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Interactions between the Asian monsoon and the El Niño/Southern Oscillation

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The other chapters in this book have devoted considerable attention to the significant role of the sea surface temperature (SST) conditions over the Pacific and Indian Oceans in climatological aspects as well as variability of the Asian monsoon. The El Niño/Southern Oscillation (ENSO) phenomenon is known to exert a strong influence on the SST patterns throughout the World's oceans. The nature of the interactions between ENSO and the atmospheric circulation over different parts of the Asian–Australian region is obviously of primary importance in understanding many facets of the monsoon system. Of particular interest is the response of the monsoon flows to local and remote SST changes that emerge during ENSO, as well as the impact of monsoon fluctuations on the evolution of ENSO episodes.

The principal goal of the present chapter is to offer a synopsis of the basic characteristics of the ENSO phenomenon, its relationships with monsoonal features in the south Asian and east Asian–western Pacific sectors in different stages of the ENSO cycle, physical mechanisms that contribute to covariability between ENSO and the monsoon system, the effects of monsoon anomalies on the oceanic temperature and circulation fields, and the implications of such feedbacks on the subsequent development of ENSO. This review of the linkages between ENSO and the Asian monsoon is based on the available observational records of the last half-century, as well as output from general circulation model (GCM) experiments that are designed to delineate pertinent aspects of atmosphere–ocean coupling.

12.1 AIR–SEA INTERACTIONS RELATED TO ENSO

Detailed historical accounts of the discovery of a myriad of oceanic and atmospheric features associated with ENSO, as well as the mechanisms contributing to ENSO variability, have been given by Rasmusson (1985), Enfield (1989), Philander (1990), Glantz *et al.* (1991), Cane (1992), Wallace *et al.* (1998), Neelin *et al.* (1998), Jin (2004)

B. Wang (ed.), *The Asian Monsoon*.

and Wang and Picaut (2004), among others. The term 'El Niño' (the child) was originally used by natives of the Peru–Ecuador coasts to refer to a warm ocean current that appears in that region during the Christmas season. It has been known for centuries that the intensity of this annual coastal warming fluctuates from year to year (Quinn and Neal, 1978). The data from an expanded monitoring network established during the International Geophysical Year (1957–1958), which coincided with an unusually strong El Niño, indicate that the warm SST anomaly associated with this episode extended westward to as far as the Date Line. Composite analyses of recent outstanding El Niño events (e.g., Rasmusson and Carpenter, 1981) confirm the large spatial extent of the SST signal associated with El Niño. These empirical studies also reveal that the anomalous SST episodes in the equatorial Pacific often exhibit cyclical characteristics, with warm events being followed by cold ('La Niña') events, and vice versa. The SST anomaly typically emerge in the eastern equatorial Pacific during the boreal spring, attaining maximum strength in the central Pacific 6–9 months later, and attenuating in the northern spring or summer of the following year.

The term 'Southern Oscillation' was coined by Walker and Bliss (1932) to refer to a global-scale east–west seesaw pattern in the sea level pressure (SLP) field. This pattern primarily depicts the out-of-phase relationship between interannual SLP variations over the western Pacific/Indian Oceans, and those over the eastern Pacific (e.g., Troup, 1965; Trenberth and Shea, 1987). This phenomenon was documented by Sir Gilbert Walker in his pursuit to improve monsoon prediction in the Indian region.

A significant milestone in our understanding of the El Niño and Southern Oscillation phenomena was reached when Bjerknes (1966, 1969) pointed out that they are strongly coupled to each other. The basic processes considered in Bjerknes' hypothesis may be summarized as follows. The marked climatological thermal contrast between the 'warm pool' in the western Pacific and cold temperatures in the equatorial eastern Pacific is accompanied by relatively low (high) SLP west (east) of the Date Line. This thermal contrast drives a thermally direct atmospheric circulation loop (the 'Walker Cell') along the equatorial zonal plane, with ascending (descending) motion over the western (eastern) Pacific, a westward air current near the surface, and eastward return flow at upper levels. An anomalous cold event in the eastern tropical Pacific (i.e., La Niña) would enhance the zonal SST and SLP gradients across the Pacific basin, which would correspond to a positive swing of the Southern Oscillation (i.e., higher SLP to the east, lower to the west). The stronger Walker Cell would be accompanied by above-normal easterly winds at the surface, which further amplify the east–west SST gradient through the effects of oceanic advection, upwelling and thermocline displacement (see description of the latter oceanic processes by Cane (1992)). Conversely, the slackened thermal and pressure contrasts across the Pacific basin during an El Niño event would be associated with a negative phase of the Southern Oscillation and weakened easterly trades at the equator, which is conducive to still further oceanic warming. The anomalous episodes of either polarity are hence sustained by the cooperative feedbacks between the atmospheric and oceanic components of the coupled system (see schematic diagrams in McPhaden *et al.* (1998; figure 1)). In recognition of the intimate relationship between El Niño and the Southern Oscillation, the term 'ENSO' has been widely adopted since the 1980s to refer to the myriad of phenomena mentioned above.

The characteristic spatial patterns of atmospheric and oceanic anomalies during the peak stage of ENSO are portrayed in Figure 12.1(a) (color section). This panel has been constructed by subtracting the composite over ten La Niña events (1950, 1954, 1955, 1964, 1970, 1973, 1975, 1988, 1998, 1999; see blue markers along the time axis in Figure 12.1(b)) from that over ten El Niño events (1957, 1965, 1969, 1972, 1976, 1982, 1987, 1991, 1997, 2002; see red markers in Figure 12.1(b)). This composite procedure has been performed for the three-month period from December of the year when the events initiated ('year 0') to February of the following year ('year 1'). We shall henceforth refer to a specific time period within the ENSO time frame by grouping the first letter of the months in that period, followed by the year(s) in parentheses. For instance, the northern winter period considered in Figure 12.1(a) is abbreviated as DJF(0/1). The patterns generated by the above procedure will be referred to as the 'warm-minus-cold composites'. The distributions of SLP (contours) and surface vector wind (arrows) in Figure 12.1(a) have been obtained from the National Centers for Environmental Prediction (NCEP) reanalysis data set (Kalnay *et al.*, 1996). The SST pattern (shading) in this figure has been computed using the data set compiled by Hurrell *et al.* (2005). These charts indicate that, when the ENSO events are fully developed, the zonal extent of the near-equatorial SST anomaly is more than 90° of longitude, and its peak-to-peak amplitude exceeds 3°C. Concurrent with the warm SST anomaly are negative (positive) SLP changes east (west) of the Date Line, with peak-to-peak amplitudes approaching 2 hPa near the two poles of the Southern Oscillation (i.e., subtropical central south Pacific and northern Australia). Anomalous westerly surface flow prevails near the equator during warm events, thus indicating a weakened Walker circulation. The composite patterns in Figure 12.1(a) also indicate considerable convergence of the surface wind to the warm SST anomaly.

The variations of the key atmospheric and oceanic features in Figure 12.1(a) (color section) are depicted in the panels below that figure, for anomalies of (b) SST in the eastern/central equatorial Pacific, (c) SLP over the central subtropical south Pacific and over northern Australia, and (d) surface zonal wind over the central/western equatorial Pacific. These indicators are obtained by averaging the monthly data over the rectangular regions shown in Figure 12.1(a). The temporal fluctuations displayed in Figures 12.1(b–d) illustrate the relationships between the SST and atmospheric changes as noted by Bjerknes.

Bjerknes' hypothesis has offered a good explanation for the amplification of ENSO-related atmospheric and oceanic anomalies through mutual reinforcement. However, further research is needed to elucidate the mechanisms responsible for the cyclical nature of ENSO events. One of the most influential paradigms that addresses this important aspect of ENSO was proposed by Schopf and Suarez (1988), Suarez and Schopf (1988), and Battisti and Hirst (1989). The essential processes considered by these investigators are shown in Figure 12.2. During an

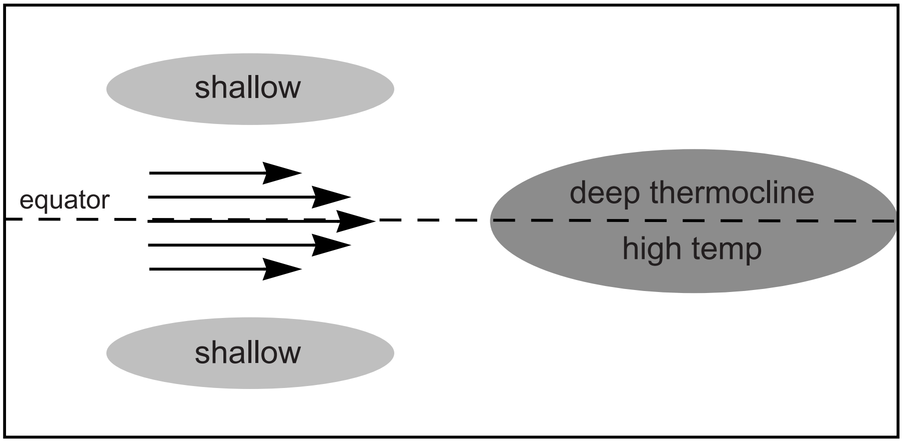

Figure 12.2. Schematic representing the spatial structures of the delayed oscillator mode during El Niño. Anomalies in surface wind and thermocline depth are indicated by arrows and shading, respectively.
From Philander and Fedorov (2003).

El Niño event, the above-normal SST and deepened thermocline in the central and eastern equatorial Pacific is accompanied by westerly surface wind anomalies in the western part of the basin (see Figure 12.1(a), color section). The curl of the wind stress associated with the latter feature leads to shoaling of the thermocline on both sides of the equator. The spatial structure of this pair of off-equatorial oceanic signals resembles that of a Rossby wave packet, which tends to carry the thermocline anomalies westward. When these wave motions encounter the western edge of the basin, a portion of their energy is converted to that associated with Kelvin waves, which first propagate equatorward near the western coast and then eastward along the equatorial Pacific. The shoaling thermocline tendency is preserved in this process at the western boundary. The oceanic signals returned by the Kelvin waves to the central and eastern Pacific are hence opposite in sign to the deep thermocline anomaly that originally resides in that region, and thereby facilitates the transition in the ENSO cycle from a warm phase to a cold phase. In this context, the behavior of the tropical atmosphere–ocean system is analogous to that of a delayed oscillator, with the SST anomaly of a given polarity 'sowing the seed' for its own demise some time later by instigating a host of atmospheric and oceanic changes that eventually generate a new anomaly of the opposite polarity. Observational evidence (e.g., Wyrtki, 1975, 1985) and diagnosis of coupled models (e.g., Zebiak and Cane, 1987) are in support of the applicability of this general paradigm for understanding certain facets of ENSO dynamics. Other oscillator modes highlighting the roles of a charging/recharging of the upper ocean heat content (Jin, 1997), zonal temperature advection in the central Pacific (Picaut *et al.*, 1996), and various aspects of air–sea interaction over the western Pacific (Weisberg and Wang, 1997a; Wang *et al.*, 1999) have also been proposed.

Philander and Federov (2003) proposed that the ENSO events observed in nature may be viewed as a hybrid of the delayed oscillator mode illustrated in Figure 12.2, and other modes arising from local air–sea interactions (e.g., the 'SST mode' examined by Neelin (1991)). Fedorov *et al.* (2003) further noted that random atmospheric disturbances, such as westerly wind bursts over the western tropical Pacific, could play a crucial role in determining the unique characteristics of individual events. The predictability of ENSO is hence limited by the amount of atmospheric 'noise' that is ever present in the climate system.

12.2 PRECIPITATION ANOMALIES IN THE ASIAN–AUSTRALIAN MONSOON REGION DURING ENSO EVENTS

The typical evolution of precipitation anomalies in the Asian–Australian monsoon region in various phases of the ENSO cycle has been documented by Ropelewski and Halpert (1987) using station records. Various other empirical studies on the impacts of ENSO on the monsoon rainfall intensity over the Indian subcontinent, east Asia, and Australia have also been reviewed recently by Webster *et al.* (1998) and Wang *et al.* (2003).

The development of the anomalous rainfall pattern through different stages of ENSO is illustrated using the warm-minus-cold composites from JJA(0) to JJA(1) in Figure 12.3. These charts have been constructed using the data set produced by the Global Precipitation Climatology Project (GPCP; see Huffman *et al.*, 1997), which incorporates measurements of both rain gauges and satellites. Due to the somewhat limited duration of this data set, only three El Niño events (1982, 1991, 1997) and two La Niña events (1988, 1998) have been included in the present composite procedure. The composite charts in Figure 12.3 bear some correspondence to the correlation maps presented by Navarra *et al.* (1999) and Miyakoda *et al.* (1999).

Through much of the JJA(0)–DJF(0/1) period (Figure 12.3(a–c)), the precipitation patterns in the equatorial zone are dominated by negative anomalies over the Indonesian archipelago and eastern Indian Ocean, and by positive anomalies from $\sim 150^{\circ}$E to the Date Line. These features are indicative of the eastward displacement of the Walker circulation during warm ENSO events (Section 12.1). Over the Arabian Sea/India/southern Bay of Bengal region, below-normal rainfall prevails during the summer and autumn of 'year 0' (Figure 12.3(a,b)). This deficiency of Indian monsoon rainfall during the summer of warm events is a well-known phenomenon (e.g., Rasmusson and Carpenter, 1983; Shukla and Paolino, 1983). Comparison between Figure 12.3(a) and (e) suggests that the summertime precipitation anomalies over the Arabian Sea and Bay of Bengal tend to change sign from year(0) to year(1). During SON(0) and DJF(0/1) (Figure 12.3(b,c)), the precipitation changes over the equatorial and southern Indian Ocean are characterized by dry conditions in the east, and wetness in the west. This rainfall pattern is evidently related to a recurrent mode of SST variability in the Indian Ocean basin, which also exhibits distinct east–west contrasts in northern autumn (Section 12.3.1 and Figure 12.5, color section).

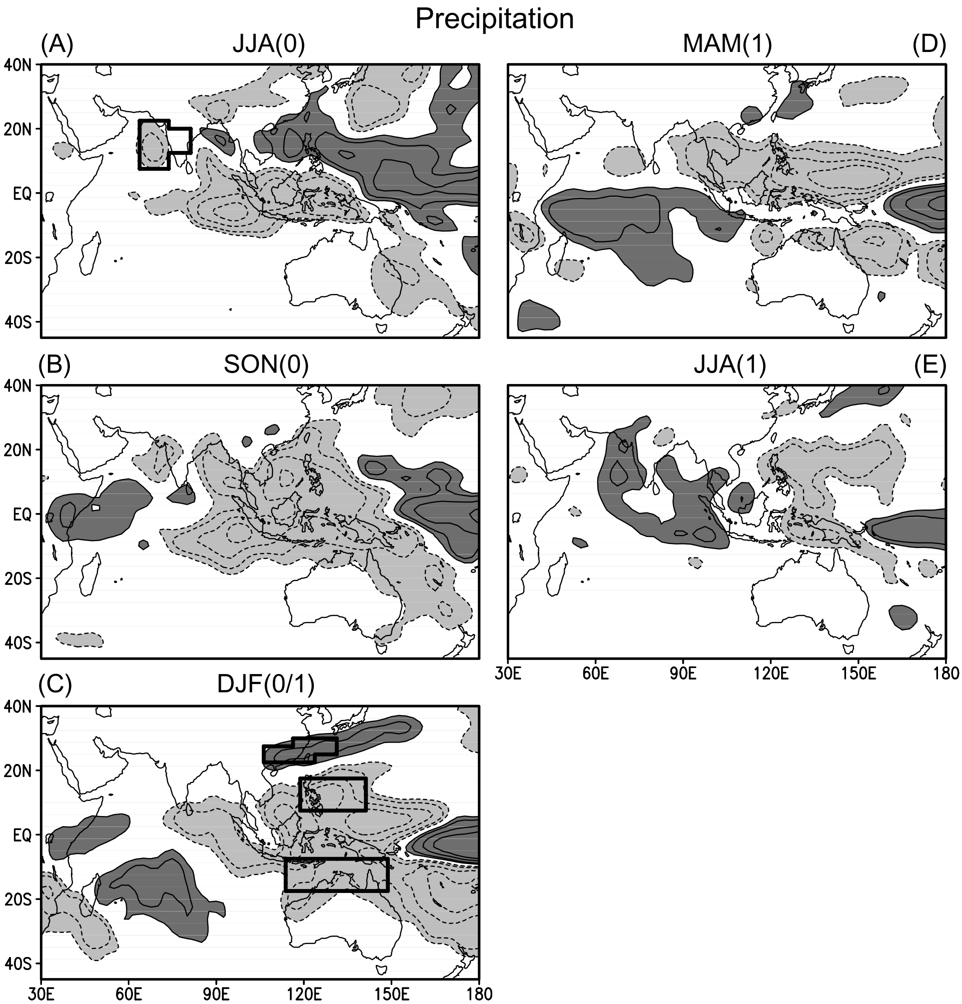

Figure 12.3. Distributions of the warm-minus-cold composites of precipitation during (a) JJA(0), (b) SON(0), (c) DJF(0/1), (d) MAM(1) and (e) JJA(1), as computed using GPCP data for the warm ENSO events of 1982, 1991 and 1997 and the cold events of 1988 and 1998. Contours are shown for ± 1, ± 2, ± 3, ± 4 and $\pm 6\,\mathrm{mm\,d^{-1}}$, respectively. The zero contours are not shown. The sites used for computing the areal averages displayed in Figure 12.4 are indicated in (a) and (c).

The most prominent precipitation anomalies in the east Asian and Australian monsoon regions appear in DJF(0/1) (Figure 12.3(c)). Below-normal rainfall is observed over the Philippines and the nearby oceans, as well as northern Australia. An elongated wet zone is also seen to extend north-eastward from southern China to the waters south of Japan. The dry anomaly in the vicinity of the Philippines is first established in SON(0) over the South China Sea

(Figure 12.3(b)). This feature migrates eastward with time, with its main center being located over the tropical western Pacific in MAM(1) (Figure 12.3(d)). Remnants of the wet anomaly over southern China and Japan are still discernible in MAM(1).

The robustness of the composite precipitation signals identified in Figure 12.3 may be assessed by inspecting the anomalies in various key regions for a larger number of individual warm and cold ENSO events, and comparing the rainfall estimates based on different data sources. In Figure 12.4 are plotted the precipitation anomalies occurring in each of six selected warm events (left panels) and six cold events (right panels), as obtained using the data sets provided by GPCP, Climate Prediction Center Merged Analysis of Precipitation (CMAP; see Xie and Arkin, 1997) and Dai *et al.* (1997). The CMAP fields are based on a combination of rain gauge observations, satellite measurements and numerical model outputs. Both the GPCP and CMAP data have global coverage, and are available from 1979 onwards. Dai's product consists of gridded analyses of rain gauge records exclusively, and covers land points over the period of 1950–1995. The data values shown in Figure 12.4 are areal averages taken at four individual monsoon regions: central India–Arabian Sea (IND), southern China and the East China Sea (CHI), the Philippine Sea (PHI), and northern Australia (AUS). The boundaries chosen for these regions are depicted in Figure 12.3(a) (for IND) and Figure 12.3(c) (for CHI, PHI, and AUS). The seasons used in computing the anomalies (i.e., JJA(0) for IND, and DJF(0/1) for CHI, PHI, and AUS) correspond to those periods in the ENSO cycle when the rainfall signals attain maximum strength in the respective sites (Figure 12.3).

The observational data points displayed in Figure 12.4 confirm the dry conditions in the IND, PHI, and AUS regions, and wet conditions in CHI in a majority of the warm events. These precipitation anomalies are reversed during most of the cold events. The consistency among individual warm and cold events is particularly strong for the PHI region (Figure 12.4(e,f)). A notable exception to the general relationship between ENSO and monsoon rainfall over IND is seen during the warm events in the post-1980 era (Figure 12.4(a)), when some of the observational data sets indicate wet anomalies in that region. Such occurrences evidently contribute to the weakened correlation between ENSO and Indian rainfall during the past two decades, as pointed out by Krishna Kumar *et al.* (1999b). For a given region considered in Figure 12.4, the estimates by the GPCP and CMAP data sets in each event are mostly in good agreement with each other. The corresponding estimates by Dai's data set deviate noticeably from the GPCP/CMAP values in some cases, partially due to the incorporation of other sources of information in the latter data sets, and to the voids in Dai's station data over the maritime portion of the four regions examined here.

The observational rainfall estimates in Figure 12.4 are displayed in juxtaposition with corresponding areal averages based on output from two experiments with a 30-wavenumber, 14-level GCM developed at the Geophysical Fluid Dynamics Laboratory (GFDL). Details of these integrations have been given by Alexander *et al.* (2002) and Lau and Nath (2003). In the control (CTRL) experiment, the observed monthly SST variations throughout the 1950–1999 period have been

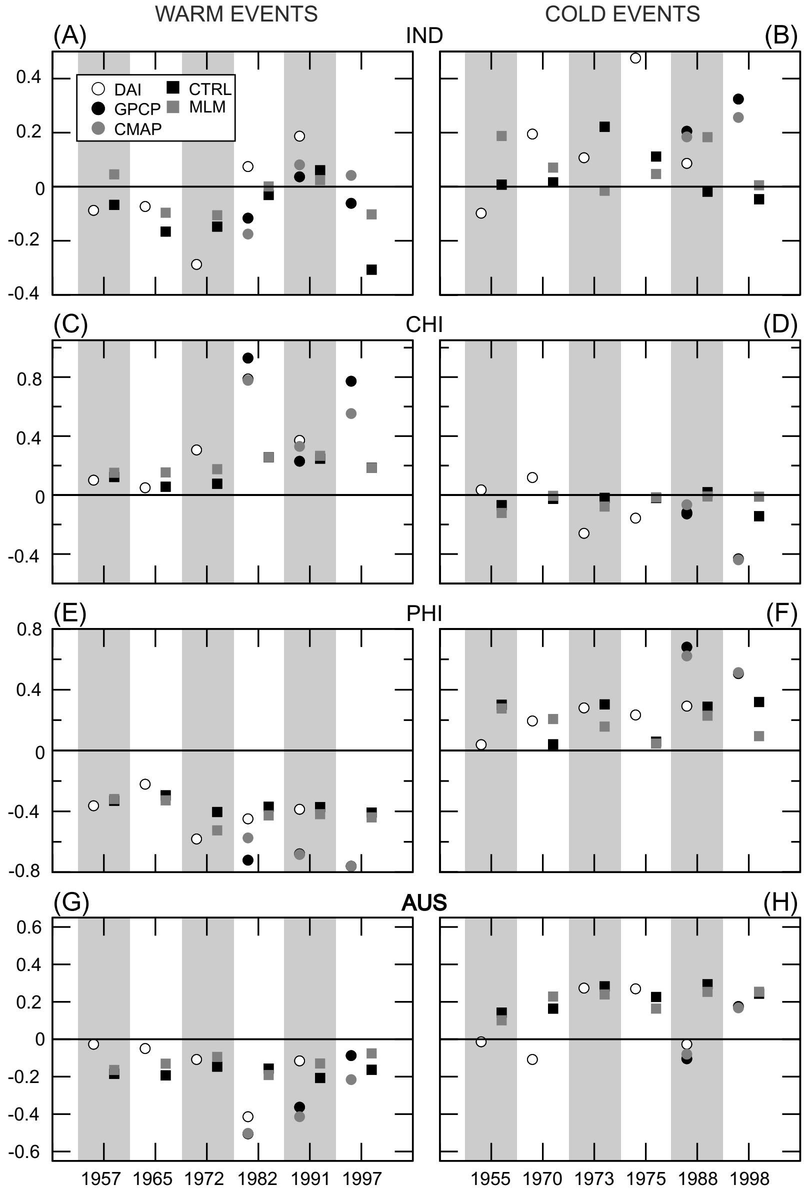

Figure 12.4. Areal averages of precipitation anomalies during six selected warm ENSO events (*left panels*) and six cold events (*right panels*), for (a, b) India–Arabian Sea (IND) in JJA(0); and for (c, d) southern China–East China Sea (CHI), (e, f) the Philippine Sea (PHI), and (g, h) northern Australia (AUS), all in DJF(0/1). Results are computed based on various observational data sets (Dai, GPCP, and CMAP, indicated by circles) and on output from the MLM and CTRL experiments (indicated by squares). All results are expressed as fractions of the local climatological mean precipitation. The regions used in computing the areal averages are indicated in Figure 12.3(a) (for IND) and Figure 12.3(c) (for CHI, PHI and AUS).

inserted in the deep tropical eastern/central Pacific (DTEP; 5°S–5°N, 172°E to the South American coast). The SST at all ocean points outside of the DTEP has been fixed at their climatological seasonal values, with no interannual variability. In the mixed-layer model (MLM) experiment, the identical sequence of temporally varying SST changes has been prescribed in the DTEP, whereas the SST conditions at remaining ice-free maritime sites have been predicted using a variable-depth oceanic mixed-layer model. A detailed description of this mixed-layer model has been provided by Alexander *et al.* (2000). Altogether eight and sixteen independent integrations have been performed using the designs for the CTRL and MLM experiments, respectively. The results presented in this chapter are based on ensemble averages over these individual realizations.

Inspection of the data entries in Figure 12.4 for the CTRL and MLM experiments reveals considerable model skill in reproducing the observed ENSO-monsoon rainfall relationships in various regions. The fidelity of the simulations is particularly evident in the PHI and AUS regions, where the polarity of model-generated precipitation anomalies is the same for all six warm events for each site (Figure 12.4(e) and (g)), and is reversed for all six cold events (Figure 12.4(f) and (h)). In accord with the weakening of the correlation between ENSO and the observed Indian monsoon rainfall in recent decades, the model simulations for the IND region (Figure 12.4(a,b)) also exhibit a relatively broader scatter among the individual events. For the CHI, PHI, and AUS regions, the precipitation anomalies for a given event in the CTRL and MLM experiments are mostly in close agreement with each other. This result suggests that the rainfall variations in these regions during the DJF(0/1) season are mostly forced by SST anomalies in the DTEP region, and that the additional air–sea feedbacks incorporated in the MLM experiment do not significantly alter the remote response to this primary forcing originating from the tropical Pacific. On the other hand, the more notable differences between the CTRL and MLM signals for some ENSO events in the IND region (Figure 12.4(a,b)) are indicative of a stronger role of local air–sea interactions in rainfall variability at that location.

In view of the demonstrable capability of the GCM to mimic the impact of ENSO on monsoon rainfall, we shall henceforth make use of the output from the CTRL and MLM experiments to delineate the mechanisms linking ENSO to the variability of the coupled atmosphere–ocean system in the Asian–Australian monsoon region.

12.3 ENSO-RELATED VARIABILITY IN THE INDIAN OCEAN BASIN

12.3.1 Atmospheric and SST anomalies

The typical atmospheric and oceanic changes in the Indian Ocean (IO) sector during ENSO episodes are illustrated in Figure 12.5 (color section), which shows the warm-minus-cold composites of the 850-hPa vector wind (arrows) and SST (shading) fields for the JJA(0), SON(0), and DJF(0/1) seasons. These patterns are based on the six warm and six cold events examined in Figure 12.4, and have been constructed using the NCEP data (left panels) and model data generated in the MLM experiment (right panels).

The most coherent atmospheric signal in both the reanalysis and MLM patterns for JJA(0) (Figure 12.5(a,b), color section) is the anticyclonic 850-hPa circulation anomaly that prevails over the Arabian Sea and the surrounding land areas. This feature is seen to extend toward the Bay of Bengal and Indo-China during the SON(0) season (Figure 12.5(c,d)). The easterly wind anomalies over much of the northern IO that accompany the anticyclone oppose the climatological westerlies over this region (see climatological streamline charts shown in Lau and Nath (2000)), and is indicative of below-normal intensity of the summer monsoon circulation over south Asia during warm ENSO events. Also evident in the composite patterns for the northern summer and fall seasons is the emergence of warm SST anomalies in both the Arabian Sea and Bay of Bengal. As noted in Lau and Nath (2000, 2003), two factors contribute to these SST changes. First, reduction in the monsoon intensity during warm ENSO episodes is accompanied by lowered wind speeds over these oceanic regions, which result in less latent and sensible heat loss to the atmosphere. Second, the decreased amount of cloud cover due to the generally dryer conditions in these areas (Figure 12.3(a,b)) leads to more heating of the ocean surface by incoming solar radiation. The observed SST increase near the Somali and Arabian coasts could also be partially caused by the reduced oceanic upwelling associated with weakened monsoon flows, an effect that is not considered in the MLM experiment.

Another noteworthy SST signal in the JJA(0) and SON(0) seasons is the cold anomaly that develops off the Sumatra–Java coasts. This feature is collocated with low-level south-easterly or easterly wind anomalies, which are parallel to the local climatological circulation (e.g., see Lau and Nath, 2000). Budget analysis performed by Lau and Nath (2003) indicates that the increased surface wind speeds in this region lead to increased latent and sensible heat loss from the ocean, as well as deepening of the local oceanic mixed layer. Both effects are conducive to SST cooling. The stronger upwelling driven by the intensified winds along the shores of Sumatra and Java, and by anomalous easterlies in the eastern equatorial IO, could further enhance the observed cold SST anomaly in those sites. This oceanic cooling in the eastern IO is in distinct contrast with the warming in the western portion of the basin described in the preceding paragraph. The occurrence of this zonal asymmetric SST anomaly pattern, which is most evident in SON(0), has been noted by Webster *et al.* (1999) and Saji *et al.* (1999). A corresponding east–west contrast in the precipitation field is also discernible in the same season (Figure 12.3(b)). These investigators have attributed this mode of variability mostly to processes operating within the IO sector. However, the appearance of the same pattern in the ENSO composites shown in Figure 12.5(c,d; color section) indicates that the remote forcing from the tropical Pacific could also influence the SST field in the IO.

During the DJF(0/1) period (Figure 12.5(e,f), color section), the cold SST anomaly in the tropical eastern IO is no longer discernible. The climatological low-level flow over this region switches from easterly to westerly in this season (e.g., Lau and Nath, 2000; Schott and McCreary, 2001), so that the easterly wind anomalies would lead to reduction of both wind speed and heat loss from the ocean. The below-normal rainfall in this area results in decreased cloud amounts and increased incoming solar radiation. Both effects contribute to warming in the eastern IO. The processes contributing to the reversal of the SST tendency in the tropical eastern IO during the JJA(0)–DJF(0/1) period have also been noted by Nicholls (1984), Hendon (2003), Li *et al.* (2003), and Shinoda *et al.* (2004). The SST anomalies in other parts of the IO remain positive in DJF(0/1), with notable amplification in the Bay of Bengal and the central IO between, 20°S and 30°S. These two sites are overlain by wind anomalies that oppose the local climatological circulation, which is oriented south-westward over the Bay of Bengal, and north-westward over southern IO during the DJF season. The resulting decrease in wind speed and in oceanic heat loss are hence consistent with the more pronounced SST warming over these areas. The basin-wide atmospheric circulation anomaly in DJF(0/1) is characterized by strong easterlies along the tropical IO, and a pair of anticyclonic cells straddling the equator, with centers located over the north-western Australia and the South China Sea. The MLM pattern (Figure 12.5(f)) is suggestive of a tendency for the anomalous wind vectors to be directed from the eastern equatorial IO (where SST is near normal) to the primary sites of warm SST anomalies (i.e., western IO, Arabian Sea, Bay of Bengal, and central IO south of 20°S). Composite SST patterns for the MAM(1) and JJA(1) seasons, as shown in Alexander *et al.* (2004) and later in this chapter (Figure 12.7(b,c), color section) using observed and MLM data, respectively, indicate that the principal warm SST anomalies in the IO basin persist through the northern summer season of year(1).

There is general agreement between the broad-scale features of the wind anomaly pattern deduced from the reanalysis data and the model output. The magnitude of the wind vectors in the JJA(0) and SON(0) seasons is noticeably larger in the reanalysis results than in the model simulation (note the different scales used plotting the left and right panels of Figure 12.5, color section). The discrepancies between the observed and model patterns are more evident in the SST field. During the SON(0) season, the cold anomaly simulated in the MLM experiment (Figure 12.5(d)) extends too far to the central and western IO as compared with the observed pattern (Figure 12.5(c)). The observed warming of the equatorial waters in the central and eastern IO in DJF(0/1) is also less evident in the model pattern (compare Figure 12.5(e) with (f)). Such discrepancies may partially be attributed to the effects of ocean dynamics that are not incorporated in the MLM experiment.

12.3.2 Atmospheric response to anomalous tropical heating

We next evaluate the extent to which the atmospheric wind anomalies depicted in Figure 12.5 (color section) may be attributed to remote forcing by ENSO-related precipitation changes in the tropical zone. By invoking analytic solutions presented by Matsuno (1966) and Gill (1980) for tropical circulations induced by heating, several investigators (e.g., Chen and Yen, 1994; Kawamura, 1998; Lau and Nath, 2000; Wang *et al.*, 2003) have interpreted the low-level anticyclones over south Asia and the southern IO (Figure 12.5) as Rossby wave responses to anomalous cooling over the Indonesian archipelago and the equatorial western Pacific. The latter heat sink is in turn linked to the eastward displacement of the Walker circulation during warm ENSO events, which results in below-normal precipitation in the equatorial zone between 90°E and 150°E (Figure 12.3(a–c)). The atmospheric responses to diabatic heating in various monsoon regions have also been examined by Rodwell and Hoskins (2001).

The effects of the altered diabatic forcing over the western Pacific on the atmospheric flow pattern have been demonstrated by Wang *et al.* (2003) and Lau *et al.* (2004) using solutions of stationary wave models for the JJA(0) and DJF(0/1) seasons, respectively. We hereby adopt the same approach to diagnose the atmospheric response to heating anomalies in the Indo-Pacific region. The stationary wave model examined in the current work is based on the dynamical framework of the GFDL GCM used for the MLM experiment (see Ting and Yu, 1998, for details). The 2-D basic state in the latitude–height plane has been incorporated in the stationary wave models using climatological data for the corresponding season. Anomalous diabatic forcing, as obtained from the warm-minus-cold composites of the heating rates generated in the MLM experiment for JJA(0) and DJF(0/1), has been applied to the stationary wave model, and linear steady-state solutions were then computed. Only composite cooling rates in the Indonesia–western Pacific sector have been considered in the calculations to be presented here. During JJA(0), there exists some cancellation between the solution in Figure 12.6(a) and the corresponding response to the enhanced heating over the central equatorial Pacific during warm ENSO events (not shown). In DJF(0/1), the response to central Pacific heating is much weaker than that displayed in Figure 12.6(b) (see Lau *et al.*, 2004).

The vector wind response in the lower troposphere (sigma level 0.935) of the stationary wave model (arrows) to the prescribed cooling anomaly (contours) is displayed in Figure 12.6, for (a) JJA(0) and (b) DJF(0/1) seasons. Some similarity exists between the wind anomaly patterns in Figure 12.5 (color section) and the stationary wave model solutions for the corresponding season in Figure 12.6, thus suggesting that the diabatic forcing in the vicinity of the tropical western Pacific accounts for some of the circulation changes appearing in the IO sector of the GCM atmosphere. Of particular interest is the generation by the stationary wave model of strong anticyclonic flow over south Asia during JJA(0), which is located to the north-west of the prescribed cooling center over the equatorial western Pacific (Figure 12.6(a)). The center of this anticyclone in the stationary wave solution is situated to the east of the corresponding features in Figure 12.5(a) and (b) (color section). Analogously, the anticyclonic centers over the east Asian and northern Australia–southern IO during DJF(0/1) are situated to the north-west and south-west, respectively, of the extensive heat sink over the tropical and subtropical western Pacific (Figure 12.6(b)). These spatial relationships between the atmospheric flow pattern and the heating field are in accord with those for the Rossby wave solutions in the Matsuno–Gill models. It is also noteworthy that the pronounced

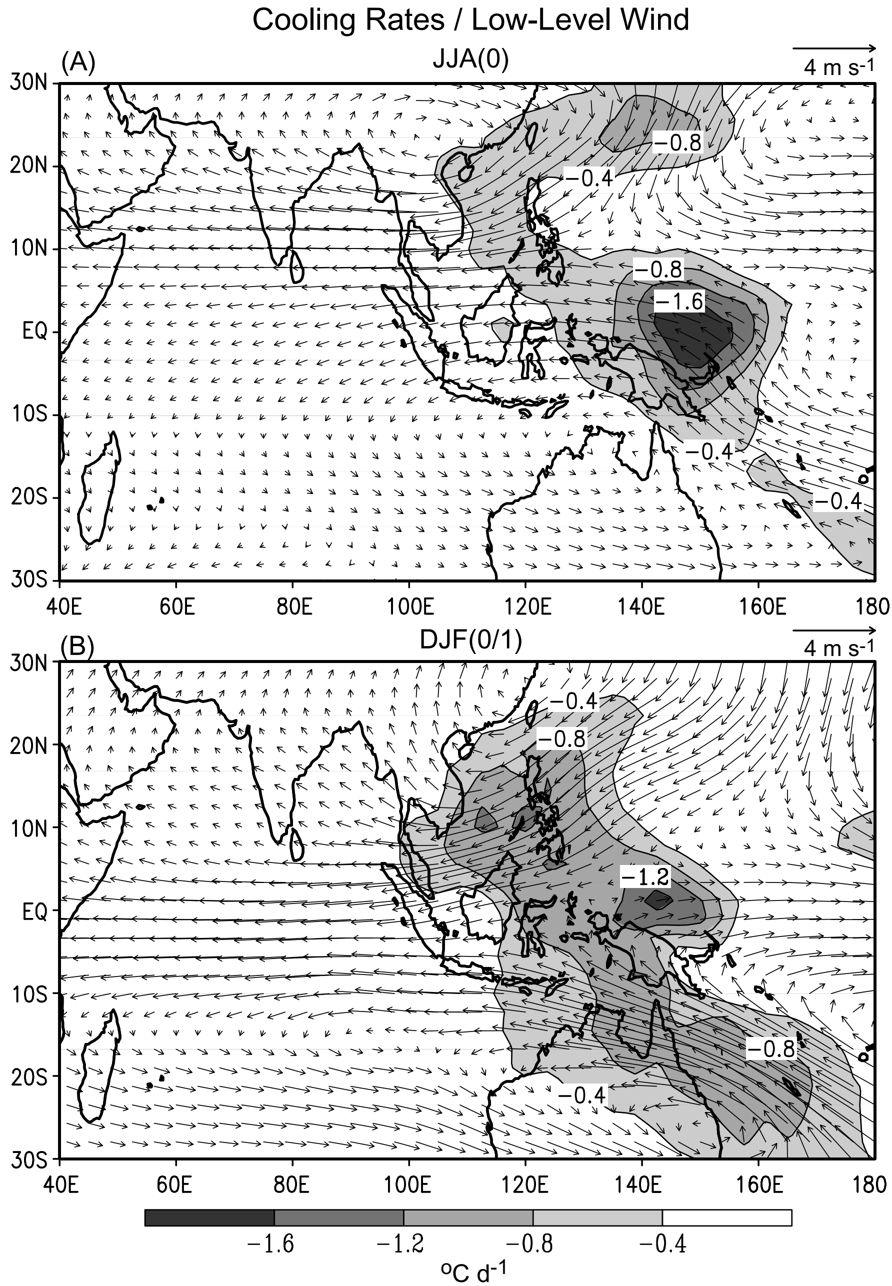

Figure 12.6. Distributions of wind response (arrows) in the lower troposphere (sigma level 0.935) to cooling in the Maritime Continent and western Pacific, as simulated by a linear stationary wave model with specified 2-D basic state for (a) JJA and (b) DJF. Warm-minus-cold composites of the cooling rates from the MLM experiment are shown in contours and shading (units: $^{\circ}C\,d^{-1}$) for the (a) JJA(0) and (b) DJF(0/1) periods.

near-equatorial easterlies over the IO basin, as seen in the observed and MLM patterns in Figure 12.5, are also evident in the patterns based on the stationary wave model.

The stationary wave solution for JJA(0) in Figure 12.6(a) exhibits a considerable degree of asymmetry about the equator, with strong anticyclonic signals over south Asia, and much weaker responses over the southern hemisphere. This interhemispheric difference in the response patterns during the boreal summer is partially related to the spatial distribution of the cold sink imposed in this season (see contour pattern in Figure 12.6(a)). As noted by Wang *et al.* (2003), another contributing factor to the asymmetry of the cooling-induced circulation anomaly about the equator is the distinct vertical structure of the basic state over the IO region in JJA. During this season, strong easterly shears with increasing height occur over the south Asian monsoon region north of the equator. The normal mode analyses by Wang and Xie (1996) and Xie and Wang (1996) indicate that moist Rossby wave responses are amplified in the presence of easterly vertical shears in the background flow, and vice versa. These theoretical results are consistent with the relative strength of the response signals in the two hemispheres for the JJA season.

12.3.3 Atmosphere–ocean feedbacks in the IO basin

The cumulative evidence presented in Figures 12.3–12.6 highlights the following chain of processes linking ENSO events in the tropical Pacific to SST variations in the IO basin: eastward displacement of the Walker circulation during warm ENSO episodes, reduced precipitation and latent heat release over Indonesia and the tropical western Pacific, generation of atmospheric Rossby wave responses to the north-west and south-west of the heat sink, and atmospheric driving of the SST field in the IO sector through modulation of surface latent and radiative fluxes as well as ocean currents. Hence the atmospheric circulation serves as a 'bridge' communicating the ENSO forcing in the DTEP to oceanic changes elsewhere (Klein *et al.*, 1999; Alexander *et al.*, 2002). In the early stages of ENSO development (i.e., during the JJA(0)–SON(0) period), the impact of the atmospheric bridge is not yet fully felt in the IO sector, and the SST anomalies in that basin are still relatively weak (Figure 12.5(b) and (d), color section). As noted in Section 12.3.1, the SST changes over the IO in this period may be interpreted as the oceanic response to ENSO-related atmospheric driving; whereas the emerging SST anomalies do not exert a strong influence on the atmospheric circulation. It is only after the ENSO-induced SST anomalies are better established (in DJF(0/1) and thereafter) that the feedbacks of the oceanic changes on the atmosphere become more evident. We shall henceforth focus on such feedbacks in the period starting from DJF(0/1). The nature of these effects could be delineated by diagnosing the model output from the MLM experiment in conjunction with that from the CTRL experiment (Section 12.2). The CTRL experiment has been designed to yield the global atmospheric responses to SST forcing prescribed in the tropical Pacific only. In addition to this 'direct' response to ENSO, the MLM experiment also incorporates two-way, air–sea interactions induced by the atmospheric bridge mechanism outside of the tropical Pacific.

Hence, the atmospheric signals that are associated with the latter feedback processes may be estimated by removing the direct ENSO response as simulated in the CTRL experiment from the MLM data. This strategy has been adopted by Lau and Nath (2000, 2003) and Lau *et al.* (2004) to isolate the impact of ENSO-induced SST anomalies in different parts of the World's oceans on the atmospheric circulation.

The distributions of SST (shading), precipitation (contours), and 850-hPa wind vector (arrows), as obtained by subtracting the warm-minus-cold composites based on output of the CTRL experiment from the corresponding composites based on MLM data, are shown in Figure 12.7 (color section) for (a) DJF(0/1), (b) MAM(1), and (c) JJA(1) seasons. On the basis of the above arguments, we shall henceforth interpret these 'MLM-minus-CTRL' patterns in terms of local feedbacks between the atmospheric and SST anomalies in the IO basin. During DJF(0/1) (Figure 12.7(a)), the 850-hPa wind vectors are directed toward the warm SST anomalies in the Arabian Sea, Bay of Bengal, and the central portion of the southern IO. The predominantly south-westerly (north-westerly) wind vectors over the northern IO (south-western IO) are oriented against the climatological flow in these regions (see Lau and Nath, 2000), so that the decreased surface wind speeds over these regions would reduce oceanic heat loss to the atmosphere, thus resulting in SST warming. These results are indicative of positive feedbacks in the atmosphere–ocean coupled system in the IO (i.e., wind responses to SST changes in the northern and southern IO tend to reinforce the SST anomalies in these locations).

Also evident in Figure 12.7(a) (color section) is a cyclonic (clockwise wind vectors) pattern centered at 20°S–55°E. The placement of this feature to the west of the warm SST site in the southern IO suggests that it is an atmospheric response to the thermal forcing associated with the latter oceanic anomaly (e.g., see arguments by Hoskins and Karoly, 1981). This circulation signal is seen to persist through the MAM(1) season (Figure 12.7(b)), with its center being shifted south-westward relative to the DJF(0/1) position. The appearance of the anomalous cyclonic center over the south-western IO in the boreal winter and spring of year(1), and its effects on local Ekman upwelling and ocean circulation, have been noted in the observational study of Xie *et al.* (2002).

The pattern in Figure 12.7(a) indicates enhanced cyclonic circulation and precipitation over the warmer waters in the vicinity of Sumatra in the MLM run as compared with the CTRL run. The presence of a positive precipitation signal in that region in the DJF(0/1) phase of the observed warm ENSO events has previously been studied by Chang *et al.* (2004c). The model finding presented here illustrates that air–sea feedbacks outside of the tropical Pacific could contribute to the rainfall signal over the western part of the Maritime Continent during ENSO events.

The warm SST anomalies in the Arabian Sea, Bay of Bengal, and South China Sea attain maximum amplitudes in MAM(1), and are still discernible in JJA(1) (Figure 12.7(b,c), color section). The corresponding wind vector and precipitation patterns in these seasons are characterized by south-westerly flows toward these warm ocean sites, and positive precipitation centers over India and Indo-China. These results imply that air–sea coupling in the IO basin tends to strengthen the summer monsoon over south Asia in year(1) of warm ENSO events. This perturbation is in opposition to that simulated in the MLM experiment in year(0), when the monsoon flow over the same region is weaker than normal (Figure 12.5(b)), and dry conditions generally prevail (see data points for MLM in Figure 12.4(a,b)). The reversal of the precipitation anomalies over the Indian monsoon region from JJA(0) to JJA(1) is also supported by the observational data shown in Figures 12.3(a,e). This tendency for some monsoonal variations to switch polarity from one year to the next may be viewed as one facet of the tropospheric biennial oscillation (TBO) (e.g., see Meehl, 1997; Meehl and Arblaster, 2002b). The model evidence shown in this section indicates that biennial changes of the south Asian monsoon may partially be the consequence of the following chain of processes: remote responses to ENSO forcing in JJA(0) (Figure 12.6(a)), generation of SST anomalies in the IO basin during SON(0)–DJF(0/1) (Figure 12.5(d,f), color section) by the atmospheric bridge, and feedback of these oceanic perturbations on the atmosphere in MAM(1)–JJA(1) (Figure 12.7(b,c)).

12.4 ENSO-RELATED VARIABILITY OVER EAST ASIA, AUSTRALIA AND THE WESTERN PACIFIC

The essential atmospheric and oceanic changes in the eastern portion of the Asian–Australian monsoon system during ENSO events are summarized in Figure 12.8 (color section), which shows the warm-minus-cold composites of (a, b, e, f) surface wind vector (arrows) and SST (shading), and (c, d, g, h) SLP (contours) and precipitation (shading), for the (top half) DJF(0/1) and (bottom half) MAM(1) seasons. Results based on NCEP and MLM output are displayed in the left and right panels, respectively. The most prominent features over the western Pacific are organized about the pair of positive SLP anomalies over the Philippine Sea and off the eastern Australian seaboard. These pressure perturbations are coincident with anomalous anticyclonic flows at the surface (see arrow patterns) and below-normal rainfall. The stationary wave solution presented in Figure 12.6(b) indicates that the two high-pressure centers are responses to the heat sink over the tropical western Pacific. Comparison between the MLM composites for the DJF(0/1) and MAM(1) periods reveals considerable eastward displacement with time of the SLP, wind, and precipitation anomalies associated with two anticyclones.

As has been pointed out by Wang *et al.* (2000), the evolution of the anomalous SST pattern is closely related to changes in the local surface circulation. For instance, the south-westerly wind anomalies to the west of the Philippine Sea anticyclone (hereafter abbreviated as PSAC) oppose the climatological north-easterly winter monsoon over that region (see Lau and Nath, 2000). The reduction in the wind speed leads to suppression of oceanic heat loss and warm SST anomalies. Conversely, the intensification of the north-easterly monsoon flow by the wind anomalies to the east of this anticyclone brings about oceanic cooling in the subtropical north-western Pacific. The resulting SST anomaly pattern with characteristic east–west contrast is seen to migrate eastward from winter to spring, in concert with the movement of the overlying PSAC. Analogous considerations of the super-position of the local wind anomalies on the south-easterly time-mean flow also account for the SST anomaly pattern off the eastern coast of Australia. Besides its impact on the underlying SST field, the weakening of the dry winter monsoon over east Asia in DJF(0/1) is also accompanied by above-normal precipitation over south China and the East China Sea. This wet anomaly is seen to persist through the following spring season. The large-scale signals in the NCEP and MLM patterns are in general agreement with each other.

The strong relationship between the temporal evolution of the anomalous PSAC and the SST pattern over the South China Sea and north-western Pacific is illustrated in greater detail in Figure 12.9, which shows the time–longitude variations of the warm composites of SLP (contours) and SST (shading), as computed using monthly mean data from (a) NCEP and (b) MLM that are averaged over latitudes between 10° and 20°N. In the summer of year(0), below-normal SLP and SST prevail over the subtropical north-western Pacific. The amplitude of the negative SLP anomaly centered near 170°E is markedly stronger in the MLM experiment as compared to observations. For both model and observations, the positive SLP anomaly emerges in September(0) over the South China Sea. The NCEP data indicate that the eastward migration of this feature is most notable during the autumn of year(0); whereas the corresponding signal in the MLM pattern exhibits the strongest zonal movement in the DJF(0/1) and MAM(1) periods. This difference between model and observation in the timing of the spatial displacement of the pressure anomaly is also discernible in the patterns of Figure 12.8 (color section), in which the eastward march of this anomaly during the winter and spring seasons is more evident in the MLM than in the NCEP composites. The patterns in Figure 12.9 indicate that, for both observed and simulated data, the establishment of the PSAC is followed by oceanic warming to its west, and cooling to its east. Corresponding to the zonal evolution of the SLP anomaly, this warm–cold SST couplet also spreads eastward with time, particularly in the MLM pattern. The SST changes attain maximum strength in January(1)–February(1) in the observed pattern, and in March(1)–May(1) in the MLM experiment.

12.5 EVOLUTION OF THE PHILIPPINE SEA ANTICYCLONE ANOMALY

The results in Figure 12.9 highlight the following stages of evolution of the PSAC: the pre-establishment phase in the summer of year(0), the onset phase in the autumn of year(0), and the fully developed phase that persists from late winter of year(0/1) to the following spring. We proceed to examine the phenomena prevailing in each of these stages by further diagnosing the MLM output for the corresponding time periods, so as to identify the processes contributing to the development of the PSAC.

12.5.1 Atmospheric preconditions in JJA(0)

The atmospheric environment prior to the onset of the PSAC is depicted by the warm-minus-cold composites in Figure 12.10, for (a) 850-hPa wind vector,

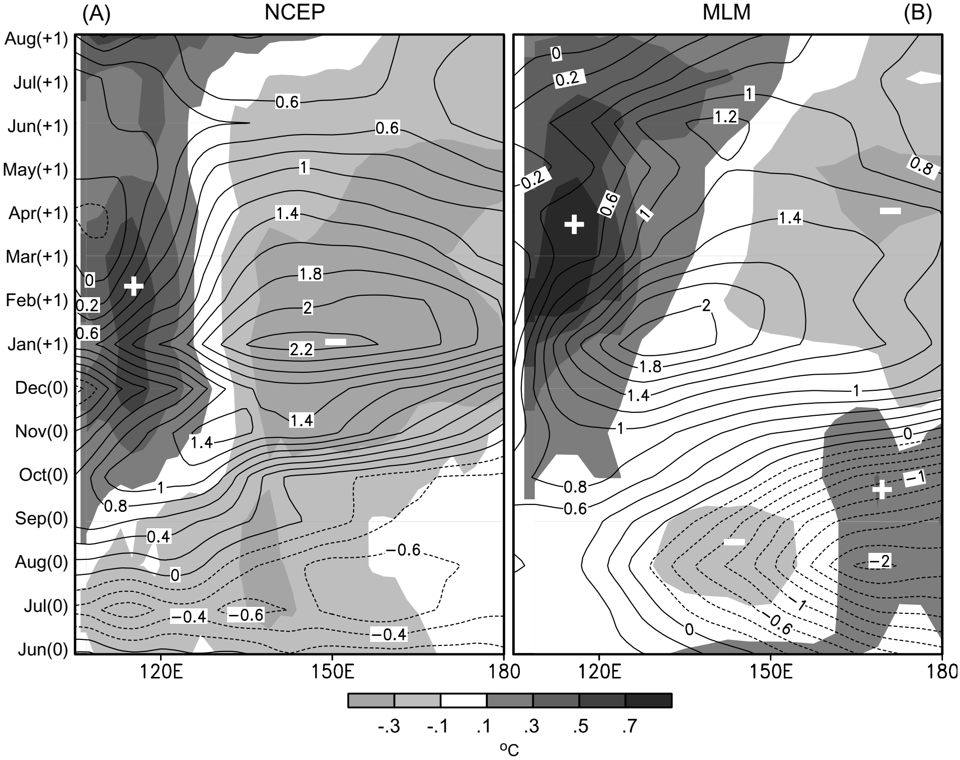

Figure 12.9. Time–longitude distributions of the warm composites of SLP (contour intervals: 0.2 hPa) and SST (shading), as computed by averaging (a) NCEP data and (b) output from the MLM experiment over the zonal belt between 10°N and 20°N.

(b), 200-hPa wind vector, (c) 500-hPa height (contours) and precipitation (shading), and (d) surface air temperature for the MLM simulation in the JJA(0) season. The circulation in this phase of the ENSO cycle is characterized by an anomalous 850-hPa cyclone and 200-hPa anticyclone over the western North Pacific. Results from stationary wave modeling (Lau and Nath, 2005) analogous to those presented in Figure 12.6 indicate that these features are essentially Rossby wave responses to enhanced condensational heating over the equatorial central Pacific, which results from displacement of the rising branch of the Walker circulation to that region during warm ENSO events. Also evident in Figure 12.10(b) is the cyclonic anomaly over northeastern Asia and the intensified westerlies within the 30°–40°N zone over east Asia. These upper tropospheric signals are indicative of a deepened trough and southward displacement of the climatological jetstream over that region. The relationships between these features at the upper level and other circulation changes over east Asia are further examined in the following paragraphs.

The northerly or north-westerly wind anomalies at 850 hPa off the south-eastern Asian seaboard (Figure 12.10(a)) are in opposition to the climatological south-

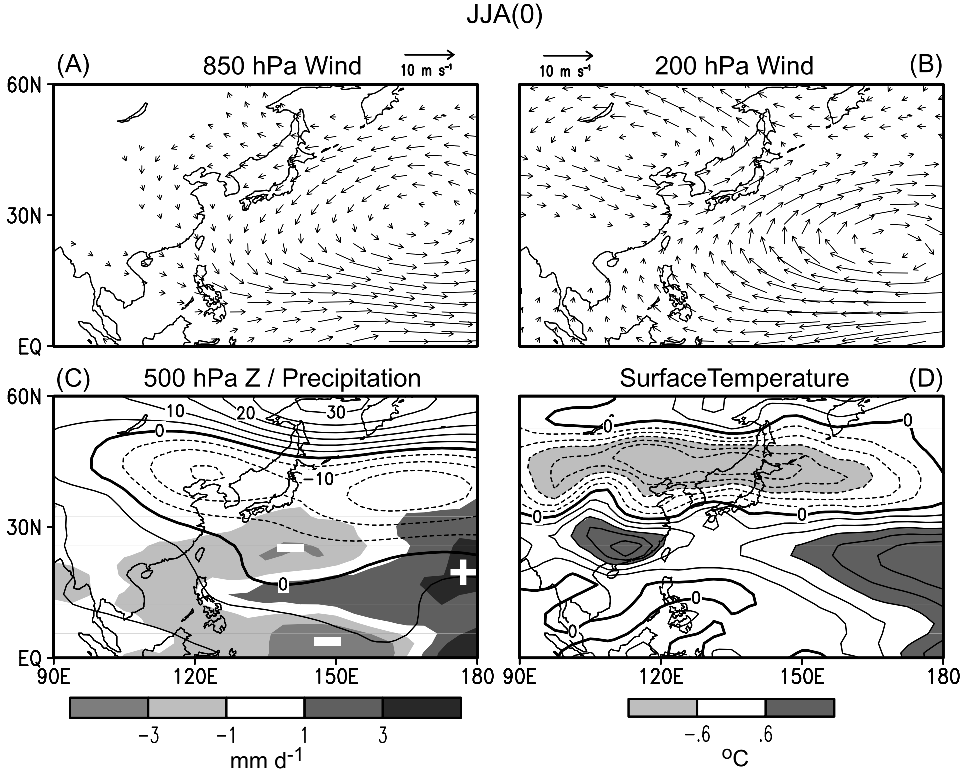

Figure 12.10. Distributions of the warm-minus-cold composites of the anomalous (a) 850-hPa vector wind; (b) 200-hPa vector wind; (c) 500-hPa height (contour interval: 5 m) and precipitation (shading); and (d) surface air temperature (contour interval: 0.2°C). Results are based on output for the JJA(0) season from the MLM experiment. Solid and dashed contours indicate positive and negative values, respectively.

easterly monsoon flow in that region. This weakening of the monsoon circulation is seen to be accompanied by below-normal precipitation over the area in 15°–25°N 120°–150°E (shading in Figure 12.10(c)). The contour pattern in Figure 12.10(c) shows an elongated negative 500-hPa anomaly extending eastward from northern China to the western Pacific, and a positive anomaly centered over the Sea of Okhotsk. The spatial relationship between the 500-hPa height pattern and the suppressed rainfall over the western Pacific in the 15°–25°N zone is reminiscent of that between the changes in atmospheric circulation and convective activity associated with the summertime 'Pacific–Japan' pattern documented by Nitta (1987), who has interpreted such a relationship in terms of Rossby wave responses to subtropical heat sources and sinks. The most prominent feature in the composite chart for surface air temperature (Figure 12.10(d)) is the cold anomaly that extends from the Asian interior to the North Pacific between 35° and 55°N. These below-normal temperatures are coincident with onshore wind anomalies at 850 hPa, as well as negative geopotential height changes and enhanced vorticity at upper levels.

The corresponding composite charts for the September(0)–October(0) period (not shown) indicate that the summertime surface temperature and circulation anomalies seen in Figure 12.10 are still evident in the early autumn. A similar set of atmospheric conditions preceding the establishment of the PSAC has been described using observational data by Wang and Zhang (2002). These investigators pointed out that such changes in the temperature and circulation environments are favorable for more intense cold air outbreaks over east Asia in early autumn. As will be demonstrated in the following subsection, the formation of the PSAC is often preceded by increased cold air activity in this region.

12.5.2 Synoptic development during PSAC onset

In order to study the synoptic phenomena associated with the onset of the PSAC, the time series of pentadal (five day) averages of selected fields simulated in each of the 16 individual runs of the MLM experiment during the six warm ENSO years (1957, 1965, 1972, 1982, 1991, 1997) have been analyzed. Following a procedure similar to that described in Wang and Zhang (2002), those MLM runs in which the SLP field over the South China Sea and Philippine Sea made a distinct transition to a persistent positive anomaly during the individual warm events were noted, and the specific pentad (hereafter referred to as the 'onset pentad', or T_o) in which this transition occurred was identified for each of such runs and El Niño episodes. By averaging the calendar dates of T_o for individual runs and ENSO events, it is found that the onset of PSAC typically occurs in early October. The SLP anomaly associated with this feature remains to be above-normal for more than four months in all cases considered here. More detailed results of the composite analyses are reported in Lau and Nath (2005).

Composites of various model fields at different temporal leads relative to the onset pentad for individual cases were constructed. The composite patterns for anomalous SLP (contours), surface wind vector (arrows), and precipitation (shading) are displayed in Figure 12.11 (color section) for the pentads centered at (a) 20 days before T_o, (b) 10 days before T_o, and (c) T_o. The most pronounced features in Figure 12.11(a) are the anticyclonic wind pattern and dryness associated with the high-pressure anomaly over the interior of the Asian land mass, the cyclonic flow and wet conditions accompanying the low center over the Philippine Sea, and the prevalent northerly wind anomalies over south-eastern China, the East China Sea, and southern Japan. These atmospheric signals bear a strong similarity to the characteristic behavior of cold air outbreaks over this region. The continental high-pressure anomaly migrates south-eastward to the Philippine Sea, so that anticyclonic flows and dry conditions are established in the latter area at about 4 pentads after T_o (Figure 12.11(c)). This sequence of simulated events during the PSAC onset is in broad agreement to that reported by Wang and Zhang (2002) using observational data. These investigators have pointed out that the deepening of the upper level trough and below-normal air temperature over east Asia in the preceding months (Figure 12.10) constitute a favorable environment for the incidence of cold air outbreaks and the subsequent PSAC formation. They have also considered the

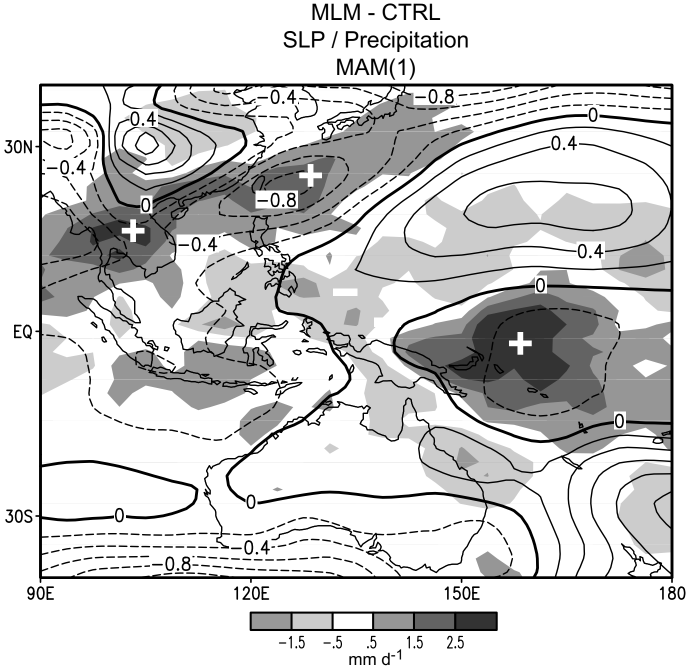

Figure 12.12. Distributions of the differences between the outputs from the MLM and CTRL experiments for the warm-minus-cold composites of SLP (contour interval: 0.2 hPa) and precipitation (shading), for MAM(1). Solid and dashed contours indicate positive and negative values, respectively.

effects of intraseasonal oscillations on the rather abrupt reversal of the wind, pressure, and rainfall anomalies over the Philippine Sea during T_o (Figure 12.11), and the role of atmosphere–ocean interactions in the seasonal dependence of the amplitude of such oscillations.

12.5.3 Air–sea feedbacks in MAM(1)

The simulated SST changes associated with the PSAC development, as illustrated in Figure 12.9(b), are seen to attain maximum amplitudes in MAM(1). By invoking the same reasoning as that applied in interpreting Figure 12.7 (color section), the local feedback of these mature SST anomalies in the western Pacific sector on the overlying atmospheric circulation may be inferred by subtracting the output of the CTRL experiment from that of the MLM experiment. The MLM-minus-CTRL pattern of the warm-minus-cold composites of SLP (contours) and precipitation for the MAM(1) season is shown in Figure 12.12. It is seen that the warm SST anomaly near the 10°–20°N, 110°–140°E region (see Figure 12.8(f), color section) is coincident with negative values of SLP and positive values of precipitation in Figure 12.12, and with a cyclonic low-level wind pattern in Figure 12.7(b). Hence, the air–sea feedbacks attendant to this warm SST anomaly leads to stronger cyclonic development and more abundant springtime precipitation along the climatological rain belt extending north-eastward from the southern coast of China to the western Pacific. Conversely, the cold SST anomaly at 10°–20°N, 150°–180°E is overlain by increased SLP and reduced rainfall in the MLM simulation relative to the CTRL experiment. The SLP couplet over the subtropical north-western Pacific in Figure 12.12, with falling (rising) pressures west (east) of the center of the PSAC (see Figures 12.8(d,h)), is conducive to the eastward migration of this anomaly. The role of air–sea interaction in the spatial displacement of PSAC has previously been noted by Wang *et al.* (2000). Analogous relationships between the SLP and precipitation signals in Figure 12.12 and the SST anomalies in Figure 12.8(f) are also discernible in the south-western Pacific. The SLP rise over the SST anomaly near the Date Line at 30°S due to local air–sea feedbacks (Figure 12.12) may contribute to the eastward tendency of the high-pressure anomaly situated off the eastern Australian coast (Figures 12.8(d,h)).

12.6 IMPACT OF THE ASIAN MONSOON ON ENSO

How ENSO affects the Asian monsoon is better understood than the influence of the Asian monsoon on ENSO. The Asian monsoon covers one-third of the area of the tropics, and is an interactive component of the climate system that can impact the slowly varying lower boundary conditions (Webster *et al.*, 1998). In the following subsections, we shall separately consider the effects of the Indian and western North Pacific monsoons on ENSO.

12.6.1 Effects of the Indian summer monsoon on the development of El Niño

The all-Indian summer rainfall anomaly is most negatively correlated with the central tropical Pacific SST anomaly occurring three to six months after the summer monsoon, implying that a weak (strong) monsoon leads the mature phase of an El Niño (La Niña) event by about one to two seasons (Yasunari, 1990). This result suggests that the Indian summer monsoon could play an active role in ENSO development. However, ENSO is primarily governed by its intrinsic coupled ocean–atmospheric dynamics in the tropical Pacific. Regardless of monsoon variability, the mature phases of El Niño or La Niña tend to occur toward the end of the calendar year. This phase-locking behavior of ENSO is largely determined by the climatological seasonal cycle in the Pacific Ocean (Wang and Fang, 1996; Tziperman *et al.*, 1998; An and Wang, 2000). Thus, the above-mentioned lag relationship between Indian monsoon and ENSO does not necessarily imply a cause-and-effect relationship. Nevertheless, it is plausible that the monsoon variability could potentially add irregularity to ENSO as suggested by Yasunari and Seiki (1992) and illustrated by Wainer and Webster (1996).

If the Indian monsoon indeed exerts an influence on ENSO, what are the key processes involved? Barnett (1984b) noted that ENSO-related westerly anomalies in the western/central Pacific originate from the Indian Ocean. It has been speculated that the eastward propagation of the westerly anomalies from the Indian monsoon region to the Pacific Ocean could serve as a trigger for ENSO events. Moreover, Webster and Yang (1992) showed that when the broad-scale south Asian summer monsoon is stronger (weaker) than normal, the tropical Pacific trade winds are also stronger (weaker) than average. This result suggests that an anomalous Indian monsoon could affect ENSO through changing the trade winds over the Pacific. However, it is difficult to determine the causal relationship between the monsoon and ENSO based on observations alone, because the two components are integral parts of the coupled climate system.

Using the Center for Ocean–Land–Atmosphere Studies (COLA) atmospheric GCM, Kirtman and Shukla (2000) examined a 50-year simulation forced by climatological mean SST to determine the tropical Pacific wind anomalies that are associated with a variable monsoon. Since the ENSO phenomenon and SST anomalies have been excluded in this experiment, the simulated monsoon variability is attributed to atmospheric internal dynamics and atmosphere–land interactions. The monsoon variability was measured by a monsoon rainfall index for south Asia (5°–25°N, 60°–100°E). Figure 12.13 shows the regressed global zonal wind in the model with reference to this monsoon index. During a strong south Asian monsoon, the Walker circulation over the tropical Pacific is indeed enhanced (Figure 12.13(a)). The opposite is true for a weak monsoon. With an enhanced summer monsoon the westerly flows in the vicinity of India and the associated monsoon cross-equatorial gyre in the tropical Indian Ocean are stronger than normal; along the equatorial Indian Ocean there are easterly anomalies (Figure 12.13(b)). Of note is that the easterly wind stress anomalies prevail throughout most of the central and eastern Pacific. This pattern indicates that a weak (strong) monsoon results in a weakening (strengthening) of the trade winds over the equatorial Pacific. This is consistent with the observed contemporaneous ENSO–monsoon relationship (Webster and Yang, 1992), even though the model wind anomalies are independent of ENSO. These numerical experimental results indicate the remote influence of the south Asian summer monsoon on the North Pacific trade winds, mainly by perturbing the climatological Walker cell over the Pacific.

In order to assess the impacts of monsoon variability on ENSO, various types of coupled atmosphere–ocean models have been used. Chung and Nigam (1999) examined the feedback of Asian monsoon on ENSO using a modified Cane–Zebiak (CZ) model (Zebiak and Cane, 1987). The original CZ model has not incorporated monsoon influence. In the modified CZ model, Chung and Nigam used Asian summer monsoon heating anomalies as an additional forcing, and referred to this suite of experiments as 'monsoon runs'. The monsoon heating anomalies were parameterized empirically by using rotated principal component analysis of tropical Pacific SSTs, residually diagnosed tropical diabatic heating, and surface winds during northern summer. In the monsoon runs, the 'interactive' summer heating anomalies in the Asian sector are included, and the ENSO events in the model occur more frequently. The presence of monsoonal interaction results in a broader frequency distribution of ENSO variability and a population shift in amplitude towards stronger El Niño events. The causes for these changes remain to be ascertained.

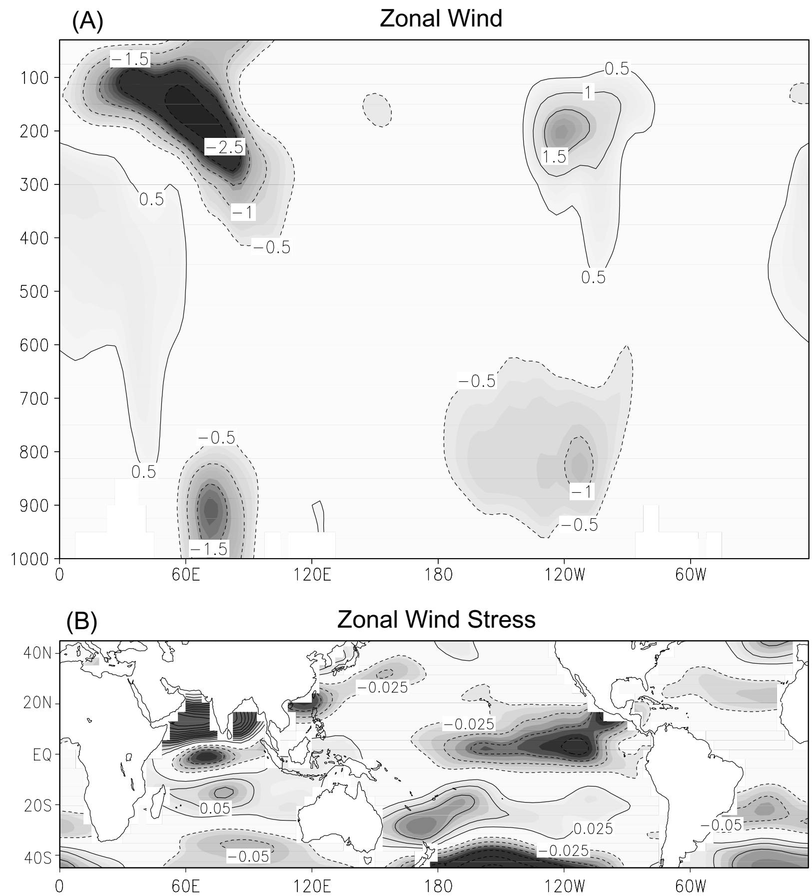

Figure 12.13. Distributions of the linear regression coefficients vs. the south Asian monsoon rainfall index of (a) zonal wind along the equatorial zonal plane (contour interval: $0.5\,\mathrm{m\,s^{-1}}$) and (b) zonal wind stress (contour interval: $0.025\,\mathrm{dynes\,cm^{-2}}$). The amplitude of the pattern corresponds to a rainfall anomaly of $2\,\mathrm{mm\,d^{-1}}$. Solid and dashed contours indicate positive and negative values, respectively. Results are based on output from the COLA atmospheric GCM.
From Kirtman and Shukla (2000).

Kirtman and Shukla (2000) implemented the CZ model by adding the effects of monsoon anomalies derived from their atmospheric GCM experiments mentioned earlier. Note that these monsoon anomalies are independent of ENSO. Their model reproduces realistically the observed lag relationship between ENSO and Asian monsoon. Their results suggest that a variable monsoon enhances ENSO variability, particularly three to six months after the summer monsoon ends. An ongoing warm (cold) event is made even warmer (colder) by a weak (strong) monsoon. They also presented model evidence on the role of a variable summer monsoon as a trigger mechanism for ENSO.

In reality, the monsoon variability is not independent of ENSO. It is thus necessary to distinguish between the impacts of monsoon variability that is related to ENSO from those that are independent of ENSO. Wu and Kirtman (2004) have examined simulations based on the COLA coupled GCM. The effects of monsoon variability that are related and unrelated to ENSO were separated using a composite approach based on simulated SST in the Niño 3.4 region (5°S–5°N, 120°–170°W) and Indian rainfall anomalies. They found that the ENSO-related monsoon variability has significant impacts on warm events but not the cold events. A weak (strong) monsoon enhances (weakens) an ongoing warm event. The monsoon impacts are manifested in the surface zonal wind stress over the western/central equatorial Pacific. The monsoon variability that is unrelated to ENSO also induces noticeable SST anomalies in the equatorial central Pacific in the following winter. A weak (strong) monsoon induces noticeable warm (cold) SST anomalies.

12.6.2 Effects of the western North Pacific monsoon on the turnabout of the ENSO cycle

The delayed oscillator paradigm (Section 12.1) attributes the reversal of the SST anomaly during the mature phases of ENSO cycles to the reflection of oceanic Rossby waves at the western boundary in a delayed manner. However, this mechanism does not explicitly explain the seasonal preference of the ENSO phase transitions. McBride and Nicholls (1983) showed that SST anomalies in the Indonesian region lead those in the eastern Pacific by 4–6 months. They speculated that air–sea interaction in the Indonesian region might in part be responsible for the turnabout of the ENSO.

In Sections 12.4 and 12.5, the observational evidence suggests, and the model results confirm, that a large-scale anomalous surface anticyclone forms over the northern Philippines in September–October during a strong El Niño episode. This feature rapidly develops from autumn to winter over the Philippine Sea and the western North Pacific (WNP), and has been referred to as the Philippine Sea anticyclone (PSAC). To focus on the turnabout during the peak phase of major El Niño events in the last 50 years, we display in Figure 12.14 the time series of ENSO anomalies within an 18-month time window centered in the November of El Niño years. In all six strong events, the maximum SST anomalies in the Niño 3.4 region occurred in boreal winter from November to January. Note that a sharp increase in SLP over the Philippine Sea (10°–20°N, 120°–150°E) preceded the corresponding warm SST peak by about one to three months. In conjunction with this pressure rise, strong anticyclonic surface wind anomalies appear over the Philippine Sea, with enhanced easterlies prevailing north of New Guinea (5°S–10°N, 120°–150°E). This temporal development is different from that inferred from the argument made by Weisberg and Wang (1997a,b), who suggested that the atmospheric high-pressure anomaly could be induced by the cooling due to oceanic upwelling Rossby waves. As discussed by Wang and Zhang (2002) and in Section 12.5, the establishment of large-scale WNP SLP and surface wind anomalies may be attributed to multiple factors: the remote forcing from heat sources and sinks over the tropical Pacific (thus

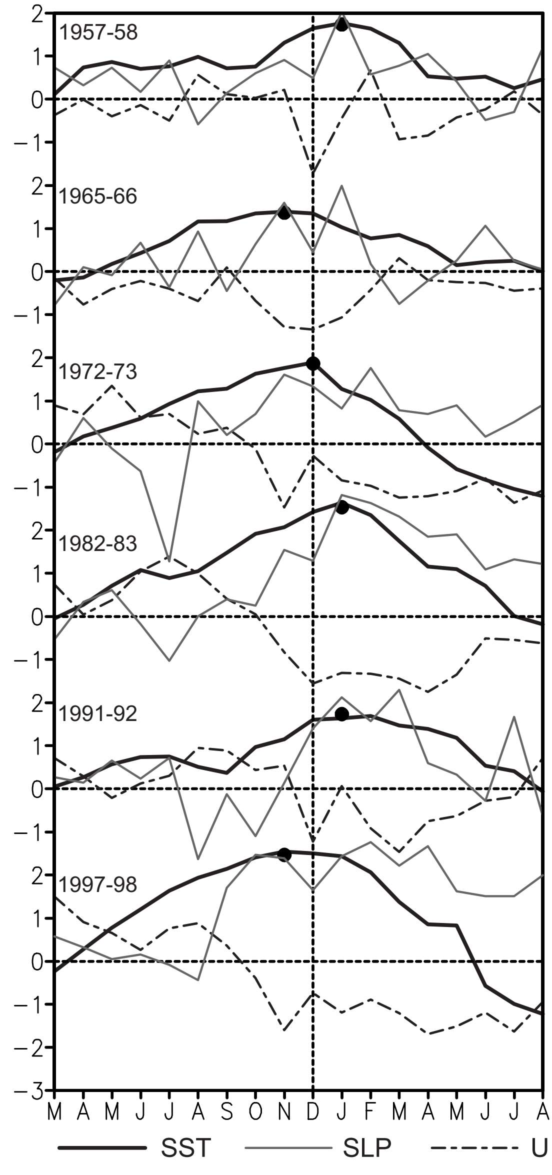

Figure 12.14. Time series of SST and atmospheric anomalies in the WNP during the six strongest warm episodes in the 1950–2003 period. The bold black curves show Niño 3.4 (5°S–5°N, 120°–170°W) SST anomalies (°C) with a heavy circle indicating the time of the maximum SST anomaly. A three-month running mean has been applied to the SST anomalies. The thin gray and dashed curves show, respectively, SLP anomalies (hPa) over the Philippine Sea (10°–20°N, 120°–150°E) and the zonal wind anomalies (m s^{-1}) north of New Guinea (4°S–4°N, 120°–150°E). The abscissa denotes a time window spanning 18 months centered in November of the year during which the warm event matures. Results are based on NCEP data.

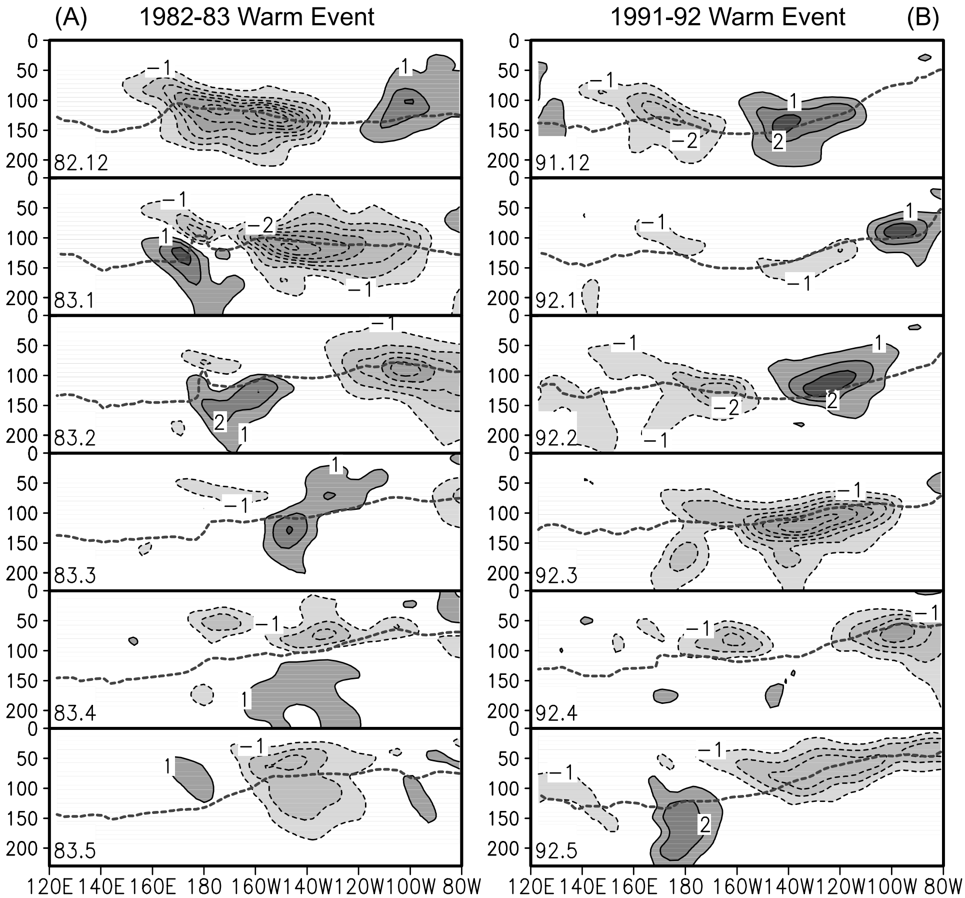

Figure 12.15. Distributions of the monthly mean local rate of change of subsurface ocean temperature in a vertical section along the equator during the turnabouts of the warm events of (a) 1982–1983 and (b) 1991–1992. The ordinate denotes the depth (m) below the surface. The dashed gray lines are the 20°C isotherm which corresponds approximately to the position of the thermocline during each month. The contour interval is 1°C per month. Solid and dashed contours indicate positive and negative values, respectively. Results are obtained from NCEP (Ji *et al.*, 1995).

contributing to the atmospheric bridge effect), extratropical–tropical interaction, and local air–sea interaction.

The sudden emergence of the equatorial easterly anomalies over the western equatorial Pacific may generate oceanic upwelling Kelvin waves that propagate along the equator into the eastern Pacific. This process might significantly perturb the slow variation of the coupled ocean–atmosphere system. Figure 12.15(b) shows local rates of change of the subsurface temperature (referred to as temperature tendency hereafter) in vertical cross sections along the equator for the 1991–1992 warm event. The tendencies of the subsurface temperature result primarily from a vertical displacement of the thermocline. In December 1991, one month before the peak of the warm event, a sudden increase of easterlies over the equatorial western Pacific (Figure 12.14) induced upwelling and raised the local thermocline, causing negative temperature tendencies at the thermocline depth (Figure 12.15). The cold temperature anomaly then migrated eastward at a speed of about 40–50 degrees of longitude per month ($\sim 1.9\,\mathrm{m\,s^{-1}}$) along the thermocline, which suggests an association with upwelling Kelvin waves. The enhancement of anomalous easterlies in February 1992 triggered a second set of cold Kelvin waves that propagated again into the equatorial eastern Pacific. Similar sequences occurred in the 1982–1983 (Figure 12.15(a)) and 1997–1998 (not shown) events. Given that the establishment of the PSAC is about two months earlier than the peak warming of El Niño, this atmospheric feature may play a role in the El Niño turnabout. Once the PSAC is established, it can persist due to positive feedback between atmospheric Rossby waves and the underlying SST dipole through wind-induced evaporation and entrainment, as well as SST-related processes in the atmospheric boundary layer (Wang *et al.*, 2000; see also Sections 12.4–12.5).

Note that both the persistent equatorial easterly anomalies and the fluctuations in these easterlies on intraseasonal timescales contribute to turnabout of the ENSO. A rise of the thermocline in the eastern Pacific due to the arrival of forced upwelling Kelvin waves may initially offset local warming, which would in turn restore the SST gradients and associated equatorial easterlies, thereby triggering a slow thermocline shoaling and SST decrease in the eastern Pacific. The consecutive generation and passage of Kelvin waves, which are forced by the wind fluctuations of a variety of timescales over the western Pacific, may facilitate slow eastward migration of the negative heat content anomalies through the 'fetch extension' mechanism (Kessler and McPhaden, 1995) or zonal advection process (Lukas *et al.*, 1984; Picaut *et al.*, 1996). These heat content anomalies act to restore the thermocline slope from a flat state during El Niño to the normal downward incline from east to west. Model simulation of the 1997–1998 warm event indicates that the upwelling Kelvin waves generated by the easterly anomalies in the western Pacific contributed to the shoaling of the thermocline and the demise of the warming (McPhaden and Yu, 1999).

The development of the PSAC may facilitate the turnabout of ENSO at the end of the calendar year, thereby enhancing the biennial tendency of ENSO events (Wang *et al.*, 1999). The positive feedback between the atmospheric Rossby wave and ocean mixed layer thermodynamics, which plays a key role in development of the PSAC, depends on the presence of climatological north-easterly trades. In the WNP this favorable basic state exists only from late autumn through the following early summer. This seasonal dependence implies that persistent western Pacific easterly anomalies occur preferentially in autumn and winter, thus favoring ENSO turnabout towards the end of the calendar year. This temporal evolution also explains why strong El Niño episodes tend to decay quickly after their mature phase, leading to a biennial tendency of SST variation.

Kim and Lau (2001) have shown that a strong biennial tendency in the ENSO cycle could result from the occurrence of wind anomalies in the western Pacific six months after the SST anomaly peaks in the eastern Pacific. The analysis of Lau and

Wu (2001) suggests that a stronger monsoon–ENSO relationship tends to occur in boreal summer immediately after a peak El Niño and before a pronounced La Niña (i.e., 1983, 1988, and 1998). Their result appears to support the idea that the summer monsoon could influence ENSO variability via the development of the PSAC. The wind forcing associated with the PSAC may be instrumental in enhancing the biennial component of the natural ENSO cycle.

It is worthwhile to mention that the above process operates mostly during strong warm events. There are three moderate warm events (1986–1987, 1968–1969, and 1976–1977) with no reversal of the warming trend at the end of the year of El Niño development. The easterly anomalies in the equatorial western Pacific were not well established and did not persist in the boreal winter of these El Niño years. An important feature common to all three prolonged events is the insufficient strength of the central Pacific warming (Niño 3.4 SST anomaly less than 1.5 standard deviation by the end of the El Niño year), whereas during the six strong events considered in Figure 12.14, the corresponding anomalies were all above 1.5 standard deviations. This suggests that the strong warming in the equatorial central Pacific is probably necessary for the robust establishment of the western Pacific wind anomalies.

12.6.3 Effects of the east Asian winter monsoon in triggering the onset of El Niño

Li (1990, 1996) noticed, for the period 1950–1979, that during the winter prior to El Niño events the Mongolian cold highs tended to be stronger and surface temperature in eastern China tended to be lower than normal. This relationship prompted him to hypothesize that a strong east Asian winter monsoon leads to development of warm episodes in the equatorial eastern Pacific. He argued that strong cold surges associated with a strengthened Asian winter monsoon could penetrate into the South China Sea and equatorial western Pacific regions (Chang and Lau, 1980; Lau *et al.*, 1982), triggering deep convection over the warm tropical ocean. The anomalous convection could further induce westerly wind anomalies to its west, which might excite downwelling oceanic Kelvin waves, thus resulting in anomalous warming in the eastern Pacific.

We have examined the above relationship between the east Asian winter monsoon and ENSO by analyzing NCEP/NCAR reanalyses data for the more extended period of 1948–2002. To avoid the possible influence of decadal variations, we have analyzed the interannual component of the data by retaining variations shorter than 8 years. The leading empirical orthogonal function (EOF) mode of the winter 925-hPa temperature field over east Asia (Figure 12.16(a)) represents the strength of the east Asian winter monsoon, which is characterized by a temperature extremum centered over northern China. The winters with strong winter monsoons are signified by large negative temporal coefficients for this mode. The time series in Figure 12.16(c) indicates that during some years in the pre-1986 period (e.g., 1950, 1956, 1962, 1967, 1971, 1975, and 1985) a strong winter monsoon indeed preceded an El Niño event. However, several prominent La Niña events such as those in 1955, 1970, and 1984 were also preceded by strong winter monsoons. In

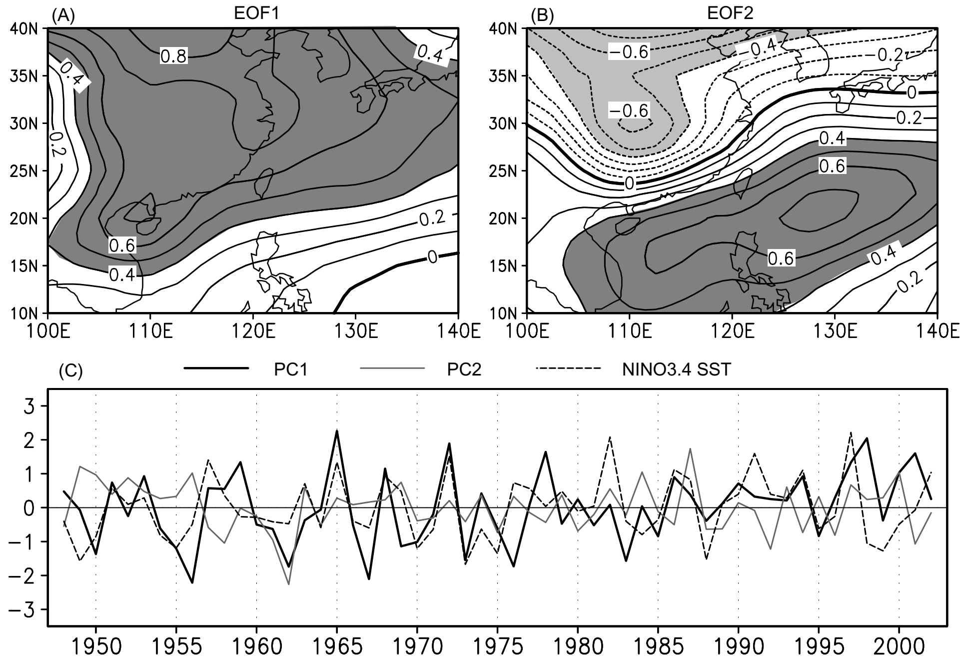

Figure 12.16. Spatial patterns (*top panels*) and temporal coefficients (*bottom panel*) of the first and second EOF modes of the winter (November–March) mean air temperatures at 925 hPa in the 1948–2002 period. The dashed curve in the bottom indicates the Niño 3.4 SST anomalies. In panel (c) the time labels refer to the calendar years for November and December of five-month winter seasons (e.g., '1950' denotes the November 1950–March 1951 period).

addition, prior to the strong El Niño events of 1965 and 1982, the east Asian winter monsoons were normal. It is also worth noting that none of the recent El Niño events (i.e., 1991, 1994, 1997, and 2002), were preceded by stronger than normal winter monsoons. The maximum correlation between the principal component of the leading EOF and SST over the Niño 3.4 region (5°S–5°N, 120°–170°W) occurs at zero lag (0.45), indicating that the Asian winter monsoon weakens during the mature phase of an El Niño and vice versa. However, when the first principal component leads Niño 3.4 SST by one year, the correlation coefficient drops to 0.18, which is not statistically significant at the 95% confidence level. Thus, the principal mode of the east Asian winter monsoon tends to vary with ENSO in tandem, but is not a reliable precursor of ENSO one year in advance. Analysis of unfiltered data yields the same conclusion.

The second EOF pattern (Figure 12.16(b)) is characterized by opposing centers of extremes over central China and the South China Sea. The time coefficient of EOF 2 negatively correlates with Niño 3.4 SST in the following year (−0.27, which is significant at the 95% level), suggesting that a strong (weak) winter monsoon over South China Sea tends to lead warming (cooling) in the central equatorial Pacific by about one year. Cooling over the South China Sea and warming over the middle Yangtze River valley may be a result of mature La Niña forcing and an indication of subsequent transition from a La Niña to an El Niño state. In view of the difficulty in predicting ENSO transition in boreal spring, the search for precursors of the eastern Pacific warming continues to be a meaningful endeavor.

12.7 DISCUSSIONS AND OUTSTANDING ISSUES

The Asian monsoon is one of the most significant signals in the climatological seasonal cycle, whereas ENSO is a principal contributor to interannual variability. The mutual interactions between these two powerful components of the climate system have important implications on monsoon predictability (see pertinent discussion in Chapters 14 and 15 of this book), as well as frequency characteristics and seasonal phase-locking of ENSO episodes. We shall conclude this chapter by discussing outstanding issues that pertain to various facets of the ENSO–monsoon relationship.

12.7.1 Biennial tendency

Perhaps one of the important consequences of the ENSO–monsoon interaction is the resultant biennial tendency of the atmosphere–ocean system. Precipitation records in Asian monsoon regions exhibit a strong biennial tendency (i.e., a strong monsoon tends to be followed by a weak monsoon), and vice versa (e.g., see Figure 12.3). A biennial spectral peak is also discernible in ENSO indices. The biennial tendency of ENSO is particularly linked to turnabouts of El Niño and La Niña episodes from one summer to the following summer, with amplitudes of these episodes peaking during the boreal winter between the two summers. Still more study is required to clarify the origin of the biennial oscillation of the monsoon–ENSO system. However, this phenomenon is likely related to the interplay between the Asian monsoon, the ENSO cycle, and myriad of oceanic and land surface processes (e.g., Meehl and Arblaster, 2002b), as well as regulation of these interactive processes by the climatological seasonal cycle.

12.7.2 Interdecadal changes

The correlation between Indian rainfall and the Southern Oscillation Index has experienced significant fluctuations in the past 120 years (Webster *et al.*, 1998). It was high during 1880–1920 and 1960–1980 (ranging from 0.5 to 0.6) but low in 1920–1940 (ranging from 0.2 to 0.3) and after the 1980s. The relationship between the east Asian monsoon and ENSO has also experienced significant changes in the mid-1970s (Chang *et al.*, 2000b; Wu and Wang, 2002). The precise causes for these secular changes are still unclear, but it may be related to changes in global-scale circulation patterns (Krishna Kumar *et al.*, 1999b), changes in the circulation over the North Atlantic (Chang *et al.*, 2001), or sampling fluctuations (Gershunov *et al.*, 2001).

ENSO properties experienced a significant change in the late 1970s (Wang, 1995), when the North Pacific experienced a notable climate shift (Trenberth and Hurrell, 1994). In model simulations of the global warming scenario, the tropical Pacific undergoes ENSO-like warming, similar to the decadal shift in the Pacific Ocean in the late 1970s (Meehl and Washington, 1996). Whether the secular change of the ENSO–monsoon relationship is related to the changes in the spatial–temporal structure of ENSO, or to global warming, remains uncertain. Further investigations are needed to identify the physical processes responsible for the interdecadal changes of ENSO–monsoon relationships.

12.7.3 Effects of atmosphere–ocean interaction

Monsoon–ocean interaction can considerably modify the monsoon–ENSO relationship in several ways. As noted in Section 12.3.2, the anomalous Walker circulation induced by eastern Pacific warming has a descending branch over the vicinity of the Maritime Continent. The suppressed deep convection over that region can generate westward propagating, descending atmospheric Rossby waves that are conducive to a weak Indian summer monsoon (Figure 12.6). However, the weakening monsoon can induce local warming in the Bay of Bengal (Figure 12.5, color section). As a result of this SST change, the Indian summer monsoon would tend to intensify (Figure 12.7, color section). Thus, through local atmosphere–ocean interaction, the ENSO-induced monsoon anomalies in turn offset the 'direct' ENSO impacts (Lau and Nath, 2000). Another example is the Indian Ocean dipole. During an eastern Pacific warming, the low-level anticyclonic pattern over the southern Indian Ocean enhances the cross-equatorial flow along the west coast of Sumatra (Figure 12.5), which subsequently enhances coastal and equatorial upwelling and decreases the SST off Sumatra. The resultant SST cooling in the equatorial eastern Indian Ocean, along with concomitant warming in the western Indian Ocean, forms a dipolar SST anomaly pattern (Saji *et al.*, 1999; see also Section 12.3.1). This dipole pattern tends to increase Indian rainfall and thus opposes the direct ENSO impacts on the Indian monsoon in year(0) (Figure 12.7). Over the WNP, the local air–sea interaction can maintain the ENSO-induced anticyclonic anomalies from the mature phase of El Niño to its decay phase, thus leading to a prolonged impact of ENSO on the east Asian summer monsoon (Section 12.5). On interannual timescales, Wang *et al.* (2003) have proposed that local monsoon–ocean interaction is one of the fundamental driving mechanisms for the biennial variability of the Asian–Australian monsoon. Questions remain regarding how the anomalies over the WNP and southern Indian Ocean interact to yield the biennial tendency of the continental-scale monsoon system.

12.7.4 Effects of atmosphere–land interactions

Land surface processes can significantly influence monsoon variability and monsoon–ENSO relationships. The snow cover and snow depth over the Tibetan Plateau and Eurasian interior might impact the strength of the Indian summer monsoon (e.g., Bamzai and Shukla, 1999). Increased snow pack increases surface albedo and reflects more solar energy from the land surface. Melting of excessive snow consumes additional heat energy. Thus, the land surface heats up slower than normal during the following spring and summer, thereby reducing land–sea thermal contrast and weakening the summer monsoon. While changes in snow pack over the Tibetan Plateau and Eurasia may be due to atmospheric internal dynamics, it may also be related to ENSO (Yang, 1996). Land surface processes could affect the ENSO–monsoon relationship by altering the energy and water cycles within the monsoon system (Chapter 11). The quantitative importance of such influences, however, has yet to be assessed.

12.7.5 Roles of intraseasonal oscillations

The Asian monsoon system exhibits prominent intraseasonal oscillations with time-scales of 30–60 days. The intraseasonal variations arise primarily from internal atmospheric moist dynamics, but could be considerably modified by air–sea interaction and land surface feedback. Anomalous behavior of these oscillations may have strong impacts on the seasonal mean monsoon climate. There exists considerable evidence for the covariability of intraseasonal activity and ENSO (Li and Long, 2002). In some regions (e.g., over the western Pacific), ENSO can influence the monsoon through regulating the Madden–Julian Oscillation (MJO) (e.g., Tam, 2003) and westward propagating biweekly oscillations (Chen and Weng, 1998). On the other hand, the intraseasonal oscillations in the Indian monsoon region could move into the western central Pacific Ocean and influence ENSO transitions through excitation of eastward propagating oceanic Kelvin waves, which can change thermal conditions in the eastern Pacific. Whether the intraseasonal oscillations of the monsoon system can trigger ENSO development or are more passive responses to ENSO remains a topic of debate.

The scientific issues raised in the above discussion can be best addressed by a combination of theoretical understanding of the basic processes involved, diagnosis of observational data, and experimentation with a hierarchy of numerical models. The history of progress in ENSO–monsoon research has shown that these alternative approaches are highly complementary to each other. For instance, insights gained from theoretical investigations have fostered fresh interpretations of empirical results and guided model experimental designs, whereas observational and model evidence has spurred theoretical advances. It is anticipated that improved analytical, observational, and modeling tools, in conjunction with coordinated intercomparison of findings derived from such diverse tools, will continue to enhance our understanding of ENSO–monsoon interaction.

12.8 ACKNOWLEDGEMENTS

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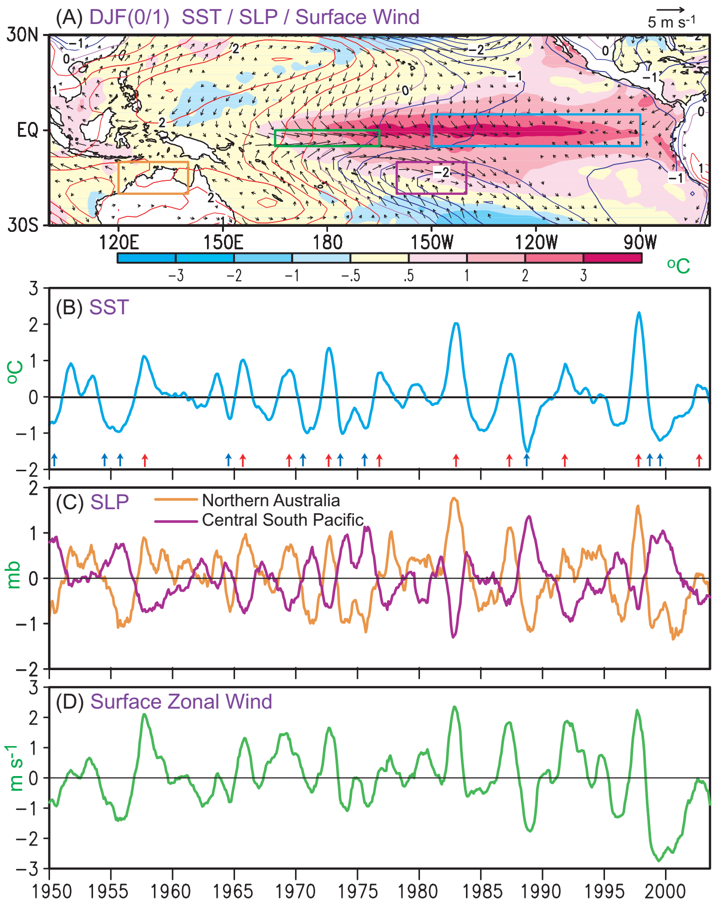

Figure 12.1. (a) Distribution of the warm-minus-cold composites of SST (shading), SLP (contour interval: 0.5 hPa), and surface wind vector (arrows) during DJF(0/1), as computed using observational data sets for ten warm and ten cold ENSO events (indicated by red and blue markers along the time axis in panel (b)). Time series of areal averages of anomalies of (b) SST in the central equatorial Pacific, (c) SLP in the vicinity of Tahiti and Darwin, and (d) surface zonal wind over the western equatorial Pacific. The rectangular sites used for computing the areal averages in (b–d) are indicated in (a), with colors matching those used in plotting the respective time series in the lower three panels.

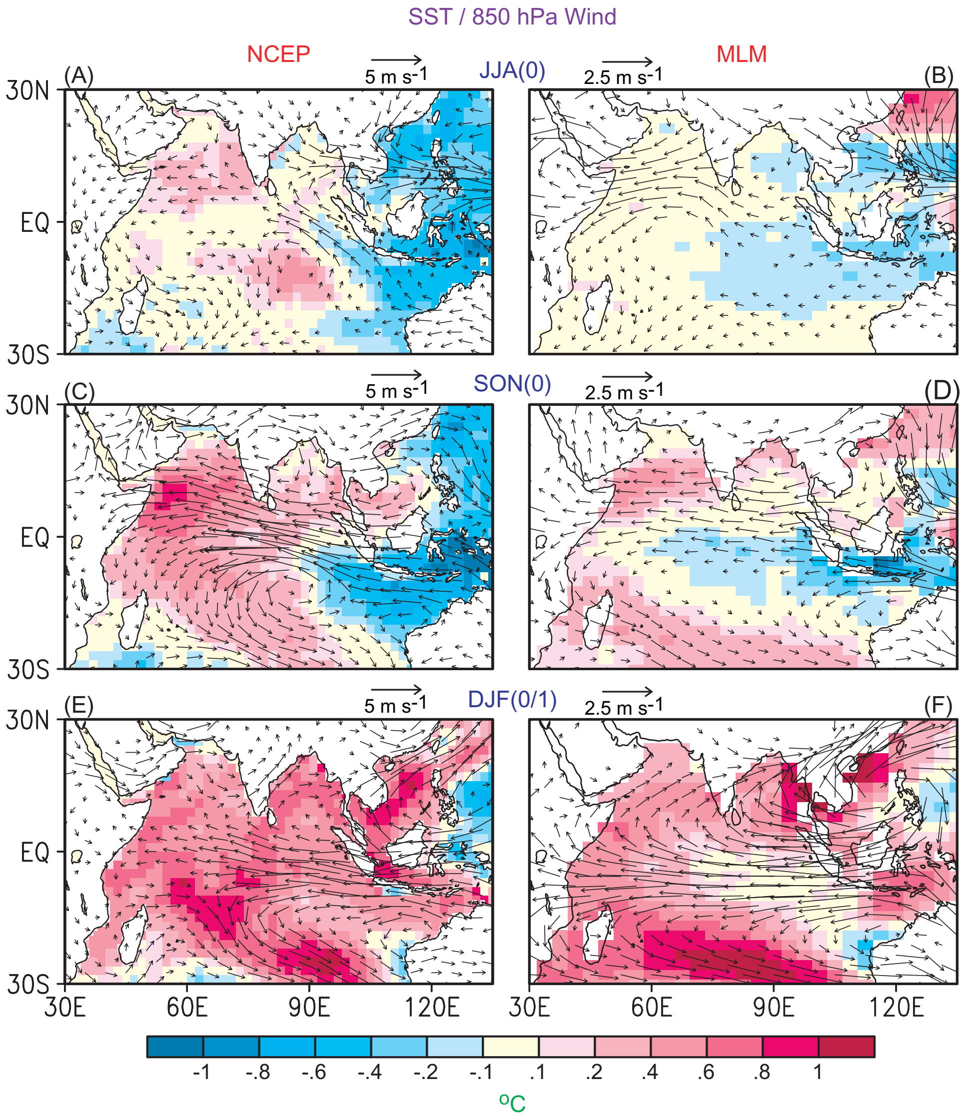

Figure 12.5. Distributions of the warm-minus-cold composites of 850-hPa vector wind (arrows; note different scales used in left and right panels) and SST (shading) fields, for (a, b) JJA(0), (c, d) SON(0), and (e, f) DJF(0/1). Results are based on NCEP data (left panels) and output from the MLM experiment (right panels) for six selected warm ENSO events and six cold events. The composite data shown in all following figures are based on the same set of ENSO events.

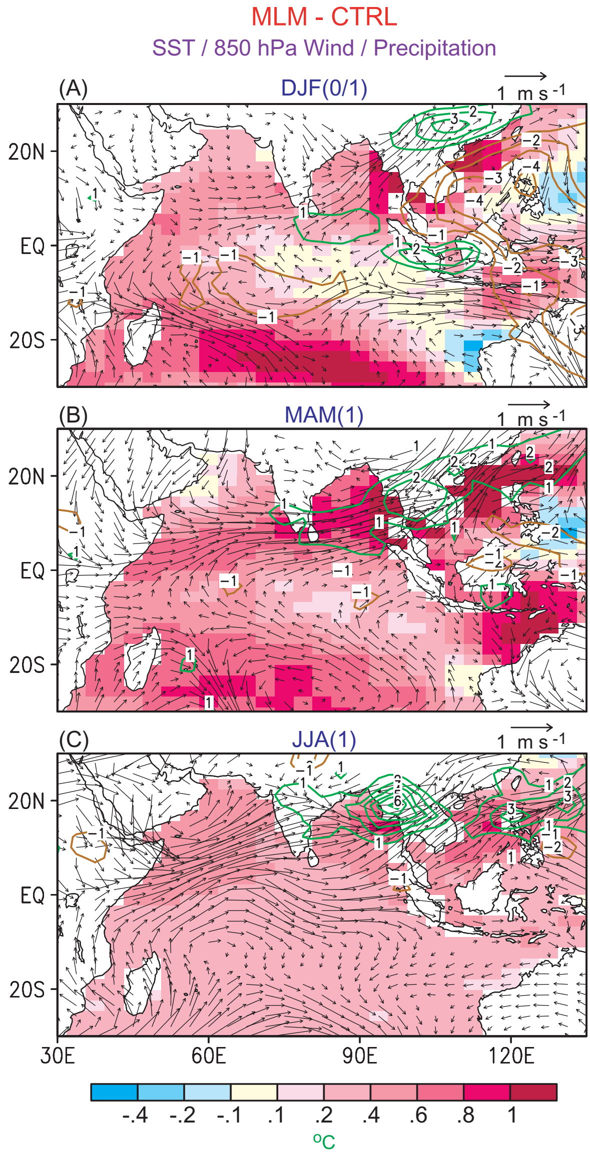

Figure 12.7. Distributions of the differences between the outputs from the MLM and CTRL experiments for the warm-minus-cold composites of SST (shading), 850-hPa vector wind (arrows), and precipitation (contour interval: 1 mm day^{-1}; zero contour omitted), for (a) DJF(0/1), (b) MAM(1), and (c) JJA(1).

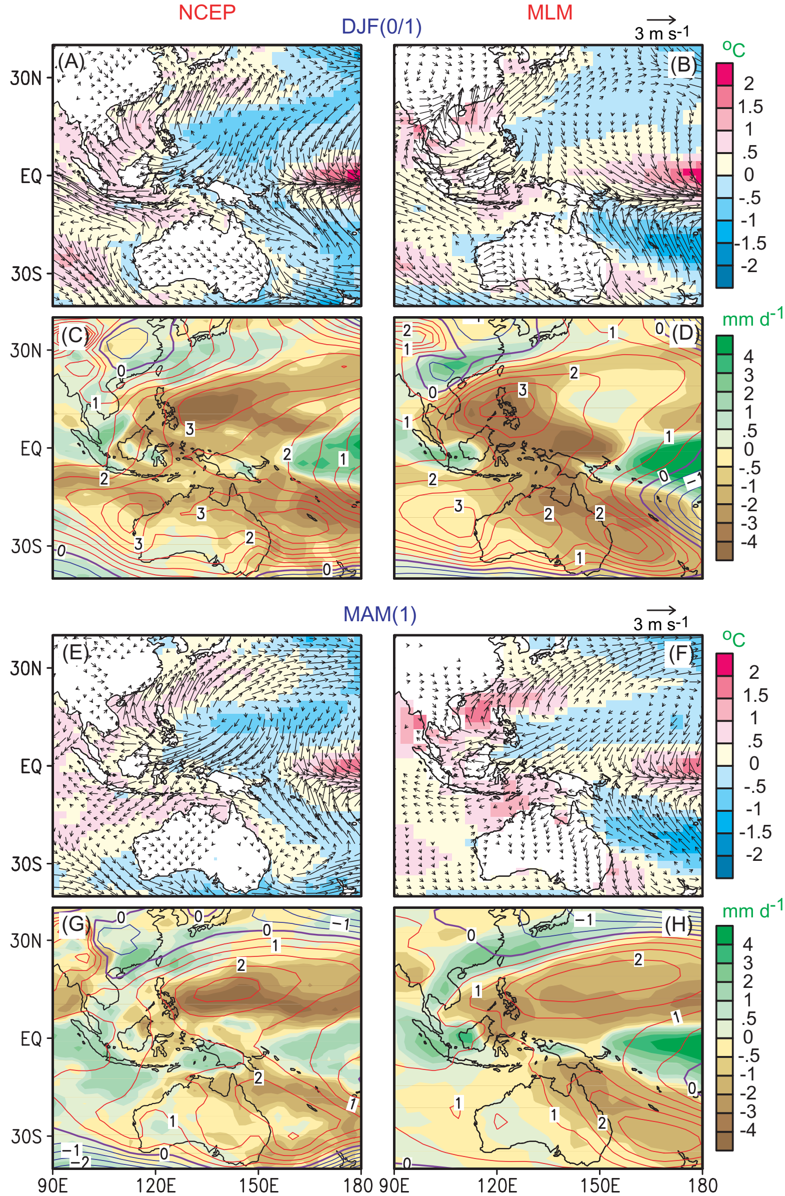

Figure 12.8. Distributions of the warm-minus-cold composites of (a, b, e, f) surface wind vector (arrows) and SST (shading), and (c, d, g, h) SLP (contour interval: 0.5 hPa) and precipitation (shading). Results are based on NCEP data (*left panels*) and output from the MLM experiment (*right panels*), for DJF(0/1) (*top four panels*) and MAM(1) (*bottom four panels*).

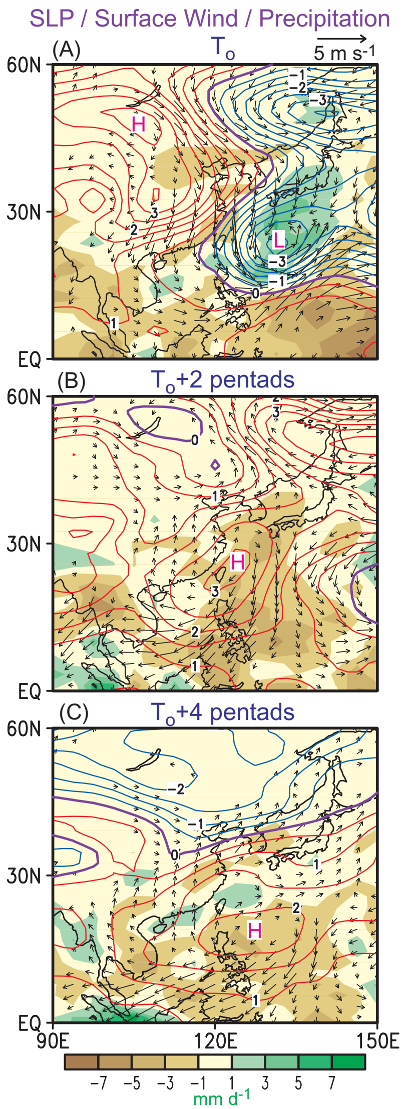

Figure 12.11. Distributions of the composite patterns of surface wind vector (arrows), SLP (contour interval: 0.5 hPa), and precipitation (shading), for the time periods of (a) T_o, (b) $T_o + 2$ pentads, and (c) $T_o + 4$ pentads, where T_o corresponds to the pentad when the SLP field over the South China and Philippine Seas makes a distinct transition to an anticyclonic pattern.