of a subtropical, 200-hPa trough. Chang and Lau (1980, 1982) found linkages between the transverse circulations associated with the east Asian jet streak and the strength of the local Hadley circulation and northern winter monsoon. Chang and Lum (1985) found that tropical convective activity during the northern winter monsoon can influence the strength of the mid-latitude jet. This coupling can often take the form of 'tropical plumes' (McGuirk *et al.*, 1988), which can amplify and spread poleward as a result of convectively generated inertial instability (Mecikalski and Tripoli, 1998). Blanchard *et al.* (1998) have identified inertial instability, often occurring on the anticyclonic side of jet streaks, as a mechanism for the upscale development of MCSs.

8.5 SURFACE–ATMOSPHERE INTERACTIONS

Throughout the monsoon regions of the World, surface exchanges represent important components of both the forcing of, and response to, the monsoon system (Ding, 1994). For example, studies have shown that strong surface sensible heat flux over the Tibetan Plateau during the spring helps set the stage for the onset of the Asian summer monsoon by heating the upper troposphere, thereby contributing to an eventual reversal in the north–south temperature gradient (e.g., Flohn, 1968; Luo and Yanai, 1984; Li and Yanai, 1996). After the summer rains begin, diabatic heating contributes further to this reversal and the overall energetics of the monsoon circulation.

Vigorous air–sea exchanges over the Arabian Sea and Indian Ocean have significant effects on the rainfall distribution over India as well as on the upper ocean. A prominent feature of the Indian summer monsoon is the abrupt cooling of the Arabian Sea following the onset of a strong south-westerly flow in June. This phenomenon was recently studied by Rudnick *et al.* (1997) using surface mooring data from the west-central portion of the Arabian Sea. A sudden onset of strong south-westerlies around June 1 is accompanied by a sharp drop in the SST and air temperature. The SST–air temperature difference decreases to near zero after onset and there is a period of upward net heat fluxes, primarily due to latent heat losses from the strong winds. The strong low-level jet over the Arabian Sea leads to a pattern of coastal upwelling (north of the jet axis) and downwelling (south of the jet axis) and an overall southward Ekman transport (Chapter 1). This upwelling contributes to the Arabian Sea cooling, and it also brings nutrient-rich water to the surface, supporting increased productivity in the upper ocean.

During the Asian winter monsoon, cold air often streams off the east coast of

Asia, leading to strong sensible heat fluxes over the bordering oceans. The boundary layer over the East China Sea during cold air outbreaks was sampled during the 1975 Air Mass Transformation Experiment (AMTEX). A visible satellite image of the cloud fields associated with a cold air outbreak over the East China Sea is shown in Figure 8.17 (color section). Narrow cloud lines are seen to expand in scale to closed cellular patterns downstream. Surface sensible heat fluxes in these cold air outbreaks can reach 1,200 $W m^{-2}$ (Agee, 1984), thus having a dramatic impact on the downstream circulation. Changes in the boundary layer accompanying cold surges were also sampled during the 1978 Winter Monsoon Experiment or WMONEX (Johnson and Zimmerman, 1986).

There are a number of complex mesoscale processes involving air–sea interactions during the monsoons. A remarkable example occurs over the East China Sea during winter (Figure 8.18 (color section), from Xie, 2004). As cold air streams off China in the winter, it cools the upper ocean. Shallower water cools more than deep water, so the SST field closely matches the bathymetry (Figure 8.18(a)). The SST field in turn influences the surface wind field. Warmer water enhances the vertical mixing, so stronger surface winds are observed over warmer waters, and a pattern of surface divergence and convergence is established (Figure 8.18(b)). Over the offshore area of convergence there is increased cloudiness, as indicated by the TRMM Microwave Imager (TMI) cloud liquid water shown in Figure 8.18(c).

8.6 GRAVITY CURRENTS AND GRAVITY WAVES

Gravity currents are horizontal flows generated by density differences in a fluid (Simpson, 1997). They commonly occur when a cool, thunderstorm downdraft reaches the Earth's surface and spreads out horizontally in a neutrally stratified boundary layer. The ensuing outflow or *gust front* often leads to the triggering of new convection throughout the Asian monsoon region. In the absence of low-level shear, the new convection occurs in arcs along the advancing outflow boundary, as illustrated in the upper left panel of Figure 8.4. When low-level shear is present, new convection is favored along the portion of the gust front that is perpendicular to the low-level shear vector (upper right panel of Figure 8.4).

If the atmosphere is stably stratified, the downdraft will generate a series of gravity waves rather than a gravity current (Haertel *et al.*, 2001). On the other hand, if the downdraft impinges on a stable layer near the surface, such as a nocturnal inversion, it can generate a *bore*. A bore consists of an increase in the depth of a fluid (or stable layer near the ground) with a series of waves on its surface that typically separate from the leading edge of the outflow and move ahead of it. Bores have the characteristic of a significant surface pressure rise and rapid propagation, such as has been observed for undular bores (called 'Morning Glories') over northern Australia in the Gulf of Carpenteria (e.g., Clarke *et al.*, 1981) and the Japan Kanto Plain (Ogura and Hoshino, 2001).

Cold fronts passing Taiwan display gravity current characteristics on the west side of the island, which impacts the properties of frontal convection there (Trier *et al.*, 1990). Farther to the south as the cold fronts (or cold surges in the winter monsoon) move equatorward, their gravity current characteristics diminish and their propagation is explained more by gravity wave dynamics (Chang *et al.*, 1983).

8.7 COASTALLY TRAPPED DISTURBANCES

The (northward) southward movement of cold air to the west (east) of mountain barriers in the northern hemisphere contributes to sudden changes in local weather conditions in monsoon regions. These disturbances often exhibit characteristics of coastally trapped gravity waves, Kelvin waves, or Rossby waves (Skamarock *et al.*, 1999), and their dynamics are still a subject of investigation. Douglas and Leal (2003) have used sounding data from the west coast of Mexico to demonstrate that Gulf of California surges during the south-west monsoon have characteristics of coastal gravity currents (coincident sharp wind shift, temperature drop, and pressure rise in the lowest kilometer). Cold fronts moving southward past Taiwan also exhibit properties of coastally trapped disturbances, with rapid propagation on the east side of the mountainous island, as illustrated in Figure 8.19 (Chen *et al.*, 2002). This rapid southward propagation along the east side of the barrier can be explained in terms of shallow-water equations. For a north–south barrier, the meridional component of the wind is given by $-fv = -g\,\partial h/\partial x$, where h is the free surface height. As the easterly geostrophic flow piles up cool air along the east coast of Taiwan, $\partial h/\partial x$

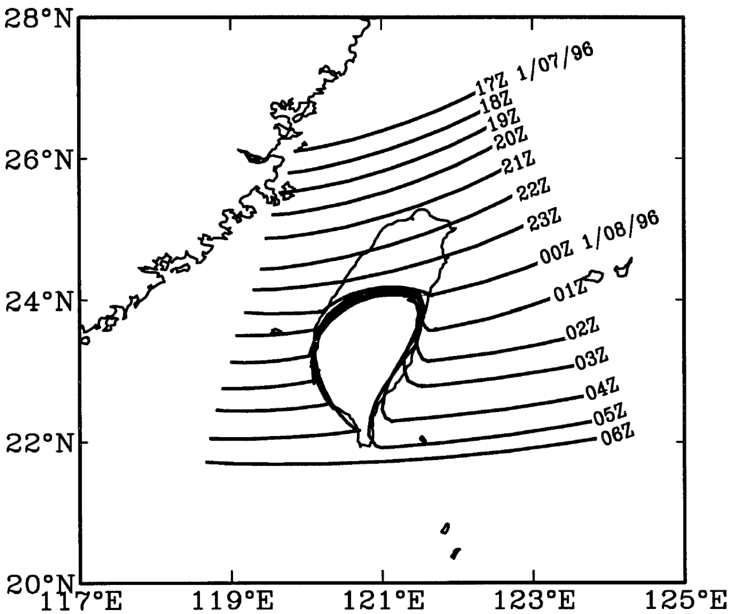

Figure 8.19. Locations of a cold front at hourly intervals on 7–8 January 1996.
From Chen *et al.* (2002).

becomes negative, yielding $v < 0$ or a northerly flow. Similar effects are seen along the east coast of China during the winter monsoon as cold fronts surge southward just along the coastline.

8.8 MESOSCALE INSTABILITIES

Numerous mesoscale instabilities exist throughout the monsoon regions, many of which influence cloud and precipitation patterns. In the boundary layer, shear-modified Rayleigh instability, inflection point instability, and Ekman layer instability may help explain the cloud streets and closed cell patterns seen in Figure 8.17. Kelvin–Helmholtz (vertical shear) instability is prevalent in many regions and accounts for closely spaced ($\sim$10 km) banded structures in precipitation systems in sheared environments. Larger scale banded structures in Mei-yu or Baiu precipitation systems which have some baroclinicity may be associated with conditional symmetric instability (Bennetts and Hoskins, 1979) or slantwise convection (Emanuel, 1983). Inertial instability has been invoked by Toyoda *et al.* (1999) to explain a series of anticyclonic vortices in a cloud band over Japan, where the instability is envisaged as arising from the vertical transport of momentum within deep convection. Cho and Chen (1995) have presented a theory for Mei-yu frontogenesis that invokes conditional instability of the second kind (CISK), arguing that in contrast to the traditional Ekman layer CISK theory, boundary layer convergence is partly induced by friction and partly by latent heat release in deep convection. To a large degree, the study of the role of mesoscale instabilities in Asian monsoon precipitation systems is still in its infancy.

8.9 SUMMARY AND OUTSTANDING ISSUES

In this chapter we have examined mesoscale processes that influence weather in the region of the Asian monsoon. Precipitation is the most important of these processes from a societal perspective, and it is also the most complex. There are a multitude of mesoscale processes that influence the location, intensity, and duration of precipitation. Field experiments in the past three decades in the tropics and monsoon regions have shed considerable light on the properties and mechanisms of deep convection. One of the most important findings is the tendency for convection to organize into cloud systems of several hundreds of kilometers in horizontal dimension (mesoscale convective systems or MCSs), such that they last considerably longer than the $\sim$30 min lifetime of individual cells. This organization has significant consequences for the distribution, intensity, and duration of precipitation, as well as the impact of convection on the large-scale flow. Various aspects of convection have been reviewed: the structure and properties of convective systems, cloud populations, latent heating profiles, convective momentum transports, and the diurnal cycle of convection.

In addition to convection, a wide range of other mesoscale processes influence the Asian monsoon. Among these are topographically forced local circulations (including flow blocking, sea and land breezes, and mountain and valley circulations), jets, surface–atmosphere interactions, gravity currents and gravity waves, coastally trapped disturbances, and mesoscale instabilities. Most of these phenomena are significantly modulated by the diurnal cycle. We have reviewed these processes and given examples that apply to the Asian monsoon. However, this list of mesoscale phenomena is not exhaustive and a number of phenomena and processes have unavoidably been omitted.

Despite progress in understanding the mechanisms of monsoon mesoscale phenomena, there remain a number of outstanding problems warranting further research, some of which are listed below:

- What determines the organization and propagation characteristics of mesoscale convective systems?
- What are the organizational patterns and life cycle characteristics of convective storms that produce flash floods?
- What roles do mesoscale convective vortices play in prolonged heavy rainfall in the Mei-yu/Baiu frontal zone?
- What are the mechanisms by which the low-level and upper level jets modify convective rainfall?
- How does convection interact with synoptic-scale disturbances (e.g., equatorial waves, monsoon depressions, etc.) to modify their development?
- What are the mechanisms for upstream development of convection along coastlines in monsoon regions?
- What processes account for the diurnal evolution and propagation of convection over oceans?
- What are the dynamics of coastally trapped disturbances in monsoon regions (e.g., coastal fronts during the winter monsoon)?
- How does the environmental flow affect the structure and organization of convective systems and their associated latent heating profiles?
- What roles do convective systems play in the onset and intraseasonal variability of the monsoon?

These are just a few of many questions pertaining to monsoon mesoscale processes which undoubtedly will motivate future observational, theoretical, and modeling studies of the Asian monsoon.

8.10 ACKNOWLEDGEMENTS

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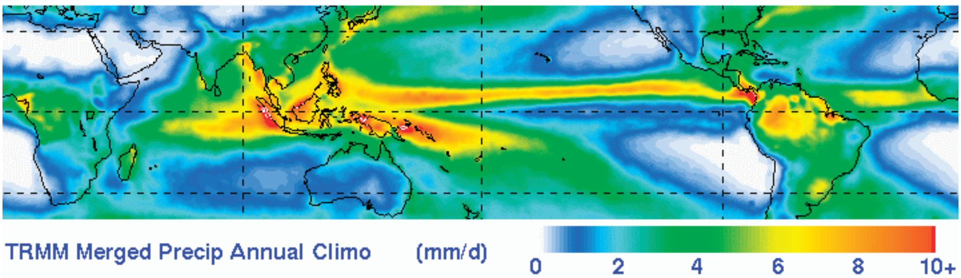

Figure 8.1. Six-year TRMM merged precipitation annual climatology (January 1998–December 2003).

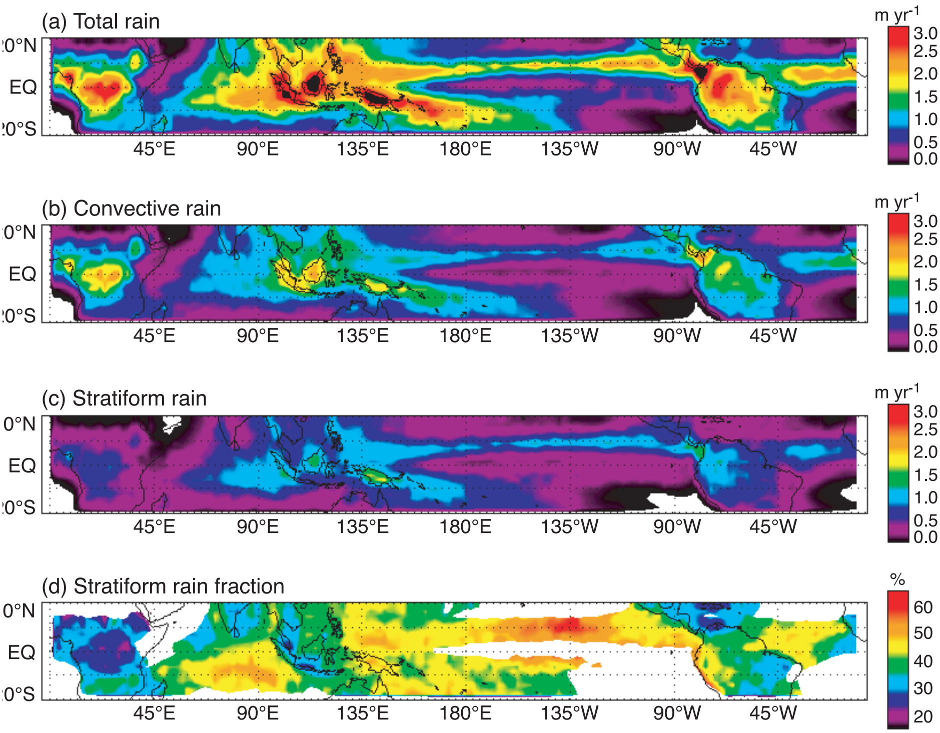

Figure 8.7. TRMM Precipitation Radar (PR) estimates of (a) total rain, (b) convective rain, (c) stratiform rain, and (d) stratiform rain fraction based on 2.5° grid averages for 1998–2000. From Schumacher and Houze (2003).

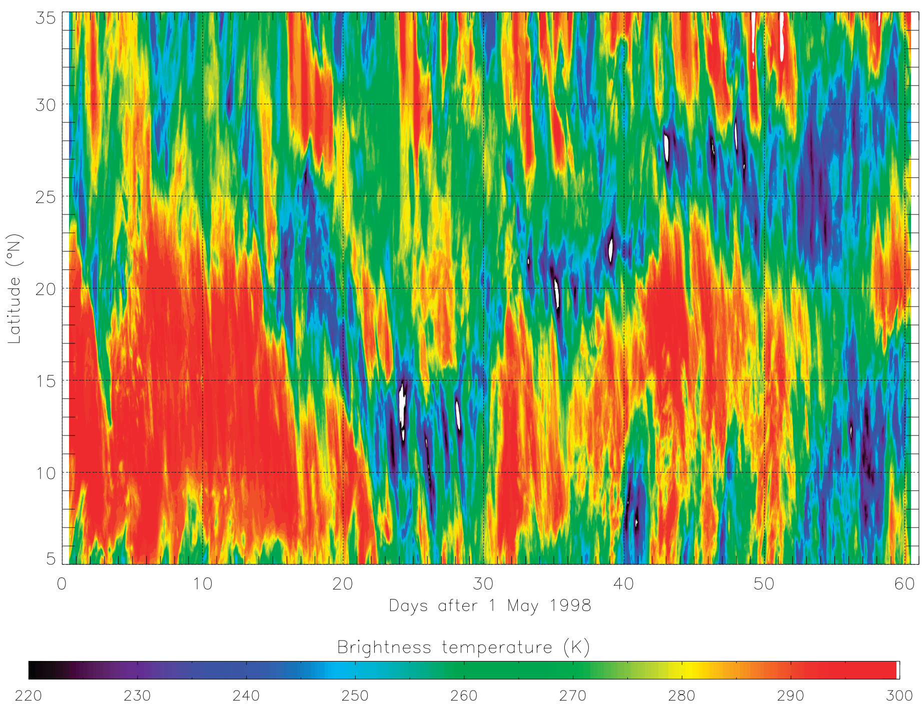

Figure 8.10. Time–latitude plot of IR brightness temperatures averaged over the South China Sea between 110°E and 120°E for 1 May to 30 June 1998.

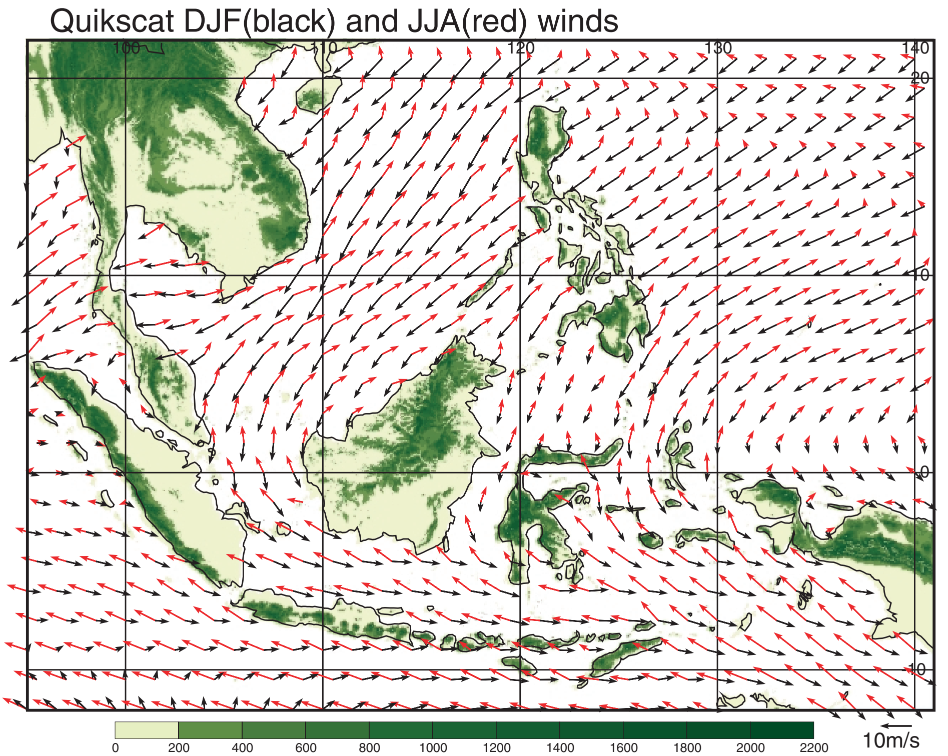

Figure 8.12. Mean QuikSCAT wind for DJF (black) and JJA (red). Topography in meters.
From Chang *et al.* (2005).

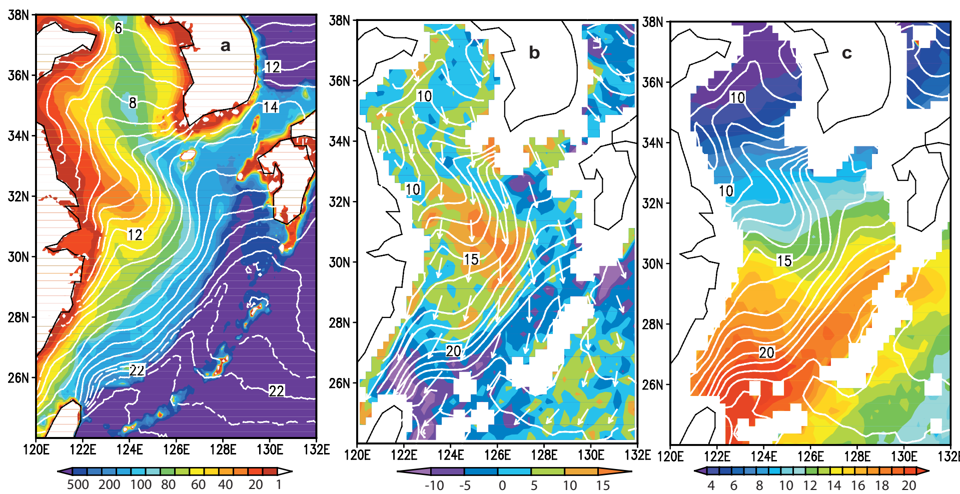

Figure 8.18. January–March SST climatology (contours in °C) over the Yellow and East China Seas, along with: (a) bottom depth (m); (b) velocity (vectors in m s^{-1}) and divergence (color in $10^{-6}\,s^{-1}$) of QuikSCAT wind; and (c) TMI cloud liquid water (10^{-2} mm). The QuikSCAT and TMI climatologies are January–March averages for 2000–2002.
From Xie (2004).

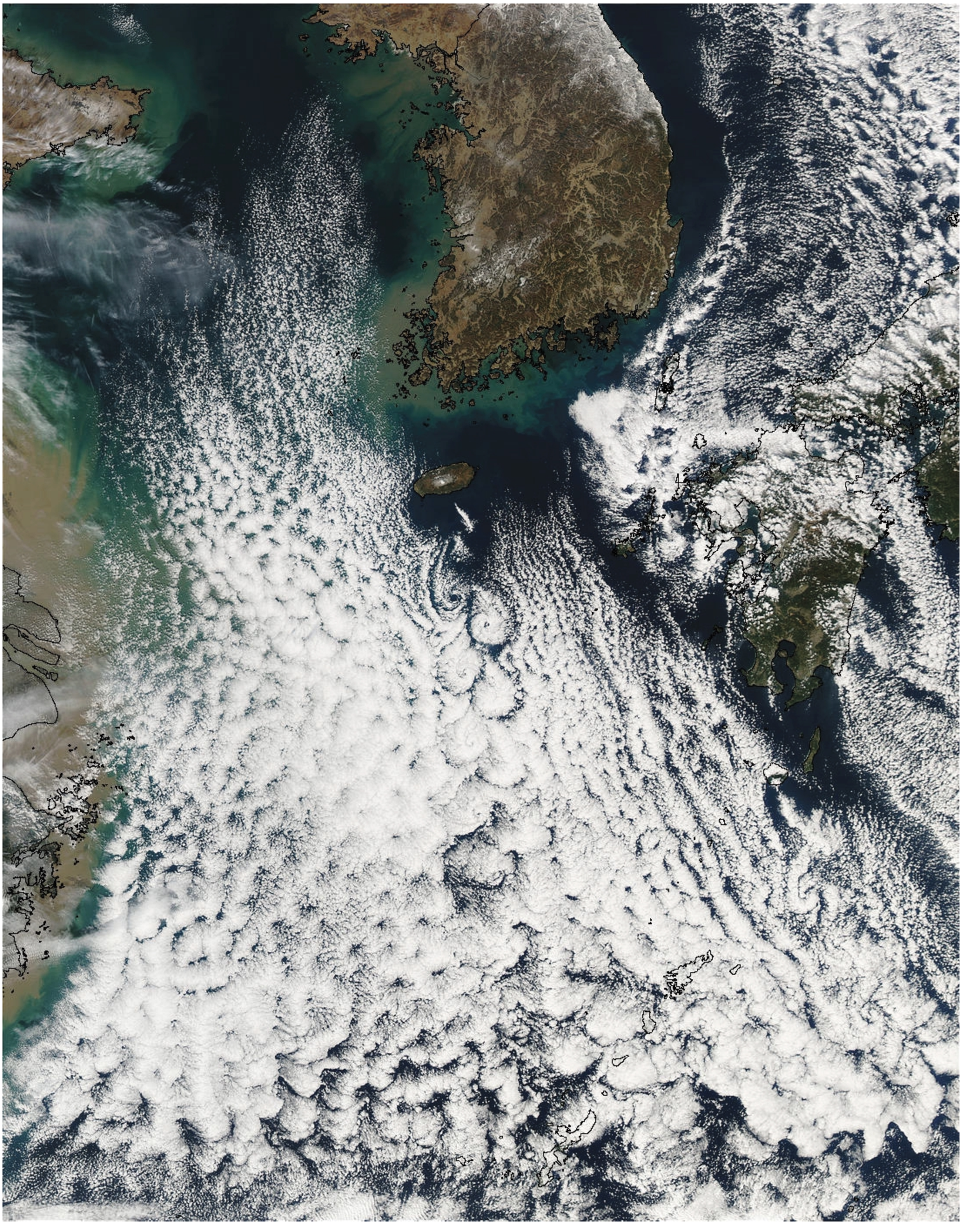

Figure 8.17. *Aqua* satellite image of cloud lines, closed cells, and vortex streets over the East China Sea on 15 January 2003 during a cold air outbreak.

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