Monsoon Convection in the Maritime Continent: Interaction of Large-Scale Motion and Complex Terrain

CHIH-PEI CHANG
Department of Atmospheric Sciences, National Taiwan University, Taipei, Taiwan

MONG-MING LU
Central Weather Bureau, Taipei, Taiwan

HOCK LIM
Department of Physics, National University of Singapore, Singapore

ABSTRACT

The Asian monsoon is a planetary-scale circulation system powered by the release of latent heat, but important features of deep convection and rainfall distribution cannot be adequately represented by the large-scale patterns. This is mainly due to the strong influences of terrain that are important across a wide range of horizontal scales, especially over the Maritime Continent where the complex terrain has a dominant effect on the behavior of convective rainfall during the boreal winter monsoon. This chapter is a review and summary of published results on the effects on monsoon convection due to interactions between the Maritime Continent terrain and large-scale transient systems.

The Maritime Continent topographic features strongly affect both the demarcation of the boreal summer and winter monsoon regimes and the asymmetric seasonal marches during the transition seasons. In the western part of the region, the complex interactions that lead to variability in deep convection are primarily controlled by the cold surges and the synoptic-scale Borneo vortex. The Madden–Julian oscillation (MJO) reduces the frequency of weaker surges through an interference with their structure. It also influences convection, particularly on the diurnal cycle and when synoptic activities are weak. When both surges and the Borneo vortex are present, interactions between these circulations with the terrain can cause the strongest convection, which has included Typhoon Vamei (2001), which is the only observed tropical cyclone that developed within 1.5° of the equator.

The cold surges are driven by midlatitude pressure rises associated with the movement of the Siberian high. Rapid strengthening of surge northeasterly winds can be explained as the tropical response via a geostrophic adjustment process to the pressure forcing in the form of an equatorial Rossby wave group. Dispersion of meridional modes leads to a northeast–southwest orientation that allows the surge to stream downstream through the similarly oriented South China Sea. This evolution leads to a cross-equatorial return flow and a cyclonic circulation at the equator, and thus a mechanism for equatorial cyclogenesis. Although the narrow width of the southern South China Sea facilitates strengthening of the cold surge, it also severely restricts the likelihood of cyclone development so that Vamei remains to be the only typhoon observed in the equatorial South China Sea.

Climate variations from El Niño–Southern Oscillation to climate change may impact the interactions between the large-scale motion and Maritime Continent terrain because they lead to changes in the mean flow. The thermodynamic effects on the interaction between MJO and the monsoon surges and Borneo vortex over the complex terrain also need to be addressed. These and other questions such as any possible changes in the likelihood of equatorial tropical cyclogenesis as a result of climate change are all important areas for future research.

Corresponding author address: Dr. Chih-Pei Chang, Department of Meteorology, Naval Postgraduate School, 589 Dyer Road, Room 254, Monterey, CA 93943.
E-mail: cpchang@nps.edu

DOI: 10.1175/AMSMONOGRAPHS-D-15-0011.1

© 2016 American Meteorological Society
1. Monsoon rainfall pattern and complex terrain

Availability of high-resolution, satellite-derived rainfall data from the Tropical Rainfall Measuring Mission (TRMM) since November 1997 has led to many studies of the interactions of the large-scale tropical convection in the monsoon regions and the terrain. Even mesoscale terrain features have a strong influence on the large-scale distribution of monsoon rainfall (Chang et al. 2004a, b, 2005b). The traditional use of area-averaged rainfall as monsoon indexes [e.g., the all-India monsoon rainfall index (Sontakke et al. 1993) that has been used for the Indian summer monsoon since the mid-nineteenth century] was based on the perspective of a large-scale pattern of the monsoon convection and rainfall that are comparable to the scale of the monsoon circulation. However, Chang et al. (2004a, 2005b) showed that such a large-scale perspective distorts the reality of deep convection and rainfall distributions because the latter are often strongly affected by the detailed terrain features throughout the monsoon region.

Among the monsoon systems around the globe, the Asian winter monsoon has the largest meridional domain that extends from the equatorial Maritime Continent and surrounding oceans to Northern Hemisphere midlatitude Siberia (Ramage 1968; Chang et al. 2006). In the midlatitude regions, the monsoon is characterized by strong baroclinicity dominated by the cold-core Siberian high, which is the strongest surface high pressure system in the world (Ding 1994). In the equatorial region, the deep convection around the Maritime Continent is the most vigorous and extensive large-scale convection system, and its effects reach far beyond the Asian monsoon region and can impact weather in North America (Yanai and Tomita 1998; Yang et al. 2002; Chan and Li 2004) and Europe (Neale and Slingo 2003). Strong interaction between the midlatitude and tropical components through cold surges and convection feedback to the East Asian jet stream (e.g., Chang and Lau 1982; Lau and Chang 1987; Neale and Slingo 2003) reinforces each circulation, which makes the Asian winter monsoon one of the most energetic planetary-scale circulation systems in Earth’s atmosphere (Chang et al. 2006). These interactions also play key roles in the interhemispheric exchanges carried out by the Asian–Australian monsoon system (e.g., Carrera and Gyakum 2003, 2007; Guan et al. 2010).

In the Maritime Continent region, significant geographical variations also exist in the seasonal march throughout the year, which has been recognized since Braak (1921, 1929; see Ramage 1971). A main reason for these variations is the complex terrain due to islands of different sizes interspersed within the surrounding seas. The complex terrain contributes to significant local and mesoscale circulations that interacts with the large-scale monsoon circulation and the diurnal land–sea and mountain breeze circulations, particularly during El Niño events when large-scale wind conditions are weaker in this region (Moron et al. 2010; Qian et al. 2010, 2013; Robertson et al. 2011).

The importance of smaller-scale terrain influences on seasonal-scale convection can be seen in Fig. 6-1, which shows the partition of the summer and winter monsoon regimes in Southeast Asia based on the TRMM Precipitation Radar (PR) data (Simpson et al. 1996) and the QuikSCAT scatterometer winds that describe winds at...
10 m above the sea surface (Liu 2002). The boreal summer and winter monsoon regimes are defined by the difference in TRMM PR rainfall and QuikSCAT winds between December–February (DJF) and June–August (JJA). Rainfall difference in warm colors means the boreal summer rainfall exceeds the boreal winter monsoon; thus it denotes the summer monsoon regime. Rainfall difference in cool colors denotes the winter monsoon regime. The general pattern reflects the mean seasonal differences that the boreal summer rainfall regime dominates north of the equator, and the boreal winter rainfall regime dominates south of the equator.

While the two regimes commingling near the equator, there is a conspicuous pattern of asymmetry in the degree of intrusion from one regime to the other regime. The boreal winter regime extends far northward into the boreal summer regime, whereas the southward extension of the boreal summer regime south of the equator is much more limited. Two effects contribute to this asymmetry: The mesoscale terrain features and the equatorward surface winds, both of which are more dominant north of the equator. Over Southeast Asia, the intrusions of the winter regime beyond 5°N occur in the following areas: east of the Philippines, northeast and northwest of Borneo in the South China Sea, east of Vietnam, the eastern coast of the Malay Peninsula, and north of Sumatra. In most of these areas the high boreal winter rainfall is due to the onshore northeasterly winter monsoon winds from the northwest Pacific and the South China Sea. These northeasterly monsoon winds are stronger than the prevailing southeasterly winds in the Southern Hemisphere because of the intense baroclinicity over the cold Asian continent during boreal winter that is significantly stronger than the baroclinicity in the Southern Hemisphere during boreal summer. There are also very few coastal areas between 5° and 10°S that face the prevailing seasonal wind, as is the case in the northern tropics.

The northeast monsoon wind is parallel to the coastline over northwest Borneo so there is little onshore wind, but the intrusion of the boreal winter regime here is also the result of the terrain effect. Over this region a quasi-stationary synoptic-scale cyclonic circulation that is associated with the heating of the island can develop in any season of the year. During boreal winter, the low-level basic-state background vorticity is also cyclonic because of both the mean northeasterly wind maximum over the South China Sea and the equatorial westerlies associated with the East Asian winter monsoon. Perturbations in this basic state can also amplify into synoptic-scale cyclonic circulations. These disturbances are often found southeast of the primary region of cold surge northeasterly winds. As a combination of these conditions, the low-level cyclonic circulation is particularly active during winter and is a prominent feature of the boreal winter climatology (Johnson and Houze 1987, see their Fig. 10.2). Although the circulation may not be completely closed on the east side over the island, it is often referred to as the Borneo vortex and is often associated with deep convection and intense latent heat release (Cheang 1977; Lau and Chang 1987; Johnson and Houze 1987; Chang et al. 2003; Trilaksono et al. 2012). Because of the increased frequency of the Borneo vortex, the deep convection and heavy rainfall frequently extend offshore several hundred kilometers into the South China Sea during boreal winter, but much less so during boreal summer.

The effect of wind–terrain interaction for the summer monsoon regime is also clearly demonstrated in Fig. 6-1, in which the reversal of the plotted wind vectors indicates the southwest monsoon wind directions north of the equator. Here the strongest monsoonal rainfall occurs along the windward side of mountains in northwestern Philippines, northern and central Vietnam, the southwest coast of Cambodia, and the west coast of Myanmar northeast of the Bay of Bengal. It has been widely accepted that monsoon depressions in the Bay of Bengal are a major contributor to the summer monsoon rainfall in India (e.g., Krishnamurthy and Ajayamohan 2010). However, Fig. 6-1 shows that the heaviest summer monsoonal rainfall in the eastern part of the Indian monsoon region is not along the main path of the Bay of Bengal monsoon depressions that traverses the bay from southeast to northwest, but is concentrated on the northeastern part of the bay away from the Indian subcontinent. This is where the southwest monsoon winds interact with the coastal mountains near the northeastern shoreline of the bay. Thus, the wind–terrain interaction also dominates the distribution of the summer monsoon rainfall and has a stronger effect in producing the monsoon rainfall than the monsoon depressions.

2. Asymmetric seasonal transitions and terrain effects

The boreal spring [March–May (MAM)] and fall monsoon regimes [September–November (SON)] may also be delineated from the TRMM Precipitation Radar and QuikSCAT scatterometer wind data. Because these are during transitional seasons, the analysis needs be done differently from the method used to partition boreal summer and winter monsoon regimes. The spring and fall regimes are defined by Chang et al. (2005b) as the areas that have higher mean rainfall than all the other seasons. Thus, the graphic representation of these two regimes is not derived from a simple seasonal
difference as shown in Fig. 6-1, but rather it is derived with two steps. In the first step, each grid point where the MAM rainfall (cool color) is the maximum among the four seasons is identified, and the difference between the MAM rainfall and the higher of the summer or winter rainfall is computed. The same procedure is followed at points where the SON rainfall (warm color) is the maximum of all four seasons. The resultant values are then plotted together in Fig. 6-2. The total QuikSCAT winds for both seasons are also plotted, with the MAM wind in black and the SON wind in red.

In Fig. 6-2 the SON monsoon rainfall dominates areas north of the equator and the western side of the domain. The MAM monsoon rainfall dominates areas south of the equator and the eastern side of the domain. Within the equatorial belt of 10°S–10°N, the two regimes are roughly divided by Borneo. The largest signals of the MAM monsoon rainfall are along windward coastlines in eastern Philippines, eastern Vietnam, and the northeastern Malay Peninsula, clearly indicative of the importance of local wind–terrain interaction. This is in contrast to the Southern Hemisphere MAM regime, where comparable significant monsoon rainfall does not exist. The contrast of the two transition regimes is consistent with the asymmetric seasonal march that has been noticed by many researchers. During boreal fall, the maximum convection moves southeastward from South Asia in a track that roughly follows the Southeast Asia land bridge, to reach southern Indonesia and northern and eastern Australia during boreal winter. During boreal spring, the maximum rainfall remains mostly south of the equator without a northwestward progression that would retrace the path of the boreal fall progression (e.g., Lau and Chan 1983; Meehl 1987; Yasunari 1991; Matsumoto 1992; Matsumoto and Murakami 2002; LinHo and Wang 2002; Hung and Yanai 2004; Hung et al. 2004).

Chang et al. (2005b) and Wang and Chang (2008) hypothesized that at least a part of this spring–fall asymmetry is due to the sea level pressure (SLP) differences between land and ocean, which are driven by the different thermal memories of the ocean and atmosphere–ocean interactions. They computed the difference between boreal spring and fall SLP and showed that in the Northern Hemisphere the land SLP is higher in boreal fall and the ocean SLP is higher in boreal spring. The reverse is true in the Southern Hemisphere. This pattern is seen throughout most of the global domain other than areas of the midlatitude storm tracks. The largest SLP difference occurs over the Asian continent, leading to the stronger SON northeasterly winds in the northern South China Sea and northeastern Pacific (area A in Fig. 6-3), which causes the deep convection east of Vietnam and the Philippines in boreal fall (Fig. 6-2). The convection in these areas is much stronger than in both boreal summer and winter, even though during boreal winter the northeasterly winds are stronger. This is because the colder and drier air and the
colder sea surface temperature (SST) make deep convection less likely to develop north of 10°N, so that strong winter cold surges actually produce drying conditions in northern and middle South China Sea (Chang et al. 2005a). In the southern South China Sea, the SLP gradient favors cyclonic flow and therefore deep convection in SON (area B in Fig. 6-3). The convection is further enhanced by the interaction of the wind with terrain on the east coast of Sumatra and west coast of Borneo.

The enhanced boreal fall convection in the eastern equatorial Indian Ocean may also be explained, at least partly, by the spring–fall SLP difference. The Bay of Bengal has higher SLP in boreal fall than in boreal spring, which suggests that in addition to the different thermal memory effect, atmosphere–ocean interaction is involved to warm the SST faster during boreal spring. One possibility is that in early spring the initially cool SST and more anticyclonic flow with weak winds cause less evaporation and more solar heating of the sea surface and downwelling in the upper ocean, so the spring SST becomes higher and the SLP becomes lower than in fall. But the land–sea redistribution of mass still contributes to lower SLP in the Bay of Bengal during boreal fall when compared to surrounding areas. The resulting difference in pressure gradient during boreal fall gives rise to cyclonic flow in the Bay of Bengal and favors increased cross-equatorial flow from the southern Indian Ocean.

The convection in and around the middle and southern South China Sea in boreal fall helps to induce southwesterly winds west of Sumatra (area C in Fig. 6-3). These southwesterly winds are enhanced by the tendency of cross-equatorial flow and the cyclonic flow in the Bay of Bengal. Other atmospheric and oceanic factors, such as the east–west pressure gradient across the equatorial Indian Ocean, may also contribute to the development of equatorial westerly winds. These winds have two effects, both of which lead to more convection. The first is the onshore flow that causes convergence along the western coasts of northern Sumatra and the Malay Peninsula. The second is the beta effect that produces convergence in the equatorial westerlies. The increased convection may further enhance the westerlies making a positive feedback possible.

South of the equator, the SLP difference between Australia and the south Indian Ocean favors counterclockwise flow toward the equator (area D in Fig. 6-3) in boreal fall. This also enhances the clockwise cross-equatorial flow that turns westerly north of the equator, which increases the wind–terrain interactions on the west coast of land areas, such as western Borneo (area B).

The combined effect of the condition favoring southerly winds from south of the equator and northeasterly winds in the northern South China Sea and the northwestern Pacific (area A) gives rise to a broadscale belt of tendency for convergence between the equator and 20°N during boreal fall (marked by the elliptic-shaped area in Fig. 6-3). This tendency of low-level convergence favors the development of deep convection. During boreal spring, the tendency of the wind directions changes sign, and this belt becomes an area with a tendency of low-level divergence, so convection tends to be suppressed. These effects are the results of the global-scale mass redistribution between the land and ocean regions that is driven by their different thermal memories. Because of the orientation of the Asian and Australian landmasses, this redistribution facilitates the southeastward march of maximum convection from the Asian summer monsoon to the Asian winter (Australian summer) monsoon, but it deters the reverse march in boreal spring. Thus, there is an inherent tendency for the onset of the Asian summer monsoon in spring to be more abrupt and difficult to predict than the more gradual development of the Asian winter monsoon.

3. Synoptic and intraseasonal disturbances during boreal winter

During the East Asian winter monsoon, the mid-latitude circulation exerts direct impacts on tropical weather through periodical and rapid strengthening of the low-level northeasterlies into the South China Sea and Maritime Continent regions following the south-eastward movement of the surface high pressure (e.g.,
The movement of the midlatitude circulation involves basically the advection of relative and planetary vorticity, but the nearly spontaneous freshening of the low-level northeasterly winds in the tropics and their large cross-isobar angles indicate a rapid progression of events on a gravity wave time scale, considerably shorter than that of the advective scale inherent in Rossby wave dynamics (Chang et al. 1983).

These processes are often referred to as “cold surges” (Ramage 1971; Compo et al. 1999; Garreaud 2001), although the rapid drop of surface temperature associated with these events rapidly diminishes over the warm waters of the South China Sea. The term “pressure surges” has also been used to identify the cold air outbreaks. This term describes well the outbreak of cold air eastward near the southern China coast (Chan and Li 2004), but it does not adequately describe the southward surges since south of about 10°N the surges are mainly manifest by the freshening of northeasterly winds rather than pressure rises. The outbreaks of cold air are almost always associated with movement of the Siberian high, which is a strong and semistationary cold-core high pressure during the boreal winter with a maximum central sea level pressure exceeding any other pressure system in Earth’s atmosphere (Ding 1994). There is no clear evidence of a relationship between the Siberian high’s interannual variation and circulation indexes such as El Niño–Southern Oscillation (ENSO) and North Atlantic Oscillation (NAO)/Arctic Oscillation (AO), but its decadal-scale variations may be related to NAO/AO (Gong et al. 2001; Wu and Wang 2002; Gong and Ho 2004; Chang et al. 2006, 2011). Its intraseasonal variation is often forced by midlatitude wave activity and can be influenced by blocking activities both to the west in the Atlantic and Ural areas and to the east in the Pacific (Takaya and Nakamura 2005a,b; Chang and Lu 2012). It can also move away from its normal position toward the western Pacific once to several times a month with durations of a few days to one week or more (e.g., Gong and Ho 2004; Lu and Chang 2009; Park et al. 2011).

As the Siberian high moves south and eastward, the cold surge winds spread equatorward around the eastern edge of low-level anticyclones and can cover a broad longitudinal span of the entire tropical western North Pacific. However, the strongest cold surges are concentrated in the South China Sea where they can reach and cross the equator. This is because the orientation of the regional topography (Fig. 6-4) acts to restrict the flow such that the low-level northeasterly flow is channeled toward the equator. Although cold surge winds are typically dry, they are moistened by the overwater trajectory (Johnson and Houze 1987; Takahashi 2012) and have been associated with enhanced upper-tropospheric outflow over the Maritime Continent, which is related to an enhanced east Asian local Hadley cell that may strengthen the East Asian jet and lead to further interactions with the midlatitude systems (Chang and Lau 1982; Lau and Chang 1987). Furthermore, the gradient of planetary vorticity together with the blocking and deflection due to topographic influences contribute to an eastward turn in the winds as they cross the equator.

The strong surges can cause heavy rainfall and are often associated with severe flooding in the equatorial zone particularly over the Malay Peninsula, Sumatra, Borneo, and other Indonesian islands (Johnson and Chang 2007; Tangang et al. 2008; S. Y. Lim et al. 2013, meeting presentation). Trilaksono et al. (2012) conducted regional model simulation of heavy precipitation events at Jakarta in West Java and found the February 2007 Jakarta flood, the worst flood event in three centuries, was associated with a strong cold surge that possessed a cold anomaly of 2°K below normal up to 1.5-km height, which is very unusual south of the equator. The cross-equatorial flow can also act to enhance the Australian monsoon trough and may contribute to tropical cyclogenesis (Holland 1984; McBride 1995).

The cold surges can even have remote influences on the tropical convection and disturbed weather away of the South China Sea. Vissa et al. (2013) documented the crucial role of the South China Sea cold surges in the development of Cyclone Sidr in the Bay of Bengal in November 2007. Cyclone Sidr was one of the most devastating tropical cyclones in the twenty-first century.
that caused several thousand deaths and tremendous property damage after landfall over the Bangladesh coast (Ákter and Tsuboki 2012). The development of the cyclone was traced back by Vissa et al. (2013) to cold surge–induced heavy rainfall episodes at the southern Gulf of Tonkin coast. The southward progression of the surge led to the intensification of deep convection on the Vietnamese coast and interaction with Typhoon Peipah, which entered the South China Sea from the western Pacific. Typhoon Peipah transported convective cloud clusters, moisture, and westward momentum to enhance the deep convection cells over the Vietnamese coast, which then moved eastward to the Gulf of Thailand and Andaman Sea where it organized into a tropical depression that later intensified to become Cyclone Sidr in the Bay of Bengal.

In addition to the synoptic-scale cold surges, the tropical convection weather in the South China Sea and the western Maritime Continent region is also affected by local sea surface temperature (Hendon 2003; Koseki et al. 2013a) and the other large-scale disturbances including the Borneo vortex and the Madden–Julian oscillation (MJO). The MJO often has peak amplitude over the Maritime Continent during boreal winter (Madden and Julian 1972), and its convection can also strengthen the local Hadley cell and the East Asian jet (Jeong et al. 2008; He et al. 2011). The MJO causes alternating periods of large-scale active and inactive convective phases with a periodicity of 30–60 days as it propagates eastward through the region (Chang et al. 2005a, 2006; Zhang 2005; Robertson et al. 2011). Thus, the frequency of the MJO is much lower than the cold surges that typically occur more than once per month, each time lasting from a few days to one week or more. Therefore, the signal of the impact on the variation of convection weather by MJO during boreal winter is less conspicuous compared to that due to the surges. Furthermore, the complex terrain causes the large-scale structure of MJO to disintegrate (Wu and Hsu 2009) so that the concept of MJO being a large-scale precipitation envelope that smoothly propagates eastward breaks down over the Maritime Continent (Peatman et al. 2013). The MJO-associated precipitation is strongest when the synoptic systems are weak (Qian et al. 2010, 2013; S. Y. Lim et al. 2013, meeting presentation), and it is mainly identified in the diurnal cycles (Qian et al. 2010, 2013; Kanamori et al. 2013; Virts et al. 2013; Peatman et al. 2013) and over ocean (Rauniyar and Walsh 2011; Oh et al. 2012).

Around the region of Borneo a low pressure area is commonly present year-round and has been variously described as an equatorial trough or an intertropical convergence zone (ITCZ). However, the Borneo vortex is an important system of the Asian winter monsoon with a climatological property that is distinct from other vortices in the tropics. This vortex arises as a result of the interaction of the basic-state shear vorticity created by the low-level northeasterly winds over the South China Sea and the much weaker winds over the mountainous western coast of Borneo, and diabatic heating supported by abundant moisture (Chang et al. 2004a; Ooi et al. 2011; Koseki et al. 2013b). The frequency of occurrence of vortices during boreal winter in the Borneo region exceeds that of any other seasons and regions, including those that are well known for seasonal vortex activities such as the Bay of Bengal during boreal summer (Chang et al. 2004a). The variability and life cycle of the Borneo vortex has an important impact on deep convection and disturbed weather throughout the equatorial South China Sea that is comparable to those due to the cold surges (Chang et al. 2006).

The larger spatial and time scales of the MJO may lead to a conception that MJO modulates the synoptic and smaller-scale convection in this region to the degree that its wet phase may be correlated significantly to the occurrence of high-impact weather. While this depiction may provide a reasonable overview of the large-scale convection during other seasons, it is too simplistic during boreal winter because the three circulation systems differ greatly in their origin. The MJO originates over the equatorial Indian Ocean, cold surges originate from the midlatitude regions of eastern Asia, and the Borneo vortex develops locally over the southern South China Sea. These three circulation systems interact in a rather complicated way, resulting in a profound impact on the variability in deep convection over the western Maritime Continent/equatorial South China Sea region during the Asian winter monsoon.

The discussion below is based on Chang et al.’s (2005b) analysis of the interaction of the three motion systems in this region (inside rectangle in Fig. 6-4). The analysis was conducted by compositing the large-scale patterns of NCEP–NCAR reanalysis 925-hPa winds and a convection index (CI) that measures the difference between the Geostationary Meteorological Satellite (GMS) blackbody temperature and a threshold of 250°C, with positive CI indicating a lower blackbody temperature. It will become apparent that the deep convection resulting from the three disturbance systems is strongly affected by the local terrain.

a. Mean convection and low-level kinematic fields in the western Maritime Continent region

Figure 6-5 shows the seasonal mean fields during December–February. In general, deep convection (Fig. 6-5a) is concentrated over the large islands of the region. The maximum convection occurs over Java,
which is connected to a second convection maximum over Sumatra with an extension into the eastern Indian Ocean as part of the ITCZ south of the equator. The lack of deep convection north of 5°N is consistent with the 925-hPa divergence (Fig. 6-5b) in the northeasterly monsoon flow. This is a significant change from boreal fall when strong convection occurs off the Vietnamese coast north of 10°N as a result of the low-level convergence produced by the northeast onshore winds, as discussed in section 2. In winter the cooler and drier mean northeast winds and the lower SST in the northern and middle South China Sea suppresses deep convection until the air reaches the southern South China Sea, where it is transformed by substantial surface sensible and latent heat fluxes (Johnson and Zimmerman 1986).

The pattern of the 925-hPa winds is largely a result of the blockings and deflections of the terrain around the South China Sea. The terrain effects and the conservation of potential vorticity cause a counterclockwise turning of the cross-equatorial winds. The two primary regions of low-level convergence (Fig. 6-5b) are associated with the maxima in deep convection (Fig. 6-5a). The convergence center near Sumatra is related to the interaction of the northeasterly monsoon flow with the terrain of the Malay Peninsula. However, the convergence center over Borneo is shifted east of the primary northeasterly wind belt and is in the region of the counterclockwise turning of the winds that cross the equator. Coincident with the convergence center over Borneo is a maximum in 925-hPa relative vorticity (Fig. 6-2c). In addition to the curvature contribution to the vorticity maximum, shear vorticity results because of the interaction between the northeasterly monsoon flow and the terrain of Borneo.

Figure 6-6 shows that the Sumatra and Borneo convection centers start the season forming a V-shape convection pattern in December, but both areas shrink and retreat southward and become separated in February. This retreat coincides with the intensification of the divergence over the South China Sea associated with the strengthening of the northeast monsoon winds. In particular, the reduction in deep convection over Sumatra is related with the increased drying influence of the strengthening northeast winds that flow toward Sumatra. The drying does not impact the convection center over Borneo as significantly since the northeast monsoon flow is nearly parallel to the Borneo coastline (Fig. 6-4), where Ekman pumping can increase moisture convergence as the strengthening northeast winds contribute to the shear and curvature vorticity of the counterclockwise circulation around Borneo. Between 2° and 4°N a substantial part of the western Borneo coastline faces northward, and as a result the blocking of strengthened northeast winds causes a direct increase in moisture convergence.

b. Cold surge and Borneo vortex

The synoptic and intraseasonal variation of deep convection in the southern South China Sea and western Maritime Continent can be described by the composites of low-level circulation and convection fields according to the activities of Borneo vortex, cold surge, and MJO. Of the total 1895 days in the 21 boreal winter seasons between 1979/80 and 2001/02, Chang et al. (2005a) identified nearly one-third as “vortex cases” and the remaining two-thirds as “no-vortex cases” based on whether one or more vortex centers appeared in the daily 0000 UTC 925-hPa streamline analysis between 5°S and 10°N and 105° and 115°E (inside rectangle in Fig. 6-4). They also identified one-fifth of the days as “surge” days when a South China Sea surge index, defined as the area-averaged northeasterly 925-hPa wind

---

**Fig. 6-5.** DJF (1979/80–2000/01) mean fields of (a) CI, (b) 925-hPa winds (m s⁻¹) and divergence (shaded, 10⁻⁵ s⁻¹), and (c) 925-hPa winds (m s⁻¹) and vorticity (shaded, 10⁻⁵ s⁻¹) (from Chang et al. 2005a).
over 5°–10°N, 107°–115°E (top thick bar in Fig. 6-4), reaches 8 m s\(^{-1}\). About 24% of the vortex days are also surge days, while only about 18% of the no-vortex days are. The frequency of both is highest in December and lowest in February.

The selection of vortex and surge days were done with unfiltered data, but in the composite of the two synoptic-scale systems the 925-hPa wind was filtered to highlight variations over the period range of 2–15 days. During no-vortex days, deep convection and low-level convergence are reduced over the southern South China Sea and enhanced to the west, southwest, and south of the South China Sea (Figs. 6-7a,c). During vortex days (Figs. 6-7b,d), the patterns are nearly identically opposite. Thus the presence of the Borneo vortex and deep convection acts to intercept transport of low-level moisture by the northeasterly monsoon flow such that convection downstream over the Malay Peninsula–Sumatra–Java region is reduced.

The composite convective index and divergence patterns for the surge and no-surge days exhibit well-recognized patterns of variability. During no-surge days, deep convection (Fig. 6-8a) and 925-hPa convergence (Fig. 6-8c) are located over Indochina, while reduced convection and low-level divergence are found over most of the remaining equatorial South China Sea region. During surge days, a near opposite pattern occurs with reduced convection (Fig. 6-8b) and low-level divergence over Indochina and enhanced convection and convergence over the remainder of the region (Fig. 6-8d), which is in agreement with previous studies of the influence of northeasterly surges on deep convection over the equatorial South China Sea (e.g., Lau and Chang 1987). During these periods, deep convection occurs in association with the blocking of the low-level winds by the terrain, which contributes to a shift of the convection pattern downstream of the maximum in 925-hPa convergence. The shift is less when the convection pattern is compared with 850-hPa convergence because there is less blocking of winds at higher elevations. The low-level convergence during surge days (Fig. 6-8d) takes on a V-shaped pattern on the windward side of the Malay Peninsula and Borneo. This pattern results from the blocking of the surge winds by the terrain (Fig. 6-4) and its location is different from the V-shaped divergence pattern associated with the monthly mean fields (Fig. 6-6), where maximum convergence centers are located over land areas.

Fig. 6-6. Individual monthly means (1979/80–2000/01) for (a)–(c) CI and (d)–(f) 925-hPa winds (m s\(^{-1}\)) and divergence (shaded, 10\(^{-5}\) s\(^{-1}\)) for December in (a) and (d), January in (b) and (e), and February in (c) and (f) (from Chang et al. 2005a).
Figure 6-9 shows the composites when the surge and vortex classifications are considered together. When neither surge nor vortex is present (Fig. 6-9a), convection is reduced over the equatorial South China Sea and enhanced downstream of the mean northeasterly wind over the Malay Peninsula, Sumatra, and Java, where the monsoon flow that has received the ocean surface sensible and latent heat fluxes interacts with terrain. This pattern is almost reversed when a vortex is present without a cold surge (Fig. 6-9b). The presence of the Borneo vortex results in a deflection of the low-level winds and convergence to the west coast of Borneo, such that the primary area of deep convection occurs more upstream near the west coast of Borneo and over the southern South China Sea. Convection over the landmasses downstream from the surge is suppressed. The presence of a cold surge basically enhances these two opposite patterns (Figs. 6-9c,d). The locations of convection during surge cases with and without a vortex are similar to those in the respective no-surge cases; however, the magnitude is much stronger with a surge. Convection over the southern South China Sea is strongest when both surge and vortex cases are present (Fig. 6-9d).

The enhancement of deep convection over the southern South China Sea during cases when both cold surge and Borneo vortex are present is sensitive to the strength of the cold surge (Fig. 6-10). As the intensity of the cold surge increases from weak (between 8 and 10 m s\(^{-1}\), Figs. 6-10a,d) to moderate (10–12 m s\(^{-1}\), Figs. 6-10b,e) to strong (greater than 12 m s\(^{-1}\), Figs. 6-10c,f), the area covered by increased convective index values and the amplitude of the convective index increase over the southern South China Sea. The increased deep
convection with surge intensity results from two processes, both of which involve terrain influences. First, increased northeast winds result in increased moisture convergence near the coastal area of Borneo. Second, the northeast winds strengthen much more over sea surface than over land, resulting in increased shear vorticity that contributes to a stronger Borneo vortex.

The impact of the presence of the vortex during strong surge events (Figs. 6-10b,e and 6-10d,f) can be examined by contrasting convection, wind, and divergence patterns of these surge and vortex cases with cases of strong surges but no vortex (Fig. 6-11). Without the presence of the vortex, deep convection throughout the southern South China Sea and along the Borneo coastline is severely reduced (Figs. 6-11a,b). However, convection over the Malay Peninsula and Sumatra is increased. This is due to the lack of the counterclockwise turning of the wind over the equatorial region (Figs. 6-11c,d), which results in reduced low-level convergence along the western Borneo coastline. Because there is no vortex to induce the clockwise turning, there is increased interaction between the northeast winds and the terrain of the Malay Peninsula and Sumatra, which contributes to increased low-level convergence and deep convection over those areas.

Another assessment of the interactions between the Borneo vortex and cold surges is to compare the location of the vortex center in the composite constructed from cases that only contain a vortex but no surge (Fig. 6-9f), all vortex cases with and without a surge (Figs. 6-5d), and the composite of vortex and surge cases (Figs. 6-9h). When a vortex is present without a surge, there is a cyclonic turning of the winds over the southern South China with no closed circulation center. In the all-vortex
composite (Fig. 6-5d), a closed cyclonic circulation is centered over the southern South China Sea. In the vortex and surge composite (Fig. 6-9h), the center of the vortex is shifted to be oriented along the western Borneo coastline. Although the presence of the surge acts to increase the strength of the vortex, the surge results in a shift of the vortex center from being located over the southern South China Sea (Fig. 6-9b) to be near the Borneo landmass (Fig. 6-9d).

An extreme case of the interaction between a strong cold surge and a Borneo vortex is the rare formation of Typhoon Vamei near the equator on 26 December 2001 (Chang et al. 2003). In addition to the vanishing Coriolis parameter, an important reason for such a formation to be extremely rare is the typical displacement of the vortex center toward Borneo under strong surge conditions. Since a tropical cyclone needs to develop over ocean, a vortex with much of the cyclonic circulation that lies over land cannot intensify into a tropical cyclone regardless of the strength of the surge. This interesting case of equatorial tropical cyclone formation will be discussed in section 4.

c. The effects of the Madden–Julian oscillation

Chang et al. (2005b) used a singular-value decomposition (SVD) on 30–60-day filtered 850-hPa winds and OLR to identify MJO activity over the Maritime Continent region. In their analysis the periods of MJO presence are separated from those of no-MJO presence. Within a period of MJO presence, four phases—dry, wet, and two transitions—defined by the two leading SVD modes are shown in Fig. 6-12. It is important to note the different meanings in the common usage of the term MJO. In the operational forecast community, the wet phase has often been called an MJO, but here the wet phase is only one of the four phases of an MJO period. Similarly, a dry phase is not the same as a period of no MJO, even though both are characterized by suppressed convection.

The convection pattern of Chang et al.’s MJO phase 1 (dry phase, Fig. 6-12a) resembles that of the Wheeler and Hendon (2004) MJO Real-time Multivariate (RMM) phases 1 and 2, in that the enhanced convection appears over the eastern Indian Ocean and the suppressed convection appears over the Maritime Continent. The convection pattern of Chang et al. MJO phase 3 (wet phase, Fig. 6-12c) resembles that of the RMM phases 5 and 6, both of which have enhanced convection over the Maritime Continent. However, there is an important difference between the two MJO index schemes. The Wheeler and Hendon index is based on the empirical orthogonal functions of the combined fields of 850- and 200-hPa zonal winds and OLR. The index does not contain meridional wind information, which is essential in the structure of synoptic-scale weather
systems. The Chang et al. MJO phases use the 850-hPa total wind and are therefore directly applicable for studying the interactions between MJO and cold surges and Borneo vortex, both of which have strong meridional wind component. It will become clear that as an MJO moves over the Maritime Continent, it is the interaction of its meridional wind components with the synoptic systems that reveals the most discernible effects of the interactions.

During the dry phase (Fig. 6-12a), convection over the Maritime Continent is reduced and equatorial easterly anomalies exist between 80°E and 150°E. During the dry-to-wet phase (Fig. 6-12b), the region is in a transition from the reduced convection regime of the MJO, which has moved eastward, to the approaching active convective regime of the MJO. Increased convection and low-level westerlies exist immediately west and south of the Malay Peninsula, Sumatra, and Java. During the wet phase (Fig. 6-12c), convection is enhanced over the eastern portion of the Maritime Continent and low-level westerly anomalies exist throughout the region. Furthermore, the Australian monsoon trough is very well defined and 850-hPa northeasterly winds increase throughout the South China Sea. During the wet-to-dry transition phase (Fig. 6-12d), the active convection regime has moved to the equatorial western Pacific and the reduced convection regime is approaching the Maritime Continent from the west. Low-level equatorial westerly (easterly) anomalies are found to the east (west) of the Maritime Continent.

In addition to affecting the convection over the Maritime Continent, the circulation of the MJO also interacts with the synoptic-scale disturbances. In particular, Rossby-type responses that are manifest in the subtropical circulations flanking the MJO-induced enhanced or reduced convection are associated with large-scale meridional wind patterns. These meridional winds may act to reinforce or weaken a cold surge event. The net effect of the MJO is to suppress cold surges (Table 6-1). When MJO is present, the chance of a cold surge during the dry and dry-to-wet phases (15%) is almost one-half those during the wet and wet-to-dry phases (28%) and during periods when no MJO is present (29%). This contrast is consistent with the anomalous 30–60-day 850-hPa wind patterns (Fig. 6-12), in that there are anomalous southerly winds during the dry-to-west phases over the South China Sea that apparently inhibit the development of cold
Chang et al. (2005b) further showed that it is mainly the weak surges (surge index between 8 and 10 m s\(^{-1}\)) that are inhibited.

The MJO has an even larger suppression effect on the Borneo vortex, with nearly twice as many vortex cases occur during no-MJO periods than MJO periods (Table 6-2). On the other hand, the distribution of the number of vortex cases varies only modestly among the four phases when MJO is present. During the MJO dry phase, the composite Borneo vortex case (Fig. 6-13a) is similar to the composite of vortex cases with no surge (Figs. 6-9b,f). There is a broad area of counterclockwise turning of the low-level winds over the southern South China Sea with no closed circulation. Convection is increased over the southern South China Sea and reduced over the equatorial Southern Hemisphere, Malay Peninsula, and Sumatra. This similarity indicates that the MJO dry phase may inhibit cold surges so the Borneo vortex most likely occurs without a surge during this phase. This is consistent with the anomalous 30–60-day subtropical ridge over the western North Pacific with southerly anomalies over the South China Sea in Fig. 6-12a. During the dry-to-wet phase, the deep convection over the equatorial South China Sea and the low-level winds associated with the Borneo vortex (Fig. 6-13b) become more organized. Northeast winds appear over the southern South China Sea, but their magnitude is much less than that in the vortex and surge composites (Fig. 6-9d). Overall, the vortex composite patterns associated with both the dry and dry-to-wet phases are not very different from composites of vortex-only and vortex–surge cases, suggesting that the MJO primarily influences the Borneo vortex by inhibiting cold surges and reducing their enhancement effect on the Borneo vortex.

Although the MJO wet phase is associated with a higher cold surge frequency than the dry phase, the cyclonic circulation of the vortex seems to be more linked to cyclonic horizontal shear associated with equatorial westerly winds rather than northeasterly winds that extend through the southern South China Sea. These increased equatorial westerlies are associated with the enhanced MJO-scale convection over the eastern portion of the Maritime Continent (Fig. 6-12c). Consequently, the center of the vortex is located over the southern South China Sea (Fig. 6-13c). During the wet-to-dry phase (Fig. 6-13d), the vortex is very weak with cyclonic shear present only over the extreme western South China Sea and Malay Peninsula. Southwesterly
anomalies exist over the primary region of the southern South China Sea, and reduced deep convection spreads northeastward along the west coast of Borneo. The relationships among the MJO, cold surges, and the Borneo vortex are summarized in Fig. 6-14. It is clear that the presence of the MJO is associated with fewer numbers of vortex cases, and the occurrence of vortex cases during periods of weak surges is most reduced during the MJO. Therefore, while the presence of a surge acts to increase the strength of the Borneo vortex, the frequency of surges is reduced when the MJO is present. Often the MJO-scale circulation pattern directly opposes the cold surge wind pattern. Therefore, weak surges may be more inhibited during periods of strong MJO. Primarily because of the impact of the MJO on cold surge intensity and frequency, 66% of the vortex cases occur during no-MJO periods. The Borneo vortex is least likely to occur when the inactive convective portion of the MJO extends to the Maritime Continent with large-scale low-level diffluence that acts to restrict the impact of cold surges on convection in the southern South China Sea. This complex relationship among MJO, cold surges, and the Borneo vortex and the effects of the topography contribute to the variability in convection patterns over a variety of space and time scales.

4. Equatorial cyclogenesis and the formation of Typhoon Vamei (2001)

The monsoon heavy rainfall events often produce heavy floods and major economic and life losses in the Maritime Continent area, particularly during boreal winter when cold surges and synoptic disturbances including the Borneo vortex are most active (Johnson and Chang 2007). However, tropical cyclogenesis has never been observed over the warm surface of the equatorial South China Sea until the development of Typhoon Vamei in December 2001. The formation of Typhoon Vamei was especially noteworthy because it formed at 1.5°N in the southern tip of the South China Sea at 0000 UTC 27 December 2001, a latitude that most previous literatures and textbooks (e.g., Gray 1968; Anthes

| TABLE 6-1. The number of surge cases with respect to the MJO and the phase of the MJO. |
|----------------------------------------|------------------|----------------|-----------------|-----------------|
| No MJO days                           | MJO days         | MJO phases 1–2 | MJO phases 3–4  |
| No. of no-surge days                  | 845              | 670            | 381             | 289             |
| No. of surge days                     | 242              | 138            | 56              | 82              |
| Percentage of surge days              | 29%              | 21%            | 15%             | 28%             |
1982; McBride 1995) ruled out for development because of the smallness of the Coriolis parameter. Thus an equatorial development needs a basic-flow relative vorticity that reaches certain minimum magnitude. Even though the Borneo vortex is a climatological quasi-stationary feature during the Asian winter monsoon, it is of synoptic spatial and time scales. The only possible mechanism to provide the required large-scale background vorticity will have to come from the sustained cold surges with periods up to a week or more.

The equatorial cyclogenesis of Typhoon Vamei was very different from drifting vortices that have been observed to cross the equator since the beginning of the satellite era (e.g., Chang and Maas 1976) because such vortices were generated away from the equator and their intensity while crossing the equator was usually weak. It was also different from twin tropical cyclones that saddle the equator (e.g., Keen 1982; Lander 1990), where the centers of these tropical cyclones were all sufficiently away from the equator for the Coriolis parameter to play a dominant role. The cyclogenesis induced by cross-equatorial cold surge discussed by Love (1985) occurred at subtropical latitudes (13°–16°).

a. Dynamics of surge-forced equatorial cyclogenesis in the absence of terrain

Two decades before the observation of Vamei, Lim and Chang (1981) proposed an equatorial cyclogenesis process to study the dynamics of a midlatitude cold surge into the tropics. They used a barotropic divergent model and argued that even though the tropical motions responding to surges are likely to be associated with cumulus convection, the shallow-water system may be regarded as a two-level approximation to a baroclinic system because in the free atmosphere the two are equivalent for a given vertical structure (Matsuno 1966). They chose a scale height for their shallow-water model such that the model waves have phase speeds close to those observed. This approximation is justified because Chang (1977) has shown that in a baroclinic model with a balance between diabatic heating and damping, the equatorial wave behavior is altered in such a way that the vertical wavelength of even the higher-frequency waves tends to be comparable to the depth of the troposphere.

Lim and Chang’s (1981) shallow-water model on an equatorial beta plane with no mean flow is

\[
\begin{align*}
\frac{\partial u'}{\partial t} + g \frac{\partial h'}{\partial x} - \beta y' &= 0, \\
\frac{\partial v'}{\partial t} + g \frac{\partial h'}{\partial y} + \beta y' &= 0, \\
\frac{\partial h'}{\partial t} + H \left( \frac{\partial u'}{\partial t} + \frac{\partial v'}{\partial t} \right) &= \Phi'
\end{align*}
\]

Here \( u', v', \) and \( h' \) are the perturbation zonal and horizontal velocity and height, respectively; \( g, \beta, \) and \( H \) are the constant gravitational acceleration, equatorial Coriolis parameter gradient, and mean height, respectively. The variable \( \Phi' \) represents a time-dependent mass forcing that is used to emulate pressure surge from the tropics and is expressed in a Fourier series form:

![Figure 6-13](https://example.com/702687.png)

Fig. 6-13. Composite maps of convective indices and 925-hPa winds (m s\(^{-1}\)) for MJO and vortex cases when the MJO is in (a) dry, (b) dry-to-wet, (c) wet, and (d) wet-to-dry phases (adapted from Chang et al. 2005a.)

---

**Table 6-2. The distribution of vortex cases (days) with respect to MJO periods and phases of the MJO.**

<table>
<thead>
<tr>
<th>MJO phase</th>
<th>No MJO</th>
<th>MJO phase 1</th>
<th>MJO phase 2</th>
<th>MJO phase 3</th>
<th>MJO phase 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>No vortex</td>
<td>675</td>
<td>542</td>
<td>148</td>
<td>163</td>
<td>157</td>
</tr>
<tr>
<td>Vortex</td>
<td>412</td>
<td>216</td>
<td>57</td>
<td>69</td>
<td>50</td>
</tr>
</tbody>
</table>
Here $\Phi_c$ and $\Phi_s$ are the expansion coefficient of the space-dependent part of the forcing function in a Fourier series in $x$ direction, $m$ is the number of waves, $\zeta$ is the nondimensionalized $y$ coordinate, $k$ is the $x$ wavenumber, $\tau$ is a time constant; various parameters in this equation are used to specify the spatial and time scale of the forcing. The time-dependent part is normalized in the sense that total time integration gives a constant value of unity. Lim and Chang (1981) studied various time scales of large-scale forcing cases by specifying the forcing to have a Gaussian spatial distribution in the form of

$$\Phi' = \exp\left\{ -\frac{1}{2} \left[ \left( \frac{\zeta - a}{\sigma} \right)^2 + \left( \frac{\lambda}{\lambda} \right)^2 \right] \right\} \left( \frac{t^2}{2\tau^2} e^{-\tau t} \right).$$

Here $a$ is the nondimensionalized $y$ coordinate; $\lambda$ and $\sigma$ are the nondimensional half-width of the $x$ and $y$ profile of the forcing function, respectively; and the $\zeta$ profile is expanded in terms of the Hermite solutions $D_n(\zeta)$:

$$\exp\left\{ -\frac{1}{2} \left( \frac{\zeta - a}{\sigma} \right)^2 \right\} = \sum_{n=0}^{\infty} p_n D_n(\zeta),$$

with the expansion coefficient $p_n$, governed by the following recurrent relation:

$$p_n = \frac{a}{n(1 + \sigma^2)} p_{n-1} + \frac{\sigma^2 - 1}{2n(1 + \sigma^2)} p_{n-2},$$

with $p_{-1} = 0$ and

$$p_0 = \left( \frac{2}{1 + \sigma^2} \right)^{1/2} \sigma \exp\left\{ -\frac{\sigma^2}{2(1 + \sigma^2)} \right\}.$$

Lim and Chang considered a forcing centered at $30^\circ$N, with $\tau = 1$ day, $\sigma = 1$ (corresponding to $-11^\circ$ latitude), $\lambda = R/5$ where $R$ is the radius of Earth, $a = 2.5$, and $c = 30\text{ m s}^{-1}$, which implies a phase speed of $\sim 10\text{ m s}^{-1}$ for the gravest Rossby mode ($n = 1$). The result computing up to zonal wavenumber 29 and Hermite mode 29, which provide better than 99.9% accuracy, is shown in Fig. 6-15.

The time evolution of velocity and geopotential depicts the geostrophic adjustment process after the forcing is turned on. On day 2, a high pressure center is building up over the area of forcing with a slightly sub-geostrophic anticyclonic outflow. The equatorward wind speed is stronger because of the smaller Coriolis parameter. It becomes noticeably cross isobaric south of $15^\circ$S and flows cross the equator, backing sharply from northeasterly to westerly in the process. The flow strengthens and becomes more geostrophic on day 3. The backing of the cross-equatorial flow becomes more pronounced resulting in a distinct east-northeast–west-southwest equatorial shear line located southeast of the high. On day 4, a vortex (area a in Fig. 6-15) develops on this shear line, and south of it a southwesterly cross-equatorial flow is established. Two days later, the equatorial vortex drifts southwestward and begins to develop around it a small high pressure area just south of the equator. The midlatitude anticyclone now shows a marked northeast–southwest tilt with a ridgeline extending toward the equator. Parallel to this ridgeline a band of low pressure appears to the east. The southwesterly cross-equatorial flow strengthens and makes a distinct wave (area d) as it swings back south to merge with the eastward-propagating equatorial westerlies. A region of counterclockwise rotation is formed at the equator to the west of area d, nearly enclosing the band of low pressure. Thus an equatorial cyclogenesis appears to take place as a result of the forcing at $30^\circ$N.

By day 9, the northeast–southwest tilt of the mid-latitude anticyclone becomes very pronounced. A streak of strong northeasterly winds surges from about $25^\circ$N to near the equator. Between this northeasterly streak and the southwesterly cross-equatorial flow, a very marked shear line is developed and extends from just south of the equator up to about $15^\circ$N (area c). This shear line also roughly divides a general area of high pressure to its west from a band of low pressure to its east, where the counterclockwise rotation becomes more conspicuous with closed circulation centered just north of the equator. This large area of complex motions drifts slowly westward with the midlatitude anticyclone. The equatorial wave (area d) and westerlies (area b) propagate farther to the east and become detached from this area of major response.

The identity of the response can be analyzed by a mode-by-mode computation. Figures 6-16a–f show the velocity field of all wave modes of $n = -1$ and 0 and the Rossby modes of $n = 1, 2, 3, \text{ and } 4$, respectively, while Fig. 6-16g shows the combined velocity field of the
Rossby modes $n = 1$ to 9, all for day 9. Comparing Fig. 6-16 and Fig. 6-15, it is clear that the main response that drifts slowly westward is a congregation of Rossby waves, while the eastward-propagating equatorial wave (Fig. 6-16d) and westerlies (Fig. 6-16b) are manifestations of the mixed Rossby–gravity and Kelvin wave groups, respectively.

The development of the main response may be interpreted in terms of wave-group behavior of the Rossby waves of different Hermite modes. For each Hermite mode ($n$), Rossby waves of different zonal wavenumber combine to form a wave group with a zonal width comparable to the longitudinal extent of the forcing. The wave groups of the lower Hermite modes, $n = 1–3$ have local wavenumber near 3, while those of higher modes, $n = 4–7$, have local wavenumber near 5. All of these wave groups have westward group velocity, but the speed is smaller for the higher modes, as can be expected from the characteristics of the Rossby wave dispersive relationship. The wave group $n = 1$ therefore moves fastest to the west, followed successively by wave groups of increasing $n$. As time goes on, the wave groups of different $n$ drift farther and farther apart. Since equatorial motions are mainly made up of contributions from lower modes while midlatitude motions are mainly made up of contributions from the higher modes, the development of a northeast–southwest tilt in the midlatitude anticyclone and the equatorial disturbances may be seen as a natural consequence of the gradual westward displacement of the lower mode wave groups relative to the higher mode wave groups. The strong northeasterly streak is built up with contributions from many Hermite modes; modes 1 and 2 make up the far southwest end of the streak, while modes 4 and 5 make up the northeast part. The southwesterly cross-equatorial flow is made up mostly of contributions from modes 2–6. As the lower modes move farther and farther westward, their influences on the eastern part of the disturbance diminish and the motions there gradually develop higher mode characteristics.

Many features in Lim and Chang’s (1981) Rossby mode of response are reminiscent of the northeasterly surges during the Asian winter monsoon. Following a pressure surge in the midlatitudes, a belt of strong winds rapidly builds up, sweeping down from about 20°N to the equator and veering gradually from north-northeasterly
to east-northeasterly. This wind belt appears to correspond to the wind surges observed over the South China Sea that often develop within 24 h after a steep pressure rise occurs in southern China. Although the flow pattern is basically a feature of the Rossby-type motion, its development however, is not through the usual mechanism of vorticity advection but through a process of separation of the Rossby mode response from the gravity mode responses. In a way, this may be regard as a geostrophic adjustment process on an equatorial beta plane. The velocity scale therefore is the gravity wave speed and not the advective speed, which is consistent with the rapid propagation of the wind surge over the South China Sea.

The northeasterly wind belt predicted by the theory, however, is too broad and penetrates farther westward than observed. More realistic predictions could possibly be made had the effects of terrain been considered. Since wind surges over the South China Sea occur mainly in the lowest 1 km of the atmosphere, frictional dissipation could significantly reduce the wind speed over land and thereby leave a narrower wind belt over the ocean. The westward extent of the surge also would be curtailed by the mountain ranges of Indochina and the Malay Peninsula as the barrier effect would be significant for the shallow layer of strong winds.

An important implication of Lim and Chang’s (1981) theory is that the counterclockwise rotation developed at the equator southeast of the main anticyclone is an inherent property of the Rossby wave group response to the cold surge even in the absence of terrain. It is consistent with the mean winter condition in the general vicinity of Borneo and southern South China Sea, and the enhancement of the Borneo vortex following surges discussed in section 3. The terrain effect of the Borneo Island is no doubt a fundamental reason for the tendency of development of low pressure and deep cumulus convection in this region, but the geostrophic adjustment and potential vorticity conservation following a cross-equatorial surge from the north can independently cause an equatorial cyclogenesis.

b. The development of equatorial Typhoon Vamei

The life cycle of the cyclone is shown in Fig. 6-17, which is based on the best track and intensity published by JTWC. The storm made landfall over southeast Johor on the southern tip of peninsular Malaysia, about 50 km northeast of Singapore, at 0830 UTC 27 December 2001, 12 h after the formation. It then weakened rapidly to a tropical depression. It continued in its west-northwest track across southern Johor and the Malacca Straits, and made landfall again in Sumatra. Upon entering the Bay of Bengal, the storm regenerated and continued its northwest track before dissipating in the central Bay of
Bengal on 31 December 2001. During the short period of 12 h as a typhoon and another 12 h as a tropical storm, Vamei caused damage to two U.S. Navy ships, including a carrier, and flooding and mudslides in southern peninsular Malaysia's Johor and Pahang states. More than 17,000 people were evacuated and five lives were lost.

The detection of Vamei as a typhoon-strength tropical cyclone was somewhat accidental (Chang and Wong 2008). Vamei was initially classified as a tropical storm by the Japan Meteorological Agency with estimated winds of 21 m s\(^{-1}\). It was later upgraded it to the typhoon category by the JTWC in Hawaii when shipboard observations of sustained winds of 39 m s\(^{-1}\) and gusts up to 54 m s\(^{-1}\) from several U.S. Navy ships within the small eyewall were reported.

Figure 6-18 shows the MODIS satellite image on 27 December 2001. Vamei’s circulation center can be estimated to be just north of 1°N, but an eye is not observable under the clouds. Even though the size of the typhoon is quite small, the spiral cloud bands emanating out from around the center clearly indicate that the storm circulation was on both sides of the equator. Figure 6-19 shows the Japanese GMS image at 0232 UTC of the same day. Feeder bands from both sides of the equator spiral into the center of Vamei, where a small eye is visible. An eye was also observed in TRMM and SSM/I images within the preceding 2 h. The size of the eye estimated from different sensors ranges from 28 to 50 km in diameter. Vamei’s small size as it formed in the southern end of the South China Sea and the short duration before its initial landfall make it difficult to observe its highest wind speed from ground-based observations or to estimate its intensity from satellite images. Without the chance passage by the U.S. ship (USS) Carl Vinson carrier group through its eyewall, JTWC may not be able to operationally upgrade the intensity of the storm to a typhoon.

The synoptic development preceding the formation is shown in Fig. 6-20, which is a time sequence of the 1° × 1° Navy Operational Global Atmospheric Prediction System (NOGAPS) 850-hPa wind and vorticity (Chang et al. 2003). Starting from 19 December 2001, a cold surge developed rapidly over the South China Sea, while the center of the Borneo vortex was located near 3°N on the northwest coast (not shown). The strength and persistence of this surge was helped by the strong meridional gradient of sea level pressure in the equatorial South China Sea during December 2001, which was above normal according to the report by the Bureau of Meteorology, Northern Territory region, published in 2002. Figure 6-20 depicts the southwestward movement of the vortex from along the Borneo coast toward the equator. By 21 December, the center of the vortex had moved off the coast over water, where the open sea region in the southern end of the South China Sea narrows to about 500 km, with Borneo to the east and the Malay Peninsula and Sumatra to the west. This overwater location continued for several days. While the vortex center remained in the narrow equatorial sea region, the strong northeasterly surge persisted, and was slightly deflected to the northwest of the vortex.

As discussed in section 3, this near “trapping” of the Borneo vortex by a sustained surge is unusual because normally the vortex center would be pushed eastward by the strengthening surge that streaks southwestward in
the middle of the South China Sea. Figure 6-21 shows the location of counterclockwise circulation centers during the 51 boreal winters of 1951/52–2001/02 based on a streamline analysis of the NCEP–NCAR reanalysis 925-hPa wind. The distribution of the frequencies of vortex centers lasting for 96 h or more in four subregions near the northwestern coast of Borneo is superimposed on the diagram. There were only six occurrences of a Borneo vortex center that stayed continuously for 96 h in the southwestern box that is in the narrow equatorial region. This unusual trapping caused the cross-equatorial flow wrapped around the vortex and provided a background area of cyclonic relative vorticity with a magnitude of $1 \times 10^{-5} \text{s}^{-1}$, which is comparable to that of the Coriolis parameter $5^\circ$ away from the equator, a latitude where formation of tropical cyclones is not considered unusual.

Cold surges are frequent events in the South China Sea during the Asian winter monsoon, but the surge preceding the development of Typhoon Vamei was especially intense and long lasting. Chang et al. (2003) reported that the period of 0000 UTC 19–0000 UTC 25 December 2001 was the most sustained and intense surge period among the approximately 15 surge episodes in the three winters that QuickSCAT wind data became available. Furthermore, none of the other 14 surge periods coincided with the occurrence of a Borneo vortex that migrated into, and stayed in, the narrow open sea region (the southwest box in Fig. 6-21).

Evidence of the strength of Vamei and its relationship with the cold surge can also be confirmed from

---

**FIG. 6-18.** MODIS satellite image of 27 Dec 2001 showing Typhoon Vamei near Singapore (from Chang et al. 2003).

**FIG. 6-19.** Japanese GMS image at 0232 UTC 27 Dec 2001 (from Chang and Wong 2008).
QuickSCAT satellite scatterometer wind data. Figure 6-22 shows that the QuickSCAT wind direction and speed at 2232 UTC 26 December 2001 captured both the signal of Vamei as it develops to typhoon strength and the remnant of the continuing surge wind upstream in the northern South China Sea. In the southern perimeter, the wind speed at 10-m height has already reached above 27 m s\(^{-1}\) over an area of about 1° latitude × 1° longitude. The northern spiral band extends to about 6°N and is detached from the cold surge wind belt farther north.

Chang and Wong (2008) examined the results of Lim and Chang’s cold surge theory and found a number of interesting features that resemble the characteristics of the Vamei development process. Figures 6-23a and 6-23b show the theoretical solutions three days apart (day 3 and day 6 in Fig. 6-15), with the location of the high center on day 6 (Fig. 6-23b) shifted eastward by 0.4L, where L is the approximate radius of the high pressure cell, to align the high centers that otherwise will show a westward propagation. Figures 6-23c and 6-23d are the NOGAPS 850-hPa wind analysis for 19 December and 22 December 2007, respectively.

As discussed earlier, Fig. 6-23a resembles a typical cold surge event that follows the southeastward movement of an East Asian surface high center with the development of a northeast–southwest tilt. This tilt is due to the dispersion of equatorial beta-plane Rossby waves in which the lower meridional modes have larger amplitudes closer to the equator and therefore propagate westward more quickly. As the northeasterly wind strengthens south of the high center, it streams southward, and after crossing the equator, it turns eastward between the equator and 15°S leading to a counterclockwise curvature flow at the equator. Figure 6-23b shows the solution three days later, in which the northeast–southwest tilt becomes even more pronounced. To the southeast of the northeasterly surge streak, southwesterly cross-equatorial winds enhance the counterclockwise circulation belt over the equator, which is now northeast–southwest oriented. The counterclockwise flow in both Figs. 6-23a and 6-23b are highlighted by color streamlines. Figures 6-23c and 6-23d show the NOGAPS 850-hPa wind analysis at the beginning of the actual cold surge (Fig. 6-23c, 0000 UTC 19 December), and 3 days later (Fig. 6-23d, 0000 UTC 22 December). Color streamlines are also sketched on Figs. 6-23c and 6-23d to highlight the counterclockwise rotation across the equator.

FIG. 6-20. NOGAPS 1° × 1°850-hPa wind and vorticity (red positive, green negative) at 0000 UTC 20–26 Dec 2001 (from Chang et al. 2003).

FIG. 6-21. Location of NCEP–NCAR 925-hPa counterclockwise circulation centers during 51 boreal winters. The four subregional boxes are enclosed by equator–7°N, and 103°–117°E. The internal partitions are 3°N and 110°E. The first number in each box indicates the frequency of persistent 925-hPa cyclonic circulation center that lasted for 96 h or more, based on the 1951/52–2001/02 DJF NCEP–NCAR 2.5° × 2.5° reanalysis. The second number is the total number of days a center is identified (from Chang et al. 2003).
that is produced by the cold surge flow. Comparing Figs. 6-23a, b and 6-23c, d, there is a strong resemblance in the overall geometry of the development of the cross-equatorial counterclockwise circulation over the 3-day span.

As pointed out in section 4b, the surge wind belt produced by the theory is too broad and penetrates too far westward. The observed intense surge wind belt is confined within the width of the South China Sea of about 750 km, which is approximately one-half of the width in the equatorial beta-plane solution in Fig. 6-23b. If the effects of topography are included, the strongest surge winds in the theory would be confined over the South China Sea, and the westward extent of the surge would be constrained by the barrier effect of the Indo-China and the Malay Peninsula mountain ranges. These effects will also add a forcing to induce the flow to turn equatorward in addition to that required by the conservation of potential vorticity.

The theory and the actual development may be made to resemble each other by scaling the east-west dimension of Figs. 6-23a and 23b to one-half of the original size ($L = 15^\circ$ longitude instead of $30^\circ$ in Fig. 6-23), or treating the highlighted rectangular area in Fig. 6-23a as being comparable to the domain of the NOGAPS plots in Figs. 6-23c and 6-23d. In addition, the shifting of the high center in Fig. 6-23b eastward by $0.4L$ in Fig. 6-23a may be accounted for by the slower propagation due to reduced zonal scale and two factors in real-world cold surge events: The eastward movement of the East Asian surface high center due to the midlatitude westerly mean flow, and the fixed location of the surge belt that is restricted geographically by the South China Sea.

c. Why is tropical cyclone development in the equatorial South China Sea so rare?

The interaction between the winter monsoon circulation and the complex terrain and the moisture from the warm ocean surface provided the latent heat and favorable vorticity conditions for development of Typhoon Vamei. However, the strong cold surge and Borneo vortex that led to the development of Vamei are commonly observed major systems of the Asian winter monsoon, and abundant low-level warm and moist air is constantly present over the South China Sea during every season. The most interesting question is not necessarily just how or why Typhoon Vamei could form so close to the equator. Rather, the question is why more typhoon formations have not been observed in the equatorial South China Sea (Chang and Wong 2008). Indeed, the many numerical simulations of Typhoon Vamei (e.g., Koh 2006; Chambers and Li 2007; Juneng et al. 2007; Tangang et al. 2007; Loh et al. 2011) have not addressed this key question. Numerical model simulations should not be considered completely successful only by simulation of the development of Vamei, but also by how successfully the model simulates the non-development of the numerous other cases of cold surge and Borneo vortex that occur in every boreal winter.

An important reason tropical cyclones did not develop in these other cases is due to the terrain factor—the narrowness of the equatorial South China Sea. This factor plays two counteracting roles that combine to make the occurrence of the typhoon formation possible but rare. One role is the channeling and strengthening of the cross-equatorial surge winds that help produce the background cyclonic vorticity at the equator. On the other hand, the open water region of approximately $5^\circ$ longitude is only sufficient to accommodate the diameter of a small tropical cyclone. It is too small for most synoptic disturbances to remain over the water for more than a day or so, particularly during strong surge events when a Borneo vortex is likely displaced to the east. In the unusual case of Typhoon Vamei, the durations of the intense cold surge and the Borneo vortex remained over water significantly longer than normal, which allowed the interaction to continue for nearly a
week until the storm was formed. Based on an analysis of the NCEP–NCAR reanalyses during the boreal winters of 1951/52–2001/02, Chang et al. (2003) estimated that the probability of an equatorial development from similar conditions to be about once in a century or longer. Although this estimate appears consistent with historical records, it is not known whether there might be other near-equatorial developments not observed during the presatellite era.

5. Summary

The Asian monsoon system, which is defined by the seasonal reversal of winds and the associated maximum rainfall seasons, is driven by the continent–ocean thermal contrast. It derives its energy mainly from the latent heat release in deep convection that produces the heavy rainfall. Even though the monsoon circulation has a very large spatial scale, the convection and rainfall patterns are heavily influenced by the complex terrain on various scales. Indeed, mesoscale and local terrain features can strongly distort the monsoon rainfall distribution, which cannot be considered from just a large-scale perspective. This terrain effect is especially prominent during boreal winter, when the center of deep convection is located in the deep tropics around the Maritime Continent region, where islands of different sizes are interspersed within the surrounding oceans.

Over this region, the equator serves as a general demarcation between the Asian summer (to its north) and...
winter (to its south) monsoon rainfall regimes. However, the seasonal convection is locally dominated by interactions between the complex terrain and the annual reversal of the large-scale surface winds. These interactions cause the summer and winter monsoon rainfall regimes to intertwine across the equator. In particular, the boreal winter regime extends far northward along the eastern flanks of the major island groups and landmasses. However, there is no complementary extension of the boreal summer regime into southern latitudes.

The seasonal march is also asymmetric during the transitional seasons, with the maximum convection following a gradual southeastward progression from the Asian summer monsoon to the Asian winter monsoon, but a sudden transition in the reverse. The mass redistribution between land and ocean areas during the transition season contributes to this asymmetric march, which can be traced to the orientations of the Asia and Australia landmasses. These orientations and various local terrain features produce sea level pressure patterns that lead to asymmetric wind–terrain interactions throughout the region, and a low-level divergence asymmetry that promotes the southward march of convection during boreal fall but opposes the northward march during boreal spring.

The deep convection and heavy rainfall in the Maritime Continent region during the Asian winter monsoon are among the most energetic tropical convection systems. In the western part of this region and over the equatorial South China Sea, the convection is strongly affected by disturbances ranging from synoptic to intraseasonal time scales. The strongest variability of convection and heavy rainfall are due to the effects of the northeasterly cold surges and the Borneo vortex and their interactions with the terrain. MJO plays a secondary role in lessening these effects when the synoptic-scale systems are of modest intensity.

The effects of a northeasterly surge occur mainly through the interaction with local topography and the dynamic response to the change in latitude. These effects contribute to the turning of the winds and localized patterns of deep convection. In days without a Borneo vortex, deep convection tends to be suppressed over the South China Sea and Borneo and enhanced downstream over the landmasses on the western and southern peripheries of the equatorial South China Sea. The pattern is reversed in days with a vortex. The simultaneous presence of a cold surge enhances this contrast. The surge also interacts with the Borneo vortex, in that the vortex is strengthened and its center shifts from over the South China Sea to over the western coast of Borneo. The frequency of cold surges and vortex days is reduced during periods when the MJO is present. Often the MJO-related circulation patterns oppose the synoptic-scale cold surge and vortex circulations. Thus, a primary impact of the MJO is to inhibit weak cold surge events, which then produces a secondary impact on the Borneo vortex via interactions between the cold surge winds and the vortex.

The strongest motion system that developed as a result of the interaction of a cold surge with a Borneo vortex is Typhoon Vamei off Singapore on 26 December 2001. With its center at 1.5°N and its circulation encompassing both sides of the equator, Typhoon Vamei was the first observed equatorial tropical cyclogenesis. The development depended on a basic flow relative vorticity that is apparently provided by an abnormally strong and sustained cold surge prior to the formation.

The mechanism for a surge-induced equatorial cyclogenesis may be explained via the linear equatorial wave theory. In the absence of terrain, a mass source at 30°N that simulates the effect of the sea level pressure rise due to the migration of the Siberian high can generate the cold-surge-like northeasterly winds over the South China Sea. After an initial period of gravity wave–type motions with strong northerly winds, the main tropical response takes the form of a Rossby wave group. A pronounced northeast–southwest tilt in this Rossby wave group develops because of the faster westward group velocity of the lower meridional modes relative to the higher meridional modes. The dispersion of the Rossby waves gives rise to a cyclonic shear zone near the equator where the Borneo vortex is often observed. These responses suggest that the equatorward strengthening of northeasterly surge winds and the cyclonic circulation can develop without the effect of terrain and latent heating, and thus may be considered the mechanism to generate and maintain the basic-state relative vorticity in the region where Typhoon Vamei formed.

Whereas cold surges and Borneo vortices are common occurrences in the South China Sea during boreal winter, equatorial formation of tropical cyclones are extremely rare because of the terrain effects that are both necessary for and hinder development. The narrowness of the South China Sea helps to channel and strengthen the cross-equatorial surge winds to produce the background cyclonic vorticity at the equator. However, it is difficult for a preexisting Borneo vortex to remain over the water long enough to develop into a tropical cyclone. The formation of Typhoon Vamei was due to unusually persistent durations of an intense cold surge and a Borneo vortex circulation that remained over open water. The probability of a recurrence of these conditions based on the climatology of cold surge and Borneo vortex behavior has been estimated to be about once in a century.
The Maritime Continent is a region that is strongly influenced by ENSO. Even though the greatest influence occurs in the dry boreal summer and boreal fall transitional season (Haylock and McBride 2001; Hendon 2003; Chang et al. 2004b), various local impacts from ENSO are also found during boreal winter (Chang et al. 2004b; Robertson et al. 2011). How the interactions of the transient motion systems are affected by ENSO, such as whether some effects may be through diurnal cycle (Qian et al. 2010, 2013), are unanswered questions. Effects of longer-term variations, from decadal to climate change scales, on these interactions are also mostly unknown. The skills of current climate models are typically low for the Maritime Continent area, with the simulated monsoon–ENSO connection generally weaker than observed, and no clear model consensus on monsoon rainfall over the second half of the twentieth century (Jourdain et al. 2013).

Another important area for future research concerns the role of the MJO. The interaction between the MJO and cold surges and the Borneo vortex discussed in this chapter relate only to the dynamic structure of the motion systems. Not much is known at this time on the thermodynamic interaction between MJO and these synoptic-scale systems under strong terrain effects over a broad area, and on how the surges and Borneo vortex may affect the MJO. The recent field program of the Cooperative Indian Ocean Experiment on Intraseasonal Variability (CINDY)/DYNAMO (Yoneyama et al. 2013; Zhang 2013) was held in the Indian Ocean to study the initiation of MJO, where the open ocean and small island topography are very different from the complex and steep terrain over the Maritime Continent. Field observation experiments, such as the planned Years of Maritime Continent (YMC), and high-resolution modeling studies will help the understanding of the interaction between MJO and the cold surges and Borneo vortex.

Regarding the rare development of Typhoon Vamei (2001), many mesoscale numerical weather prediction models were able to simulate the equatorial cyclogenesis based on initial conditions that are not obviously dissimilar from those observed on many other December or January days. Numerical experiments to simulate both Typhoon Vamei and the nondevelopment of these other days are needed to elucidate the reasons that Typhoon Vamei was the exception. Longer-term climate model integrations may give an indication as to whether there may be a shortening of the century-scale return period of the equatorial tropical cyclogenesis under climate change scenarios.

Acknowledgments. This review was based on research in the past decade supported by the National Science Foundation, the Office of Naval Research, and the National Science Council of Taiwan, R.O.C. We wish to thank Professors Russ Elsberry and Bob Haney and three anonymous reviewers for their helpful comments.

REFERENCES


