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Indian Monsoon Depression: Climatology and Variability

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1. Introduction

The monsoon climate is traditionally characterized by large amount of seasonal rainfall and reversal of wind direction (e.g., Krishnamurti 1979). Most importantly this rainfall is the major source of fresh water to various human activities such as agriculture. The Indian subcontinent resides at the core of the Southeast Asian summer monsoon system (Fig.1) with the monsoon trough extended from northern India across Indochina to the Western Tropical Pacific (WTP). Large fraction of annual rainfall occurs during the summer monsoon season, i.e., June – August[1], with two distinct maxima. One is located over the Bay of Bengal with rainfall extending northwestward into eastern and central India, and the other along the west coast of India where the lower level moist wind meets the Western Ghat Mountains (Saha and Bavardeckar 1976). The rest of the Indian subcontinent receives relatively less rainfall.

Various weather systems such as tropical cyclones and weak disturbances contribute to monsoon rainfall (Ramage 1971). Among these systems, the most efficient rain-producing system is known as the Indian monsoon depression[2] (hereafter MD). This MD is critical for monsoon rainfall because: (i) it occurs about six times during each summer monsoon season, (ii) it propagates deeply into the continent and produces large amounts of rainfall along its track, and (iii) about half of the monsoon rainfall is contributed to by the MDs (e.g., Krishnamurti 1979). Therefore, understanding various properties of the MD is a key towards comprehending the veracity of the Indian summer monsoon and especially its hydrological process.

However, it may be noted that earlier research on the formation and the water vapor budget of the MD may be constrained by limited observation over oceans adjacent to India, especially the Bay of Bengal. Because of this reason, many previous MD studies

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[1] It'll be denoted as JJA hereafter.
[2] Indian monsoon depression, monsoon depression, or MD are used exchangeably in this chapter.

(e.g., Nitta and Masuda 1981, Saha and Chang 1983, Saha and Saha 1988, and many others) mainly focused on a few cases during special field experiments (e.g., the Monsoon Experiment, MONEX[3]). It was not too long ago that a couple of more comprehensive studies on the water vapor budget, life cycle, and dynamical–hydrological cycle interaction were published with more modern datasets (e.g., Yoon and Chen 2005, Chen et al. 2000, Chen et al. 2005, and others).

The MD generally forms around Bay of Bengal and propagates westward or northwestward with the typical life span of three to six days (Ramage 1971, Krishnamurti 1979, Daggupaty and Sikka 1977, Krishnamurti et al. 1975, and 1976). Most of its rainfall occurs in the southwest quadrant of a MD with a recorded maximum of 100 – 200 mm over a 24-hour period (Daggupaty and Sikka 1977, Stano et al. 2002, Saha and Saha 1988). A previous study by Mooley (1973) estimated that MDs could contribute about 11 – 16% of total Indian summer monsoon rainfall using data from six stations (Calcutta, Allahabad, Delhi, Goplur, Nagpur, and Ahmadabad). However, more recent studies (e.g., Krishnamurti et al. 1979, Yoon and Chen 2005) found much higher contribution.

The Indian summer monsoon undergoes an active and break periods of rainfall. This change in rainfall is known as the 'life cycle' of the Indian summer monsoon. The typical life cycle consists of onset, break, revival, and withdrawal (e.g., Krishnamurti and Subrahmanyan 1982). This life cycle is regulated by two intraseasonal modes: the 30-60 and the 10-20 day monsoon modes. Following Yoon and Chen (2005), we call the extreme phases of each mode as active (break) phase for the 30-60 day mode and maximum (minimum) phase for the 10-20 day mode. During the active/maximum (break/minimum) phase of these two intraseasonal modes, the monsoon rain intensifies (weakens). Numerous studies have also reported that the monsoon trough located in northern India deepens and weakens as the 30-60 or 10-20 day mode varies (Krishnamurti and Subrahmanyan 1982, Chen and Yen 1986, Murakami 1976, Krishnamurti and Bhalme 1976). Furthermore, it was suggested that the MD is also affected by the two intraseasonal modes in its occurrence frequency (Chen and Weng 1999) and rainfall intensity (Yoon and Chen 2005), due to changes in the large-scale circulation and convergence of atmospheric water.

Our goal in this chapter is to review recent findings of the MD with focus on the following aspects:

1. Precipitation produced by the MD from genesis to demise and its contribution to monsoon rainfall over central India.
2. The atmospheric water budget of the MD regarding what maintains associated rainfall.
3. Coupling between the hydrological and dynamical processes of the MD.
4. Interaction between the MD and slowly-varying large-scale circulation change such as the Madden-Julian Oscillation (MJO; Madden and Julian 1993), the 10-20 day monsoon mode and the El Nino and Southern Oscillation (ENSO), along with its implication on long-lead climate prediction.

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[3] The Monsoon Experiment was conducted in 1978 – 1979, as an important part of the First GARP (Global Atmospheric Research Program) Global Experiment (FGGE) (e.g., Murakami 1979).

The chapter is structured along with these focal points mentioned above. In Section 2, the atmospheric water budget and data used are introduced. The climatology of the Indian summer monsoon and intraseasonal variability affecting the Indian summer monsoon are discussed in Section 3. Life cycle, water budget, and dynamical-hydrological processes of the MD are described in Section 4. Interaction between the low-frequency variability, such as tropical intraseasonal oscillation or interannual variability and the MD is demonstrated in Section 5. Concluding remarks are given at Section 6.

2. Atmospheric water vapor budget and data

2.1 Atmospheric water vapor budget

The hydrological cycle of the Indian summer monsoon and the MD can be analyzed with the following atmospheric water vapor budget equation:

$$\frac{\partial W}{\partial t} + \nabla \bullet \vec{Q} = E - P \, , \tag{1}$$

where W is atmospheric precipitable water defined as vertical integral of specific humidity ($W = \frac{1}{g}\int_{p_T}^{p_S} q \bullet dp$), $\vec{Q}$ is vertically-integrated water vapor flux ($\vec{Q} = \frac{1}{g}\int_{p_T}^{p_S} \vec{V}q \bullet dp$), E is evaporation, P is precipitation. g is the acceleration due to gravity, and p_S and p_T are surface pressure and pressure at the top of the atmosphere, respectively[4]. All the analysis is performed at 00Z and 12Z, when upper air sounding is launched and assimilated in atmospheric reanalysis product. Using these high-frequency data, daily- and seasonal-means are computed. Computational details in obtaining each term of the water vapor budget are summarized as follows:

- Convergence of vertically-integrated water vapor flux ($\nabla \bullet \vec{Q}$) is computed using atmospheric reanalysis at every 00Z and 12Z.

- Storage term ($\frac{\partial W}{\partial t}$) at 00Z and 12Z is computed by taking difference 6-hour before and after precipitable water from atmospheric reanalysis. For example, storage term at 12Z is computed taking different between precipitable at 18Z and 6Z, i.e., $\frac{\partial W}{\partial t}\Big|_{12Z} \approx \frac{\Delta W}{\Delta t}\Big|_{12Z} = \frac{W_{18Z} - W_{06Z}}{12h}$. A couple of different methods, Such as using daily mean values, were also tested. However, various different methods yield the same result (Yoon 1999).

- Precipitation (P) is only available as daily-mean value. However, outgoing longwave radiation is available twice a day (Liebmann and Smith 1996). In computing water budget, daily mean precipitation is used, while OLR is used to indicate strong convective regions in tracking the MD.

- Evaporation (E) is neither directly observed nor assimilated. Although it is computed by data assimilation system, evaporation is estimated as a residual in area-mean atmospheric water budget at a given domain (denoted as [ ], and defined as $[\ ] = \iint_A da$):

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[4] Surface pressure is taken as 1000mb, while the top of the atmosphere is 300mb in our computation. It is because most of atmospheric water vapor resides in the lower troposphere (Peixoto and Oort 1992).

$$[E] = [\frac{\partial W}{\partial t}] + [\nabla \bullet \vec{Q}] + [P]$$

(2)

This could include biases or analysis incremental terms of the data assimilation system.

The moisture flux can be separated in to rotational and divergent component ($\vec{Q} = \vec{Q}_R + \vec{Q}_D$), where $\vec{Q}_R$ is rotational and $\vec{Q}_D$ is divergent component of the moisture flux and only divergent component is necessary in the water budget analysis because rotational component does not contribute to the divergence of flux. The importance of divergent water vapor flux in maintaining tropical rainfall was discussed in Chen (1985). Thus, Eq.(1) can be rewritten as

$$\frac{\partial W}{\partial t} + \nabla \bullet \vec{Q}_D = E - P$$

(1a)

Further, stream function and potential function of the water vapor flux is computed following Chen (1985):

$$\vec{Q}_R = \hat{k} \times \nabla \psi_Q$$

(3)

$$\vec{Q}_D = \nabla \chi_Q$$

(4)

$$\nabla^2 \chi_Q = \nabla \bullet \vec{Q} = \nabla \bullet \vec{Q}_D$$

(5)

$$\nabla^2 \psi_Q = \hat{k} \bullet \nabla \times \vec{Q} = \hat{k} \bullet \nabla \times \vec{Q}_R$$

(6)

Eqs.(5) and (6) are computed in terms of the spectral method based on a horizontal resolution of T31, which is approximately 2.5° latitude × 2.5° longitude.

Considering the long-term mean water budget, one can neglect the storage term ($\frac{\partial W}{\partial t}$) because the atmosphere can hold water about 7 days. Eqs.(1) and (1a) can be rewritten as follows:

$$\nabla \bullet \vec{Q} \approx \nabla \bullet \vec{Q}_D \approx E - P$$

(7)

Further, the water vapor budget of the MD and the Indian summer monsoon is presented by area-averaged values of individual terms in Eqs. (1a) or (7). For the MD, we compute the area-averaged budget terms with a 10° × 10° box centered at the center of the MD, which is about 1,000 km x 1,000 km. Previously the maximum radius of influence affected by the MD is roughly 3,000 km (Godbole 1977, Sikka 1977, Krishnamurti et al. 1975, 1976). However, a relatively conservative estimate was used to analyze all 143 cases.

2.2 Data

Various precipitation estimates are used in this study. For the period of 1979 to 1997, rainfall proxies generated with satellite infrared (IR) observations at Goddard Space

Flight Center (hereafter GPI, Susskind et al. 1997) are used. This dataset is based on daily satellite IR observations with $1^\circ \times 1^\circ$ horizontal resolution. In addition, outgoing longwave radiation (OLR) observed by NOAA's polar orbiting satellites (obtained from the Climate Diagnostic Center[5], NOAA) is also used as an indicator of deep convection in the tropics for the early period when satellite retrieved rainfall data was unavailable. As for the later years (1998 – 2002), two different sources of rainfall estimates were used: Tropical Rainfall Measuring Mission (TRMM[6], Simpson et al. 1996)-3B42 and Global Precipitation Climatology Project (GPCP[7], Huffman et al. 1997). Both provide daily rainfall with $1^\circ \times 1^\circ$ spatial resolution.

It should be noted that one of the problems in satellite rainfall estimate (especially GPI in our case) is that heavy rainfall over land may not be properly detected. To overcome this deficiency, GPI was blended with station reported rainfall. Merging of two rainfall datasets was done by the methods described by Yoon (1999) and Yoon and Chen (2005). In summary, the Cressman scheme (Cressman 1959) was applied to produce station-reported rainfall at uniform $1^\circ \times 1^\circ$ grid first. Then, this was merged with GPI with weighting factors depending on distance to the land. For TRMM and GPCP products used in our study, station-reported rainfall was already merged. Other meteorological variables (e.g., wind field and specific humidity) were obtained from the ECMWF reanalysis data (Gibson et al. 1997) for the 1979 – 2002 period.

3. The Indian summer monsoon: Climatology and intraseasonal variability

3.1 Climatology

More than one billion people live in the south Asian monsoon regions. Previous studies have depicted the important features of the Indian summer monsoon (e.g., Fein and Stephens1987, and many others):

- The monsoon trough extends from northern India across Indochina to the western tropical Pacific (Fig.1a).
- South of the monsoon trough, strong westerly winds at the lower troposphere (extending from the Somali Jet) prevail with a maximum value of 20m/sec over the western Arabian Sea, and a significant eastward deceleration (i.e. low-level convergence) over the eastern Arabian Sea, the Bay of Bengal, and the South China Sea (Fig.1b).
- Heavy rainfall occurs at the western coast of India and the Bay of Bengal. Between these two major wet regions, a semi-arid region situates inside the Indian subcontinent (Fig.1b).

Our analysis domain (80°E – 95°E and 15°N – 27°N indicated as a box in Fig.1b) covers part of the Bay of Bengal and northeast India, which used for area-averaged atmospheric water budget (Fig.2b). This selected area is also quite close to that used by Goswami et al (1999) and more importantly covers the majority of MD tracks (e.g., Yoon and Chen 2005).

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[5] It is obtained from http://www.cdc.noaa.gov
[6] It is obtained from http://trmm.gsfc.nasa.gov
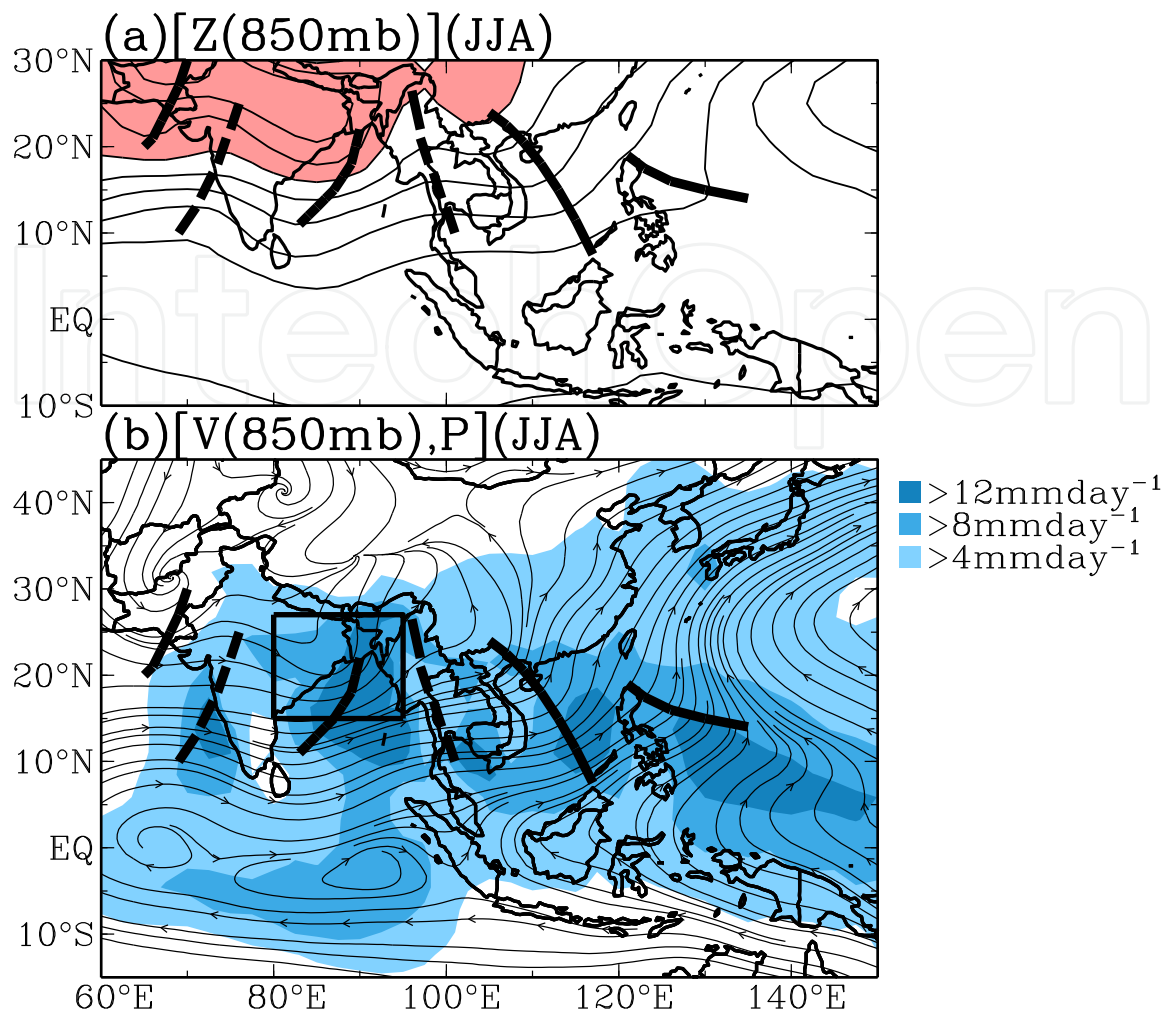[7] It is obtained from http://precip.gsfc.nasa.gov/

Fig. 1. Summer-mean circulation at the lower troposphere and rainfall. (a) geopotential height at 850mb, and (b) streamline of wind at 850mb and rainfall during northern summer monsoon season (JJA). Monsoon low (lower than 1460m) is colored in (a). Also, trough (ridge) are marked as thick solid (dashed) lines. Analysis domain (80°E – 95°E and 15°N – 27°N) is indicated as a box in (b).

The hydrological processes of the Indian summer monsoon are depicted in Fig.2a with potential function of atmospheric water vapor (χ_Q), divergent water vapor flux ($\vec{Q}_D$) and rainfall (P). Overall the Indian summer monsoon region sits within the vigorous convergent branch of the global hydrological cycle centered at the western tropical Pacific (Chen 1985). Fig. 2 clearly shows that the atmospheric water vapor fluxes converge toward the monsoon trough (located over northern Indian, Indochina, and the WTP; Fig.1). This converging atmospheric moisture maintains the monsoon rainfall centered over the Bay of Bengal, northeast Indian and the western coast of India (Fig. 2a). This feature is further substantiated by area-averaged atmospheric water budget (Fig. 2b) for our analysis domain. The area-averaged rainfall ($[P]$) in Fig.2b is about 12 mmday[-1] and the convergence of water vapor flux ($-\left[\nabla \bullet \vec{Q}\right]$) is approximately 7 mmday[-1], which accounts for about 60% of rainfall. The rest of rainfall is likely maintained by evaporation ($[E]$). This is reasonable because our analysis domain covers part of the Bay of Bengal where evaporation could be substantial during the northern summer season.
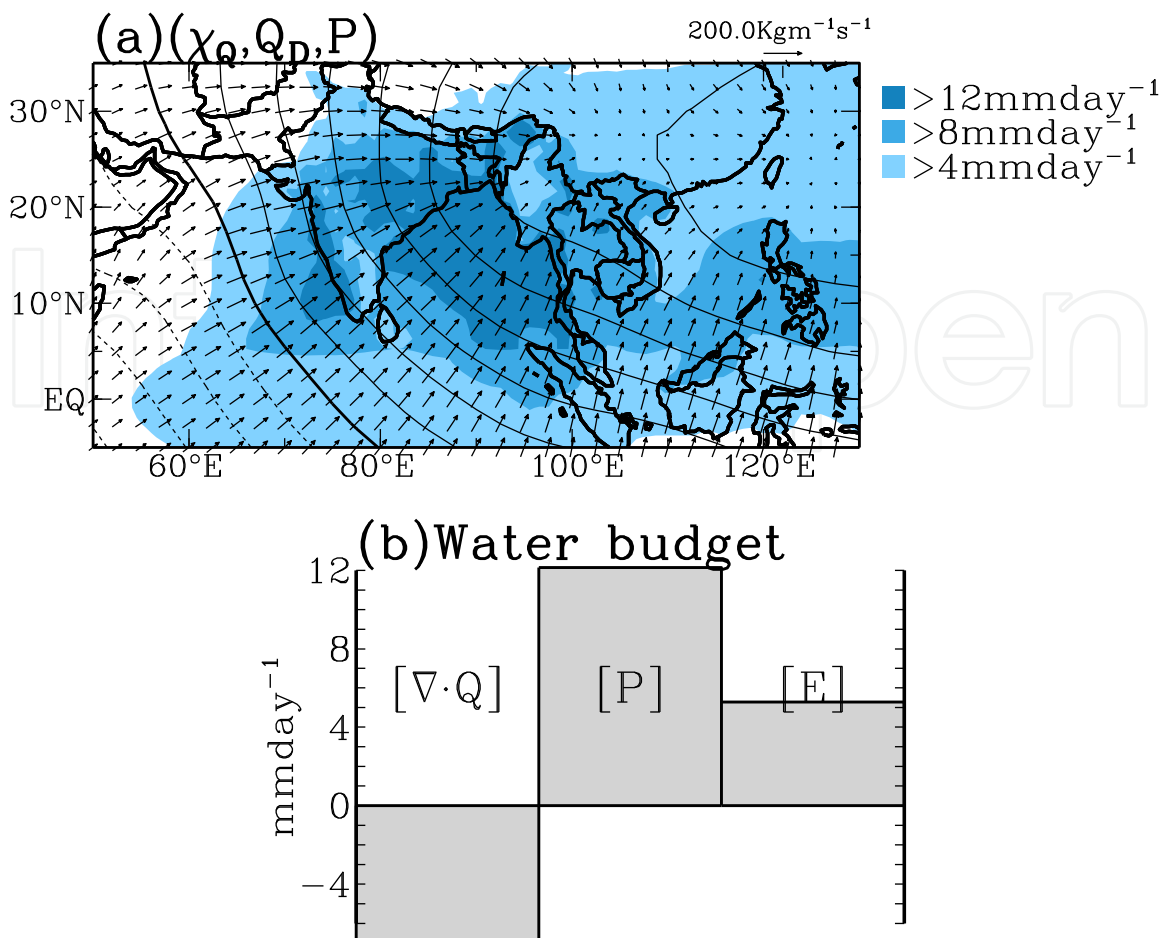
Fig. 2. (a) Summer-mean potential function of water vapor flux (χ_Q), the divergent water vapor flux ($\vec{Q}_D$), and precipitation (P). Contour interval is $2 \times 10^9 \mathrm{m^2 s^{-1} g \cdot kg^{-1}}$ for (a). The analysis domain used in (b) was marked as a thick black box in Fig.1(b).

3.2 Intraseasonal variability

The evolution of the Indian monsoon follows a periodical annual cycle but is also affected by quasi-periodic tropical intraseasonal oscillations. For example, the Indian summer monsoon undergoes a 30 – 60 day cycle with active and break phases which are linked to the northward migration of monsoon trough/ridge (Krishnamurti and Subrahmanayam 1982, Joseph and Sijikumar 2004, Krishnamurti and Shukla 2007, Pai et al. 2009) and to the global eastward propagation of the 30-60 day mode (Chen et al. 1988, Madden and Julian 1993). Monsoon trough intensifies (weakens) as monsoon westerlies, the convergence of water vapor flux, and monsoon rainfall intensifies (weakens) during an active (break) phase of the monsoon. To visualize the effect of the 30-60 day mode on the Indian summer monsoon, composite of the active and break phases of the 30-60 day monsoon mode is constructed. To properly isolate this intraseasonal monsoon mode, a band-pass filter designed by Murakami (1979) was used and a 30-60 day filter variable is denoted as (~). The 30 – 60 day band-pass filtered rainfall index over our computational domain (Fig.2a) and the zonal wind ($\tilde{U}(850mb)$) over the Arabian Sea (65°E, 15°N) were used as monsoon indices. Traditionally, the zonal wind index was used to represent the intra-seasonal monsoon life cycle (Krishnamurti and Subrahmanyam 1982). Based on these monsoon indices, all the days above 0.8 (below -0.8)

standard deviation of both indices were composited to describe the active (break) phase of the 30-60 day monsoon mode.

Figure 3 exhibits composite filtered anomaly of active/break phases of $\tilde{\chi}_Q, \tilde{\vec{Q}}_D, \tilde{P}$ (Figs.3a and b) and the area-averaged water vapor budget ($[\nabla \bullet \tilde{\vec{Q}}_D], [\tilde{E}], [\tilde{P}]$) (Figs.3c and d). Anomalous convergence (divergence) of atmospheric water vapor flux collocates with enhanced (suppressed) monsoon rainfall and trough, consistent with observations of Murakami et al. (1984) and Cadet and Greco (1987). The area-averaged water vapor budget shown in Figs.3c and d reveals that 25% (about 3 mmday^{-1}) increase (decrease) in both precipitation and convergence of water vapor flux during active (break) phase. However, unlike the climatology in Fig.2b the evaporation changes very little by the 30-60 day mode (Figs.3c and d). This may be due to (i) the fact that the 30-60 day mode changes relatively fast so that evaporation cannot respond to this fast-varying mode, or (ii) limitation of residual method used in our study. It is clear that the 30-60 day mode affects the atmospheric water budget through change of the atmospheric water vapor flux.
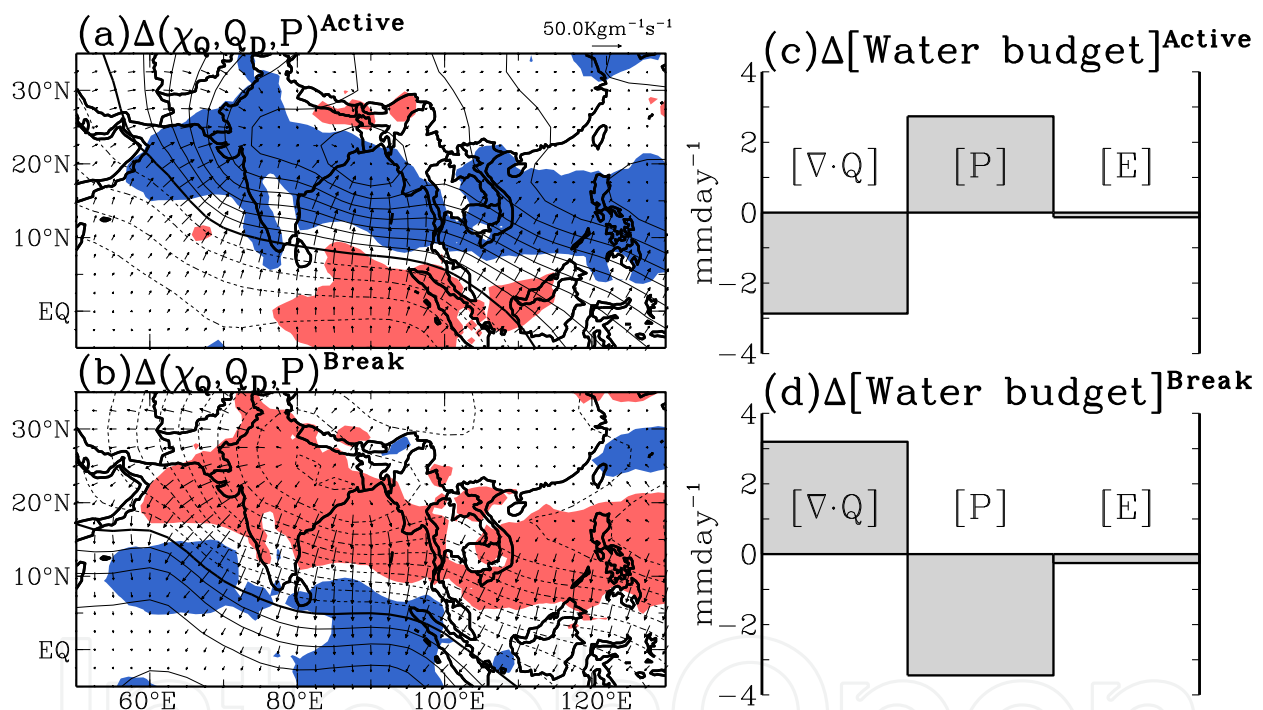

Fig. 3. The 30-60 filtered anomaly of water vapor flux and precipitation ($\tilde{\chi}_Q$, $\tilde{Q}_D$, $\tilde{P}$)active in (a), and ($\tilde{\chi}_Q$, $\tilde{Q}_D$, $\tilde{P}$)break in (b), where (~) is the 30-60 filtered anomaly. Area-averaged water budget departures from its summer-mean value during phases of (c) active and (d) break phase of the 30-60 day mode. The active and break phases of the 30-60 day monsoon mode are denoted by ( )active, and ( )break, respectively. Contour interval is 2×10^{8}m^{2}s^{-1}g·kg^{-1} for (a) and (b).

In addition to the 30-60 day mode, another important intraseasonal mode affecting the life cycle of the Indian summer monsoon is the 10-20 day mode. In fact, it has been found that the onset of the 1979 summer monsoon was caused by a phase lock of the 30-60 and the 10-20 day modes, not just the 30-60 day mode alone (Krishnamurti et al. 1984). Following similar methodology as used in Fig. 3 with a different frequency (10-20 day), the composite of the 10-20 day monsoon (denoted as (^)) mode was constructed. The synoptic structure of 10-20 day mode and its impact on the Indian summer monsoon were well documented in

previous studies (e.g., Murakami 1976, Krishnamirci and Ardanuy 1980, Chen and Chen 1993). Consistent with findings in these previous studies, a dipole structure of potential function of water vapor flux was observed in Fig.4a. In other words, converging center over the Indian subcontinent and diverging center at the equator were observed. It is obvious that the circulation center over the India affects the Indian summer monsoon more directly. On the other hand, the area-averaged water vapor budget indicates that about 17% change in rainfall and convergence of water vapor flux is contributed by 10-20 day mode (Figs.4c and d). This contribution is slightly smaller than that of the 30-60 day mode (Figs.3c and d), but not negligible. Also, change in evaporation is very small as the 30-60 day mode.
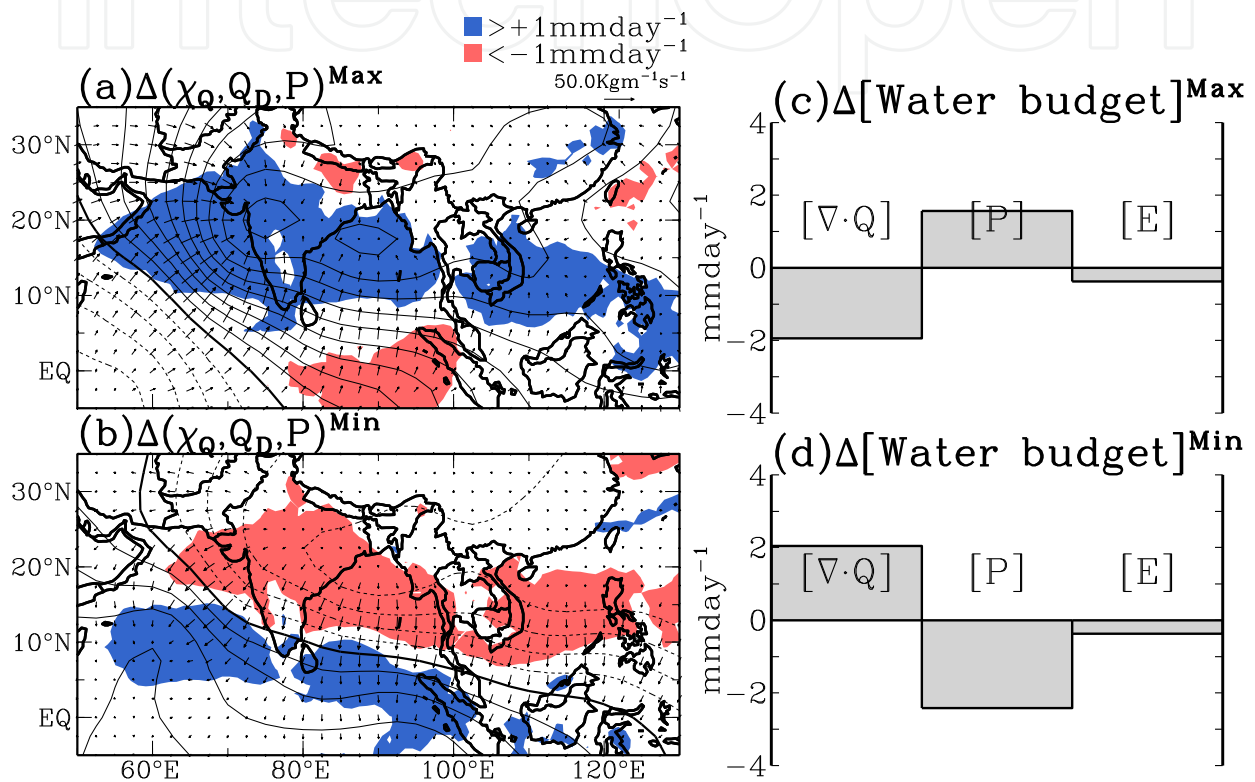

Fig. 4. Same as Figure 3, except for the 10-20 day monsoon mode. Maximum and minimum phases are used instead of active and break phases.

4. Indian monsoon depression

4.1 Life cycle of the MD

Detection of the MD was done with the same method as developed by Chen and Weng (1999) and Yoon and Chen (2005). Tracks of MDs were identified for 1979-1994 by Chen and Weng (1999) and were later expanded to 2002 by Yoon and Chen (2005). One hundred forty three (143) cases were identified over 24 summers (1979 – 2002); the dates were summarized in Fig.1 of Yoon and Chen (2005). Each MD was traced back to its origin position. It is clear from Fig.5 that most of the MDs are formed or intensified over the Bay of Bengal and have their predecessors as residual lows over the Western Tropical Pacific – South China Sea (WTP-SCS) region (Krishnamurti et al., 1977, Saha et al., 1981, Chen and Weng, 1999). There are six genesis mechanisms of the MDs according to Chen and Weng (1999) and Saha et al. (1981). Initial location of the MD with genesis mechanism is summarized in Fig.5.

Population of Monsoon Depressions (1979−2002) and related weather disturbances

(a) First appearance of
• Monsoon Depressions
6 Tropical Cyclones
+ Weak disturbances generaged by TCs
□ Weak disturbances by land−genesis
∗ 10−20day monsoon low
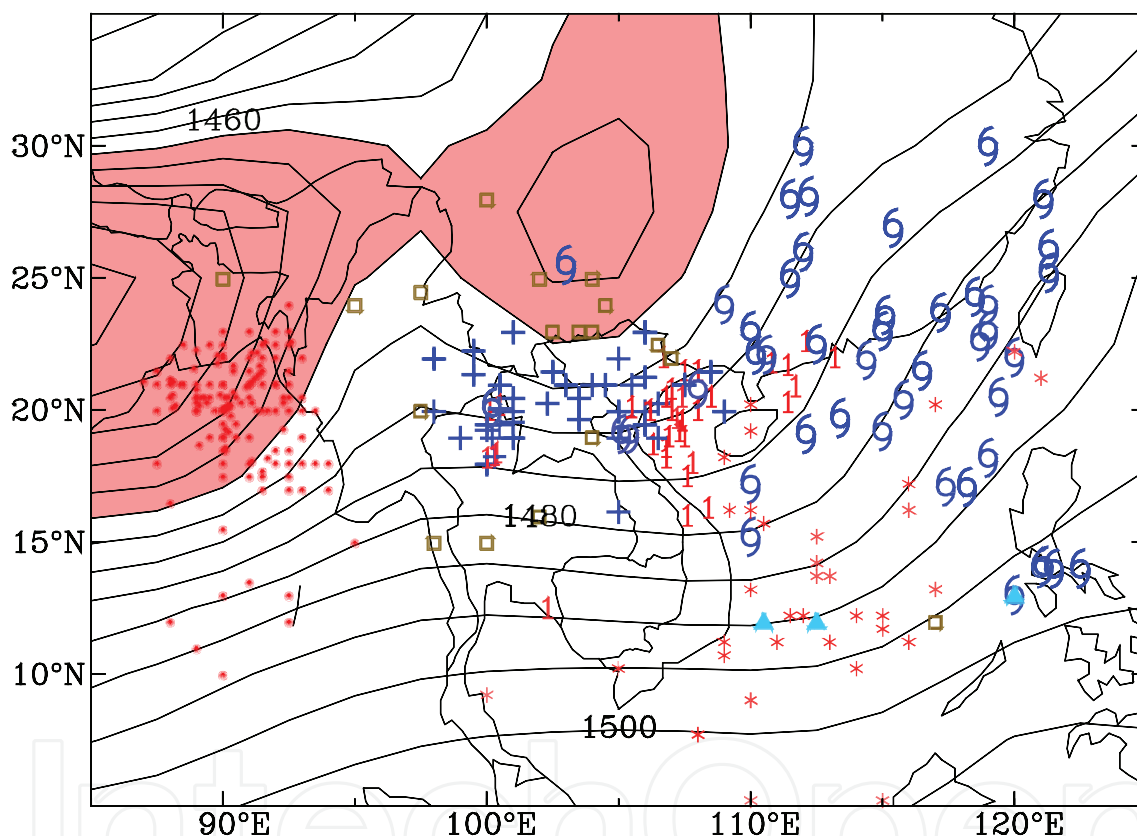▲ easterly waves

(b) Last appearance of
1 Tropical Cyclones

Fig. 5. Genesis locations and mechanisms of the MDs consistent with Fig.5 of Chen and Weng (1999) but for longer record of the MD for 1979 – 2002. First appearance of MDs (red dot), that of weak disturbances apparently linked to tropical cyclones (open tropical cyclone symbol), genesis over land, especially Indochina (brown open square), last appearance of tropical cyclone (marked as 1), 10-20 day monsoon lows (pink diamond), equatorial waves in the WTP-SCS (marked as triangle) are shown on top of the geopotential height at 850mb.

After its formation in the Bay of Bengal, a MD migrates westward or northwestward into the Indian subcontinent with a phase speed of roughly 5° longitude•day-1 (e.g., Godbole 1977; Sanders 1984). Because the distance between Bangladesh (~95°E) and northwest India (~70°E) is 25° in longitude, it is estimated that five days are needed for a MD to migrate

across the Indian subcontinent. This is consistent with the typical life span of the MD ~ 5 days (e.g., Krishnamurti et al. 1977, Saha et al. 1981, Nitta and Masuda 1981, Chen and Yoon 2000a, and many others). One can estimate that a MD initiated locally or transformed from a residual low over the WTP-SCS gets fully developed over the Bay of Bengal and moves into the Indian subcontinent in 5 days.

To quantitatively illustrate this time evolution of a MD, a composite scheme was developed based on its centered location following Yoon (1999) and Yoon and Chen (2005). First, the average location of all MDs within a 5° – longitudinal zone is obtained and centers of all MDs within this longitudinal zone are adjusted to match this averaged center. For example, all MDs with centers located between 95˚E to 90˚E, over the Bay of Bengal are averaged to form the MD at Day 1. We repeat this for 2 days prior to and 5 days post its formation over the Bay of Bengal. During the former period, the MD is classified as the prior depression phase (equivalent to the residual low). For the later period, the depression is generally with its rainfall larger than 25 mmday^{-1}, identified as Phase 2 (Day 2 – 4). Note here that some MDs with either slower or faster phase speed than 5° longitude·day^{-1} may exist, which implies our composite method may not ideally fit to these cases. However, phases of these MDs are still decided by their longitudinal location only in our study.

The evolution of MDs prior to and post of phase 2 is classified as phase 1 (Day 1) and Phase 3 (Day 5) when the system exhibits development over the Bay of Bengal and decaying at the western and northwestern part of India following Yoon (1999) and Yoon and Chen (2005). Also, composite of prior depression phase was constructed for two days. All different phases with corresponding location and composite days are listed in Table 1.

Phase	Location	Composite Days
Prior depression	Indochina	Days -2 and -1
Phase 1 or Developing	Bay of Bengal	Day 1
Phase 2 or Mature	Inside the Indian subcontinent	Days 2, 3, and 4
Phase 3 or Decay	Western India	Day 5

Table 1. List of phases, location and days in our composite analysis. More details can be found in Yoon (1999), Yoon and Chen (2005), and Chen et al. (2005).

The life cycle of a MD depicted with wind and divergent circulation in the lower troposphere along with different phases are depicted in Fig.6. For better isolation of this synoptic-scale disturbance from the background flow, zonal-wave filter using Fast Fourier Transform (FFT) is applied. The maximum horizontal scale of the MD is reported to be about 3,000 km (Godbole 1977; Krishnamrti et al. 1975) which is corresponding to zonal waves numbers 12 – 13 at 20°N. Considering variable sizes of individual MD, we applied a Fourier spectral filter with zonal wave number of 6 – 25 (denoted as [ ]^s) following previous studies (Yoon 1999, Chen and Yoon 2000a, Yoon and Chen 2005, Chen et al. 2005).
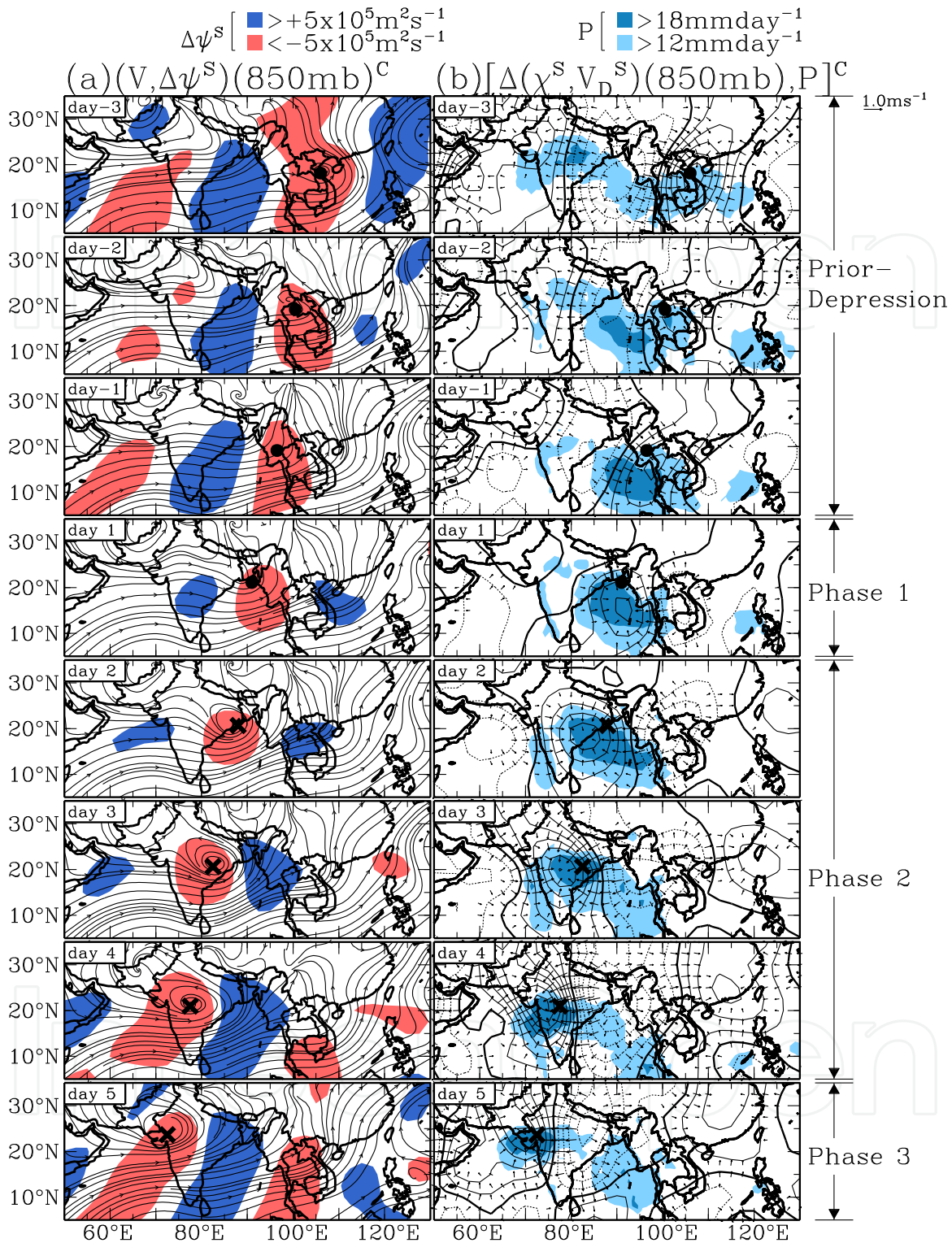
Fig. 6. A composite of MD during twenty four summers (1979−2002) for 7 days (5 (2) days after (before) formation of the MD, listed in Table 1) using (a) wind at 850mb superimposed with departures of the short-wave stream function at 850mb $[\Delta(V,\psi^S)(850mb),P]$, and (b) composite 850-mb velocity potential and divergent wind departures in the short-wave regime superimposed with precipitation, $[\Delta(\chi^S,\mathbf{V}_D^S)(850mb),P]$. Contour interval is $2\times10^8m^2s^{-1}g\cdot kg^{-1}$. $\Delta()$ is departure from the summer-mean value of ().

Here we summarize several important dynamical and physical characteristics based on the composite MD as revealed in Fig.6 and from previous studies (e.g., Krishnamurti et al. 1975, Krishnamurti et al. 1977, Saha et al. 1981, Saha and Saha 1988, Sikka 1977, Daggupaty and Sikka 1977, Godbole 1977, Chen and Yoon 2000a, Douglas 1992a and b, Rajamani and Sikdar 1989, and others):

- The horizontal scale of a MD is in the range of 1,500 to 3,000 km with central pressure down to 990mb. Its propagation speed is 3 – 5 degree per day. Of note, it propagates westward against the strong monsoon westerlies in the lower atmosphere (Fig.6a).

- The vertical extent of a MD is only up to 300mb restrained by the existence of the Tibetan high during Indian summer monsoon season. It implies that the upper level monsoon easterlies may not steer a MD westward (Chen and Yoon 2000a).

- The maximum rainfall and rising motion of a MD were detected in the southwest quadrant of the system (Fig.6b). This is consistent with early observation during MONEX periods (Saha and Saha 1984). Maximum rainfall can be as high as 100mm per day.

- Because the MD does not stay long enough over the Bay of Bengal, energy input from the surface boundary appears unimportant.

- Its evolution and dynamical characteristics have various stages. At its early stage, the barotropic dynamics may play an important role. On the other hand, baroclinic and Conditional Instability of the Second Kind (CISK) dynamics are more important at later stage.

- It is a tropical weather disturbance so that planetary vorticity (f) cannot be larger than relative vorticity (ζ). In other words, one cannot ignore relative vorticity advection (Chen and Yoon 2000a).

4.2 Water vapor budget of the MD

Fig.6b shows that the MD brings a large amount rainfall into central India as it moves westward. To acquire a quantitative measure of how much rain is produced by a MD, we perform area-averaged water vapor budget with a moving window following the MD's position. The research questions addressed here are: *(1) what is the contribution of the MD to the total monsoon rainfall?* and *(2) how this rainfall is maintained within the depression?* Understanding of this hydrological process helps explain the dynamical evolution of a MD, as explained later in Section 4.3. To answer the above questions, composite maps of water vapor fluxes and rainfall are shown with area-average water vapor budget in Fig.7. Important findings obtained from Figs.6 and 7 are summarized here:

1. *Prior depression phase (Days -2 – -1)*: A weak low-pressure system moves across Indochina as shown by 850-mb streamline with the short-wave filtered stream function anomalies (colored) in Fig.6a. Although it is a relatively weak system with less-organized circulation structure, a residual low brings some amount of precipitation (Chen and Yoon 2000b) and is maintained by low-level convergence (Fig.6b). As indicated by previous studies (Chen and Weng 1999, Saha et al. 1981), most of the MDs have their predecessors over Indochina. Therefore, it is important to identify or track these systems.

2. *Phase I (over the Bay of Bengal at Day 1)*: After reaching the Bay of Bengal, this weak low pressure system transforms into a MD. Our composite captures its transition stage, which signifies the MD's development. Apparently, the cyclonic vortex intensifies (Fig.6a) in accordance with stronger convergence of low-level flow (Fig.6b) as well as water vapor flux and rainfall (Fig.7).
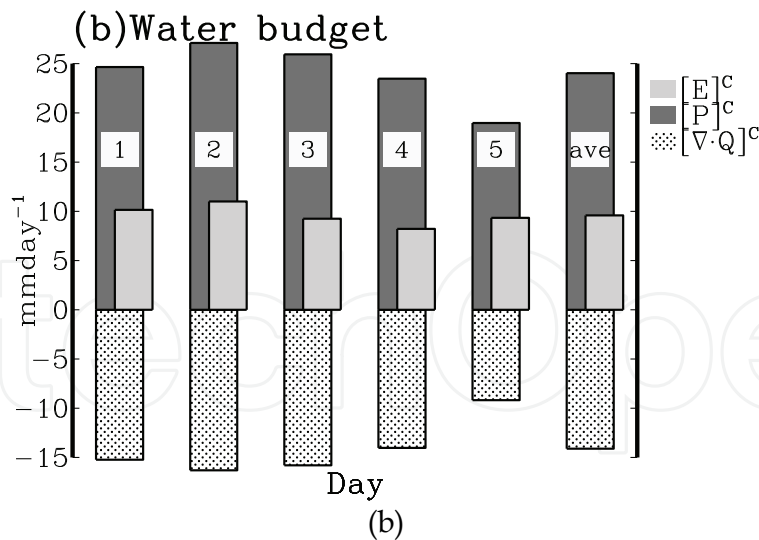
(a)
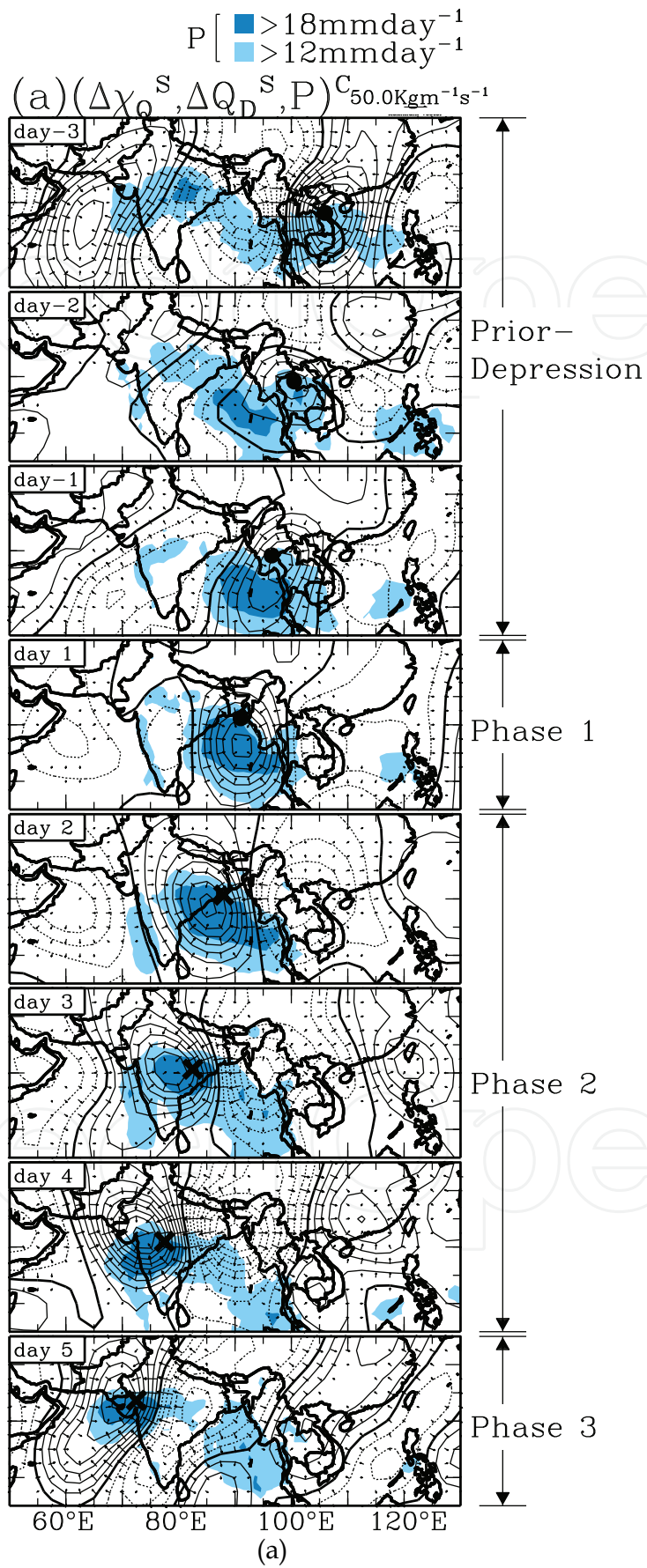
(b)Water budget

Fig. 7. (a) same as Fig.6 except potential function of water vapor, divergent water vapor flux, and rainfall [$\Delta(\chi_Q^S, \bar{Q}_D^S), P$] and (b) area averaged water vapor budget for Day 1 – 5 and average of these five days.

3. *Phase 2 (Days 2 – 4):* At this phase, a MD starts to progress into the Indian subcontinent and produces large amounts of rainfall inland. An important finding here is that as rainfall increases, the convergence of atmospheric water vapor flux intensifies as well (Fig.7a). This implies that the major source of MD rainfall is not from surface source through evaporation but through atmospheric hydrological process. This is further substantiated in terms of area-averaged water vapor flux shown in Fig.7b. At the last day of this phase (Day 4), the entire system starts to weaken.

4. *Phase 3 (Decay at Day 5):* This is the last stage of a MD's lifecycle. However, the low-pressure system can be easily identified and it continues to produce rainfall inside the continent. The demise of the system is clearly seen in the amount of the water vapor flux converging into the system (Fig.7).

It is noted here that during the pre-depression stage, rainfall is less organized but located mainly at the east side of the system. However, as it develops into a MD, major rainfall is concentrated on the western or southwestern corner. Corresponding lower-level divergent circulation is formed across a MD with a convergent (divergent) center located east (west) of the system. Based on area-average water vapor budget, rainfall and convergence of water vapor flux reach their maximum at Day 2 and maintain their strength till Day 4. On the last day (Day 5), convergence of water vapor flux is reduced to about half compared to Day 2.

For most tropical weather systems, water vapor is mainly supplied by evaporation from warm ocean surface (Riehl 1954). However, in the case of the MD, atmospheric water vapor transport and convergence is by far the most important source. In this sense, hydrological processes of the MD are close to those of the mid-latitude cyclones (Chen et al. 1996).

Typically, about six MDs (e.g., Chen and Weng 1999) develop every monsoon season over the Bay of Bengal. It is shown by the composite analysis that a MD could stay about 3 – 4 days (from Day 1 to Day 4) over the Indian subcontinent with rainfall over 25 mmday^{-1}. A simple estimation of the total rainfall by MDs (3 – 4 days × 25mmday^{-1} × 6/season = 450-600 mm/season) is equivalent to about 45 – 55% of the total monsoon rainfall (92

days/season × 12mmday^{-1} = 1104 mm/season) over the computational domain (75°E – 90°E, 15°N – 27°N). This value is higher than that estimated by Mooley (1973), probably due to the different methods and data employed, but is close to the result shown by Dhar et al. (1981). Mooley (1973) used only six stations along the eastern coast of India to estimate the MD contribution to the total rainfall over these stations, while the current study applied composite analysis using rainfall of GPCP, TRMM, and GPI to estimate the contribution by the MD to the total monsoon rainfall. Based on our estimate, about half of the total rainfall over the eastern coast of India is generated by the MDs.

4.3 Coupling between hydrological and dynamical processes of the MD

Our next question is how the aforementioned hydrological processes are linked to dynamical properties of the MD. To answer this question, we'll use the stream function tendency equation, which is another form of the vorticity equation (e.g., Holton 1992). Let's briefly review the vorticity (Eq.(8)) and the stream function tendency equations (Eq.(9)):

$$\frac{\partial \zeta}{\partial t} = -\mathbf{V} \cdot \nabla(\zeta + f) - (\zeta + f)\nabla \cdot \mathbf{V}$$

$$\zeta_t \qquad\qquad \zeta_A \qquad\qquad\qquad \zeta_\chi$$

$$(8)$$

First, vorticity tendency (ζ_t) represents whether vorticity at one location becomes more positive (cyclonic circulation) or negative (anti-cyclonic circulation). Second this vorticity tendency is determined by sum of advection (ζ_A) and stretching terms (ζ_χ). Analyzing each term reveals which process is important in a particular system. For example, vorticity advection term by westerly jet in the upper troposphere becomes a dominant process in eastward moving mid-latitude storms (Holton 1992). Vorticity in the tropics is at least one order smaller than that in the mid-latitudes. Thus, it is difficult to apply the vorticity tendency equation to the MD.

To overcome this difficulty, the stream function tendency equation is constructed by applying Inverse-Laplacian to the vorticity equation (Eq.8):

$$\nabla^{-2}(\frac{\partial \zeta}{\partial t}) = \nabla^{-2}[-\mathbf{V} \cdot \nabla(\zeta + f)] + \nabla^{-2}[-(\zeta + f)\nabla \cdot \mathbf{V}]$$

$$\psi_t \qquad\qquad \psi_A \qquad\qquad\qquad \psi_\chi$$

$$(9)$$

This stream function tendency equation (Eq.9) was applied for a case during FGGE-MONEX by Sanders (1984) and Chen and Yoon (2000) to illustrate dynamical properties of the MD. The dynamical implication of each term in Eq.(9) is the same as that of Eq.(8). Name of each term in both equations are summarized in Table 2. Consistent with our analysis earlier, a short-wave filter in zonal direction was applied.

First, let us examine the existence of an east-west asymmetric circulation across a MD proposed by Saha and Saha (1988) based on heat budget analysis. As shown in Fig.6, lower tropospheric rotational and divergent flow has a spatial quadrature relationship. In other words, local maxima or minima of potential function at 850mb ($\Delta\chi^S(850mb)^C$) located at nodal points of the stream function $\Delta\psi^S(850mb)^C$, in between positive and negative centers. On top of this lower tropospheric convergence, upward motion is located in the west and

southwest corner of the MD (Fig.4 of Chen et al. 2005). Although its downward branch is less clear, it is conceivable that an east-west divergent circulation exists across the MD.

Vorticity Equation	Stream function tendency equation			
$\dfrac{\partial \zeta}{\partial t}$	$\nabla^{-2}[\dfrac{\partial \zeta}{\partial t}]$	ζ_t	ψ_t	Vorticity (stream function tendency)
$-\mathbf{V}\cdot\nabla(\zeta + f)$	$\nabla^{-2}[-\mathbf{V}\cdot\nabla(\zeta + f)]$	ζ_A	ψ_A	Advection term
$-(\zeta + f)\nabla\cdot\mathbf{V}$	$\nabla^{-2}[-(\zeta + f)\nabla\cdot\mathbf{V}]$	ζ_χ	ψ_χ	Stretching term

Table 2. Vorticity and stream function tendency terms.

Second, each term of Eq.(9) was computed at all the phases of the MD. The result indicates that stretching term is much larger than advection (Chen et al. 2005). Thus, one can approximate this equation as follows:

$$\psi_t^S \sim \psi_\chi^S \tag{10}$$

Negative stream function tendency is found west and southwest corner of the depression center, which makes the system, i.e., negative stream function center, moves westward (not shown). Convergence at the lower troposphere can generate negative stream function tendency by vortex stretching. Further, total vorticity ($\eta = \zeta + f$) in vortex stretching term can be assumed as a constant and the budget equation can be simplified as follows:

$$\psi_t^S \sim -\eta\chi^S \sim -(\zeta + f)\chi^S . \tag{11}$$

Our results confirm findings by previous studies (Sanders 1984, Chen and Yoon 2000a, Chen et al. 2005) that advection of total vorticity is negligible in the westward propagation of MDs. Indeed the stream function tendency generated by vortex stretching over the west-southwest sector of a MD is vital to its westward propagation.

Finally, collocation of strong convective rainfall, convergence (divergence) at the lower (upper) troposphere, and upward motion implies that the divergent circulation is closely linked to hydrological processes. In summary, (1) release of latent heat at the center of convection, which is west side of the MD center, produces strong convergence (divergence) at the lower (upper) atmosphere and its counter part to the east of the MD center, and (2) this strong upward branch exerts negative stream function tendency through vortex stretching term (Fig.6 of Chen et al. 2005). This is well depicted by a schematic diagram in Fig.8. Based on the composite $\Delta\psi^S(850mb)$ budget of MDs, it is clear that $\Delta\psi_A^S(850mb)$ (stream function tendency generated by total vorticity advection) is not an effective dynamic process in generating $\Delta\psi_t^S(850mb)$. Instead, $\Delta\psi_\chi^S(850mb)$ coupled with the east-west asymmetric circulation of the MD is the primary dynamic process in propagating a MD westward. The convergent center of the east-west circulation west of a depression center overlaps the negative stream function tendency. Therefore, the depression is propagated westward by this negative tendency.
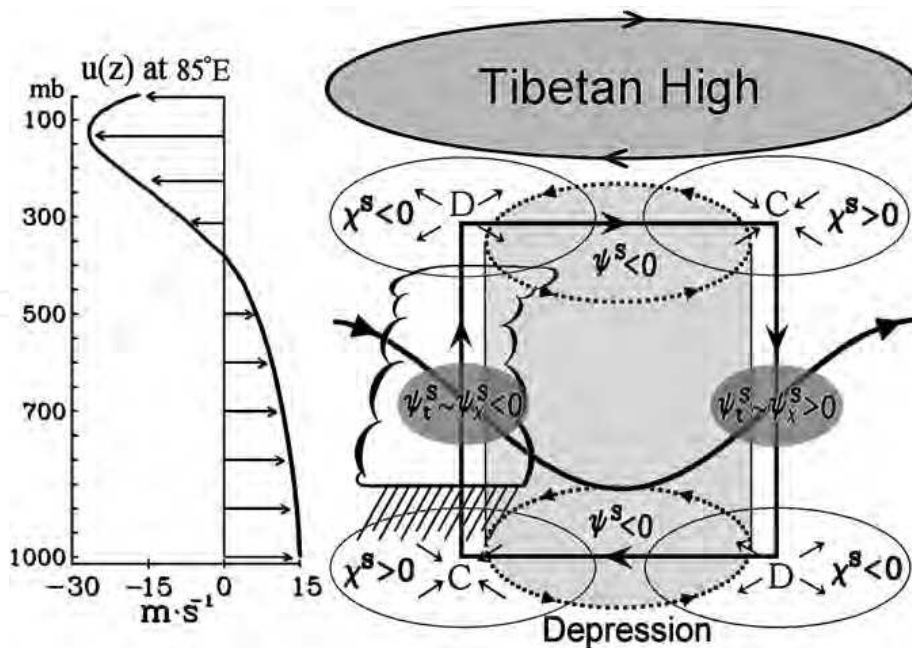
Fig. 8. A schematic diagram of the MD (a stippled cylinder, identified as $\psi^S < 0$) embedded in the monsoon westerlies in the lower troposphere (indicated by the zonal wind profile left) and the Tibetan high in the upper troposphere (the elliptic stippled area at the top of the right diagram). The combination of the depression and monsoon westerlies is portrayed by the thick sinusoidal streamline. The east-west asymmetric circulation of the MD is portrayed by solid lines with shafts encircling a cylindrical-shaped depression. The divergent circulation (χ^S, and divergent wind) coupled with the east-west circulation is denoted by thin-line circles centered at Ds (divergent centers) and Cs (convergent centers) at the upper and lower levels. Cumulus convection/rainfall west of the depression center is coincident with the upward branch of the east-west circulation. This depression is moved westward by the negative (positive) stream function tendency (ψ_t^S) generated by the vortex stretching (ψ_χ^S) associated with the upward (downward) branch of the east-west circulation. This was recreated from Fig.1 of Chen et al. (2005) with permission.

5. Interaction between the low-frequency variability and the MD

5.1 Modulation of the MD by the intraseasonal variability

It is shown in Section 3 that the Indian summer monsoon system is modulated by two tropical intraseasonal modes: 30-60 and 10-20 day monsoon modes. Because the MD is the major rainfall contributor to the Indian summer monsoon rainfall, it is conceivable that both the 30-60 day and 10-20 day modes also have a large impact on the MDs and that this impact is achieved by changing the atmospheric water flux.

A previous study by Chen et al. (1999) found that more low-pressure systems propagate into the Bay of Bengal during the active phase of the 30-60 day monsoon mode than the break phase due to changing monsoon westerlies. In this study, we rather focus on changes in the atmospheric hydrological processes. To show the modulation of the MD by the 30-60 day mode, we group all the MD cases into two categories: One with those occurring during the active phase of the 30-60 day mode, and the other during the break phase. The phase of the 30-60 day mode was determined by two monsoon indices mentioned in Section 3.

A MD during active phase of the 30-60 day mode is shown in Fig.9. At its first day over the Bay of Bengal on August 6th 1979, a MD exhibits a clear intensification, moves inland on August 8th, and reaches the other side of the Indian subcontinent on August 10th. Noteworthy here is that even at its last day of propagation, this system still maintains strong rainfall and a well-organized cyclonic flow. Also, another low-pressure system appears over Indochina. In other words, more low-pressure systems can propagate during active phase of the 30-60 day monsoon mode (Chen et al. 1999). In contrast, a case during the break phase is overall weak and less organized (not shown here but refer to Yoon 1999 or Yoon and Chen 2005).

It is found that change in the large-scale circulation, especially the converging atmospheric water vapor flux is responsible for modulation of a MD by the 30-60 day monsoon mode (Fig.9b). The difference in area-averaged atmospheric water budget of the MD during active and break phase of the 30-60 day mode is shown in Fig.10a to further elucidate this process. During the active phase of the 30-60 day monsoon mode, rainfall is about 7mm/day higher than that during the break phase. Relatively large values of atmospheric storage term ($\frac{\partial W}{\partial t}$) and evaporation ($E$) were also observed. This has a physical implication that the surface process is important in tropical intraseasonal variability (Shinoda et al. 1998). However, this large contribution of evaporation in our budget analysis could be due to (1) biases in atmospheric reanalysis used in our study or (2) a sampling bias that numbers of MDs during active are more than break phase so that uncertainty becomes larger in evaporation or storage terms. The former can be fixed using more modern atmospheric reanalysis such as MERRA[8], CFSR[9], or ERA-interim[10], and the latter needs to be tested using more longer analysis period.

The same procedure was repeated for the MD and the 10-20 day monsoon mode. A MD case during maximum phase (June 22nd 1979 – June 26th 1979) is shown in Fig.11. When a MD is collocated with the 10-20 day monsoon low, vigorous and well-organized convection and clear cyclonic flow were observed in Fig.11a during the maximum phase of the 10-20 day mode. During the minimum phase, a clear opposite was observed (not shown here but refer to Yoon 1999 or Yoon and Chen 2005). It is noted here that at last day (June 26th 1979), less intense rainfall and organized cyclonic flow are observed than August 10th 1979 in Fig.11. Although it is only a case, the result implies the strength of the 10-20 day monsoon mode is weaker than that of the 30-60 day mode.

Comparison of the MD water budget between the maximum and the minimum phases of the 10-20 day monsoon mode (Fig.10b) reveals that the hydrological processes of the MD are also affected by the 10-20 day monsoon mode through change of convergence of water vapor flux, just like the 30-60 day mode. Only difference is the weaker intensity. Using average from Day 1 to Day 5, impact of MD by the 10-20 day monsoon mode is about 15% weaker than the 30-60 day mode (Yoon and Chen 2005).

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[8] MERRA: Modern Era Retrospective-Analysis for Research and Application. More details can be found in http://gmao.gsfc.nasa.gov/merra/

[9] CFSR: The Climate Forecast System Reanalysis. More details can be found in http://cfs.ncep.noaa.gov/cfsr/

[10] ERA-interim: newer version of ECMWF reanalysis. More details can be found in http://www.ecmwf.int/research/era/do/get/era-interim
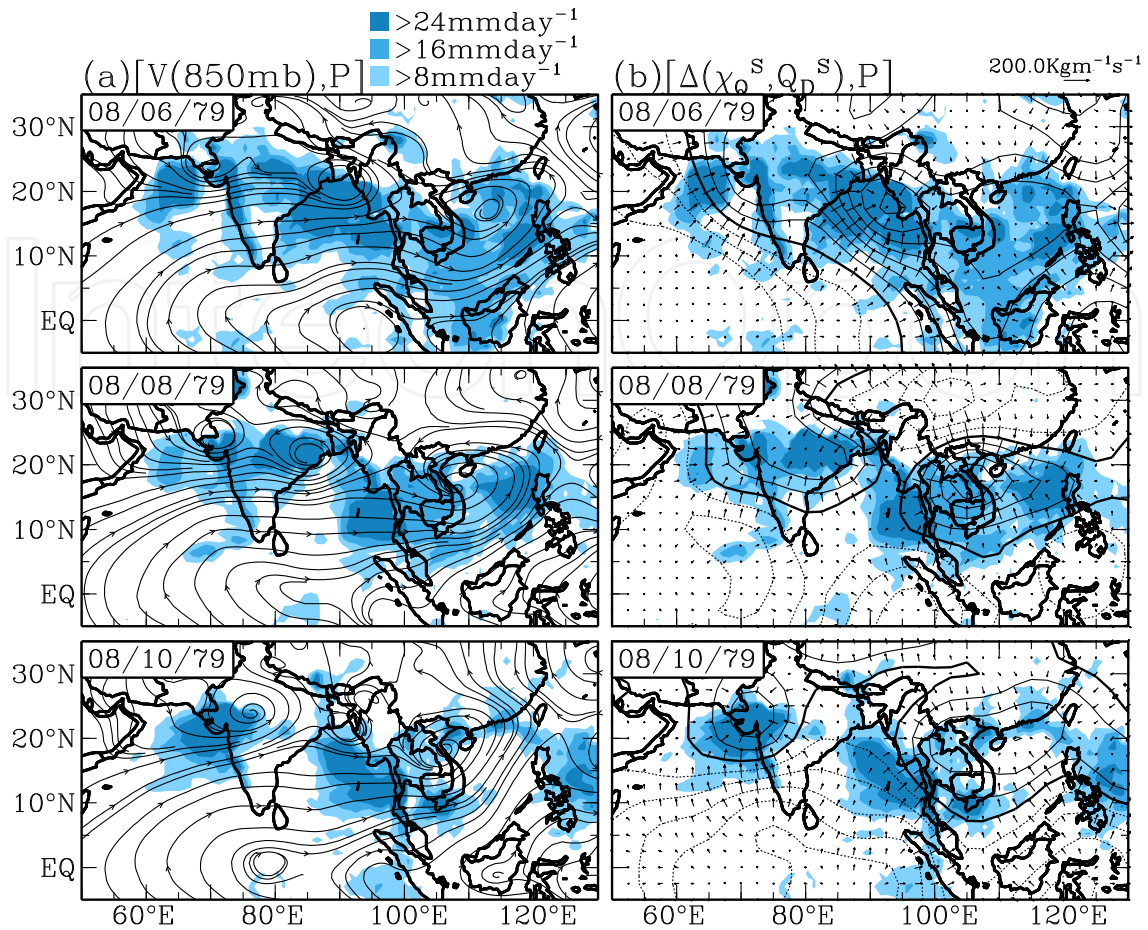
Fig. 9. (a) The 850-mb streamline charts superimposed with P, and (b) [$\Delta(\chi_Q, Q_D), P$] of a MD case during the active phase of the 30-60 day mode from August 6th 1979 to August 10th 1979. The contour interval of $\Delta(\chi_Q)$ is $2.0 \times 10^8 m^2 s^{-1} g \cdot kg^{-1}$, and $\Delta(\ )$ is departure from the summer mean.
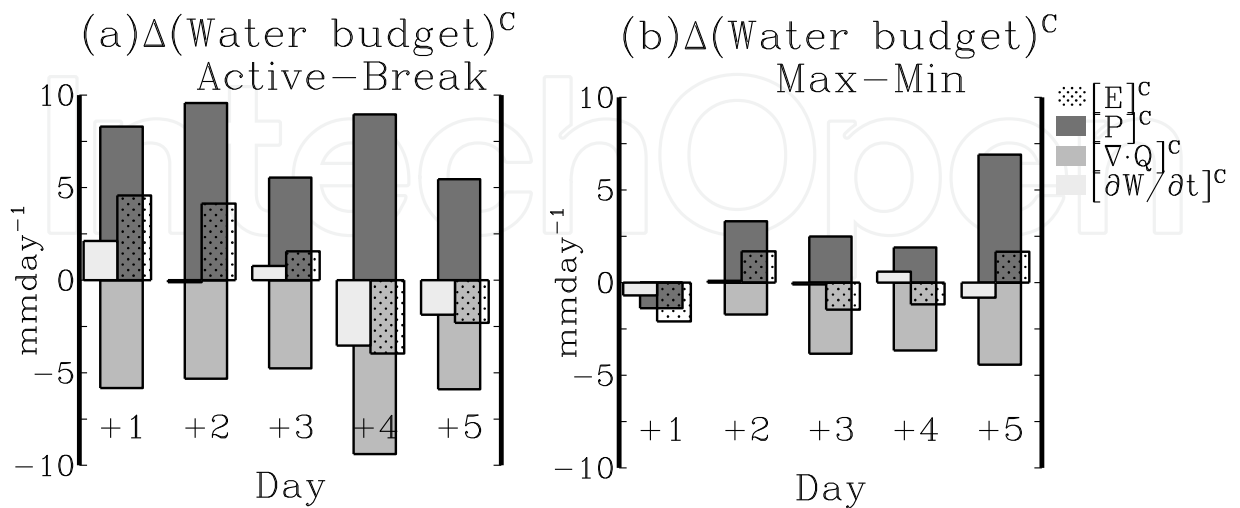

Fig. 10. Difference of the MD water vapor budget between the active and break phases of the 30-60 day monsoon mode (a) and the maximum and minimum phases of the 10-20 day monsoon mode (b).
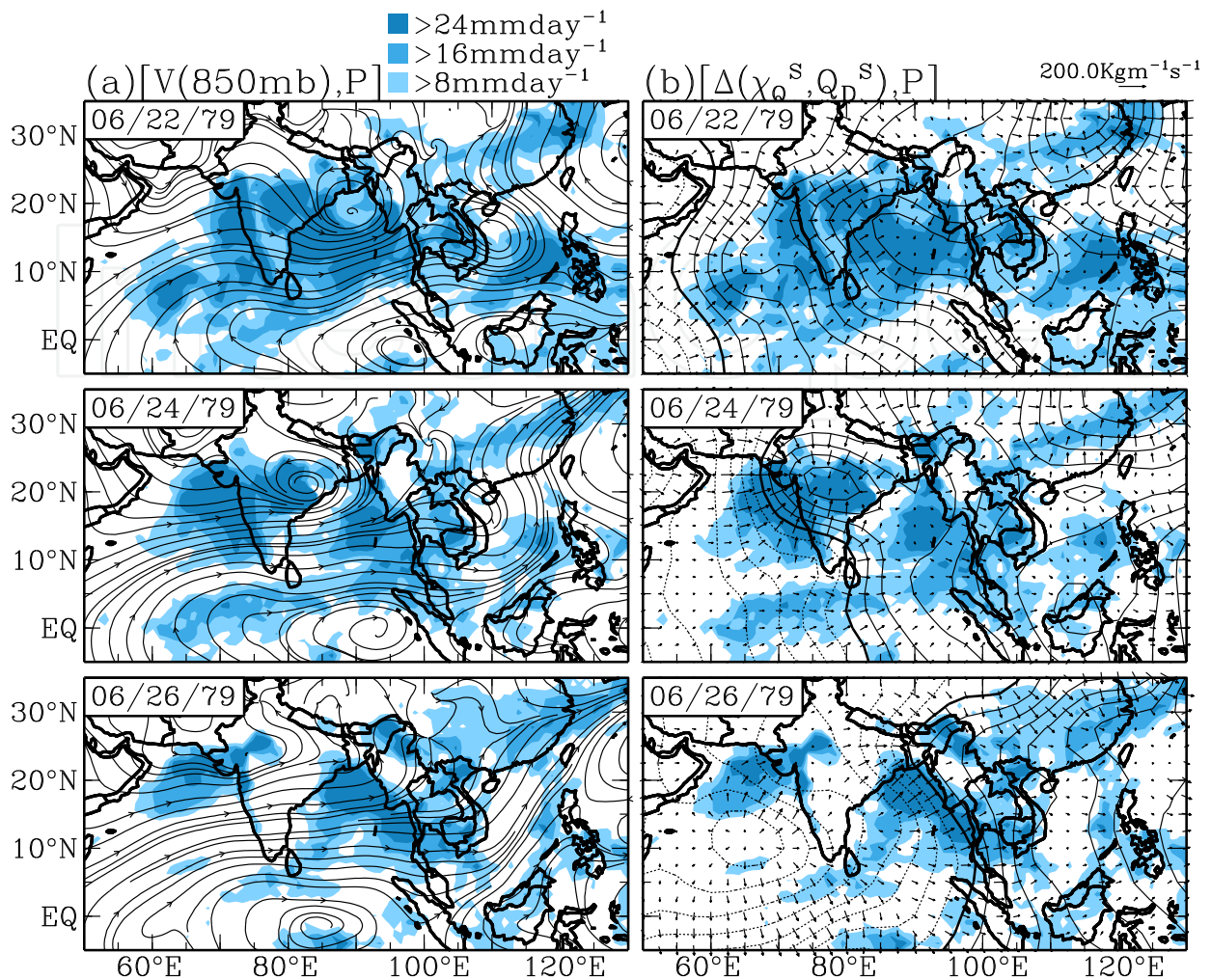
Fig. 11. Same as Figure 9 except the maximum phase of the 10-20 day mode from June 22nd 1979 to June 26th 1979.

5.2 Interannual variation of the MD

A significant year-to-year change of rainfall is found over the Indian summer monsoon region. Most well-known cases are drought occurred in 1987 and flood in 1988 (Krishnamurti et al. 1989). Numerous attempts have been made to explain the mechanisms responsible for this interannual variation of the Indian summer monsoon, such as (i) interannual variation in the sea surface temperature over the tropical Pacific (Palmer et al. 1992, Chen and Yen 1994, Ju and Slingo 1993, Slingo and Annamalai 2000), (ii) land-surface feedback including both long-memory in soil moisture (Meehl 1994) or snow coverage over the central Eurasian continent (e.g., Shukla 1987, Verneka et al. 1995), and (iii) interannual variation of occurrence frequency of MDs (Chen and Weng, 1999). As shown earlier, the MD can be responsible up to 45 – 55% of the total monsoon rainfall. Thus, it is conceivable that any change in the number of the MD could result in significant change in total monsoon rainfall.

To test this hypothesis, total precipitation over the northern part of India (65°E – 90°E and 15°N – 25°N) and occurrence frequency of the MD during summer monsoon season are displayed together in Fig.12 with warm and cold years indicated with red and blue color.

Definition of warm/cold years[11] were adopted from Chen and Weng (1999) which is based on the sea surface temperature (SST) over the NINO3 region (150°W – 90°W and 5°S – 5°N) as an indicator of thermal condition of the eastern tropical Pacific. Drought in 1987 and flood of 1988 are well captured by the total Indian monsoon rainfall index (Fig.12b). On the other hand, more (less) MDs occurred during cold (warm) years of NINO3 (Chen and Weng 1999, Yoon 1999). However, overall correlation between the occurrence frequency and total monsoon rainfall is very low. This clearly indicates that hypothesized mechanism (iii) that occurrence frequency of the MD regulates interannual variation of the Indian summer monsoon rainfall cannot be substantiated.
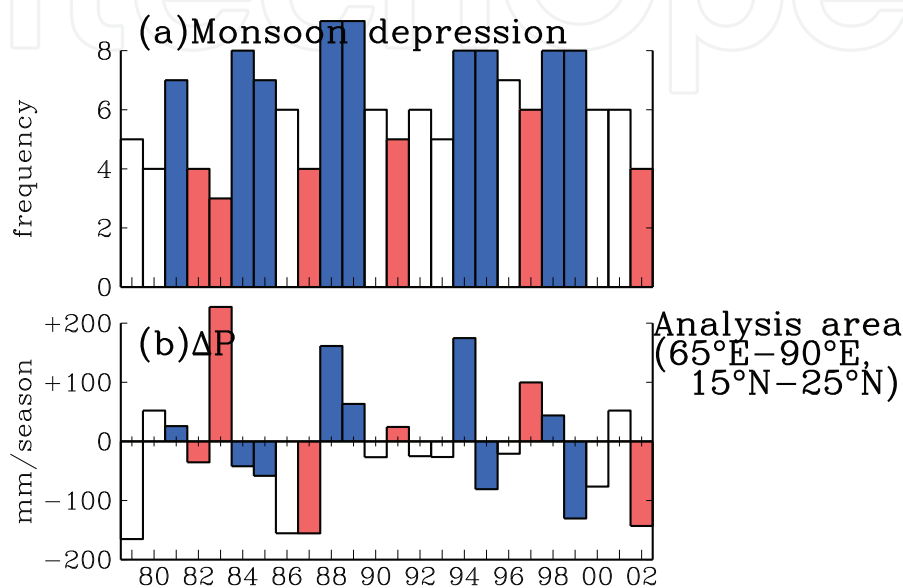
Fig. 12. Histogram of occurrence frequency of the MD at each year from 1979 to 2002 in (a), and total monsoon rainfall over the northern India (65°E – 90°E and 15°N – 25°N) in (b). Warm and cold years of the eastern Tropical Pacific was indicated by red and blue colors. This was based on SST anomaly over the NINO3 region (150°W – 90°W and 5°S – 5°N).

The occurrence frequency of MD (Fig.12a) appears to be well coherent with the NINO3 index, an indicator of the tropical Pacific climate anomaly during summer season. The tropical Pacific serves as a source for the tropical synoptic/meso-scale disturbances that propagate into the Bay of Bengal and then become the MD (Fig.5). In other words, occurrence frequency of the MD is more an indicator of the tropical Pacific weather and climate, but not a direct indicator of the strength of the Indian summer monsoon (Chen et al. 1997, Chen and Weng 1999, Yoon 1999). However, it is noted that the interannual variation of occurrence frequencies of these low-pressure systems has a significant impact on the total rainfall over Indochina (Chen and Yoon 2000b).

In summary, although MDs are responsible for a large fraction of Indian summer monsoon rainfall, the MD itself is not a decisive factor of the total monsoon rainfall change from one year to another. Second, interannual change of the MD occurrence frequency is closely linked to that of the WTP-SCS region which is closely related to the East Asian summer monsoon (Chen and Yoon 2000b, Chen and Weng 1999). Third, the interannual variation of

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[11] We have warm summers of 1982, 1983 1987 1991, 1997, and 2002 and cold summers of 1981, 1984, 1985 1988, 1989, 1994, 1995, 1998, and 1999.

the Indian summer monsoon is independent from that of the East Asian monsoon. For example, Lau et al. (1999) and Chen and Yoon (2000b) propose the dynamical and hydrological difference between two major summer monsoon regions: East Asian summer monsoon and South Asian summer monsoon. These studies imply more complicated interaction between two monsoons with global SST on the interannual timescales. Also, there are some prominent hypotheses were not fully discussed or tested here (e.g., Krishnan et al. 2010). Based on these results, more comprehensive research on the interannual variation of the Indian summer monsoon is needed.

6. Concluding remarks

Summer monsoon rainfall is a critical component of human activity over the south Asian regions, where more than one billion people live. Various weather systems contribute this monsoon rainfall. One of the prominent systems is the MD. In this article, we have reviewed various characteristics of the MD. Compared other tropical storms such as hurricane or typhoon, this low-pressure system stands out because it stays mainly over the landmass not over the warm tropical oceans. Because of this special circumstance, its dynamical and hydrological properties are different from other tropical storms. The most important property is that its major water source is not from evaporation from the warm ocean but rather from converging atmospheric water vapor flux.

Several important findings are summarized as follows:

- Our estimation based on the composite of the MD reveals that up to 45% – 55% of total summer monsoon rain is brought by MDs. About 60% of total rainfall is maintained by the converging atmospheric water vapor flux (Fig.5b).
- The life cycle of the Indian summer monsoon (onset – break – revival – withdrawal) is constituted by two intraseasonal modes. As the Indian summer monsoon undergoes its life cycle, convergence of the water vapor flux over India fluctuates coherently.
- The stream function tendency generated by vortex stretching over the west-southwest sector of a MD is vital to its westward propagation, which is located over the upward branch of an east-west circulation across a system. Furthermore, this upward branch is maintained by the latent heat released due to strong convective rainfall. The dynamical and hydrological processes can be established by feedbacks among convective rainfall, diabatic heating, east-west circulation, and divergent circulation.
- The hydrological cycle of the MD is intensified (weakened) by the convergence (divergence) of water vapor flux associated with two intraseasonal modes during active/maximum (break/minimum) monsoons. Also it is revealed that the 30-60 day mode is more effective in modulating the MD than the 10-20 day mode.
- A poor correlation between the MD occurrence frequency and total Indian monsoon rainfall is observed in interannual timescales, despite the fact that MD is responsible for about half of total rain. It is likely that other mechanisms listed in Section 5 is critical in determining the interannual variability of the Indian summer monsoon. Further work is needed in this regard.

Close interaction between the diabatic heating due to strong convection and divergent circulation is similar to the CISK mechanism in the tropics. Difference between the MD and a conventional CISK can be seen in its propagation property (e.g. Hayashi 1970, Lindzen 1974, and many others). This rather complicated dynamical and hydrological processes of the MD

provide some challenges to the regional and global climate models. For example, to properly simulate extreme rainfall under global warming, these need to be well simulated by regional or global climate models (e.g., Ratnan and Cox 2006, Vinodkumar et al. 2008 and 2009).

About half of the Indian summer monsoon rainfall is produced by MDs. Thus, the prediction of MDs is important. In order to accurately predict the intensity, propagation, and rainfall of a MD, basic features in various spatial and temporal time scales, such as the Tibetan high, the monsoon trough, monsoon westerlies, thermal contrast, and slowly-varying tropical variabilities should be properly simulated. Also, close interaction between the diabatic heating due to strong convection and divergent circulation is similar to the CISK mechanism in the tropics. Difference between the MD and a conventional CISK can be seen in its propagation property (e.g. Hayashi 1970, Lindzen 1974, and many others). This rather complicated dynamical and hydrological processes of the MD provide some challenges to the regional and global climate models. For example, to properly simulate extreme rainfall under global warming, these need to be well simulated by regional or global climate models (e.g., Ratnan and Cox 2006, Vinodkumar et al. 2008 and 2009). Also, simulation of this complicated interaction will serve as a good test-bed for newly developed physical parameterization.

One of the aspects that have not been discussed in this chapter is the long-term change of the MDs. This is an interesting issue given the fact that (1) our climate has been rapidly changing due to human activities and (2) more importantly that most of the extreme rainfall cases are associated with the MD (Dhar and Nanargi 1995, Sikka 2000). It was found that the number of extreme rainfall events have increased since 1976 (Goswami et al. 2006). At the same time, the number of the MD has been reported decreasing since 1976 (Stowasser et al. 2009, Kumar and Dash 2001, Sikka 2006, Ajayamohan et al. 2010, Rao et al. 2008, Rao et al. 2004, Mani et al. 2009). This contrasting trend – increasing extreme rainfall but reducing total number of depressions – is an interesting area that the community needs to invest resources. Another aspect that needs our attention is the role of aerosol on Indian monsoon, which was pursued as a forcing agent in long-term change of the summer monsoon (e.g., Chung and Ramanathan 2007, Lau and Kim 1999, Bolasina and Nigam 2008). Due to rapid increase of anthropogenic emissions over Asian countries including India and China, roles played by aerosol on the Asian monsoon system need to be well understood.

7. Acknowledgement

8. References

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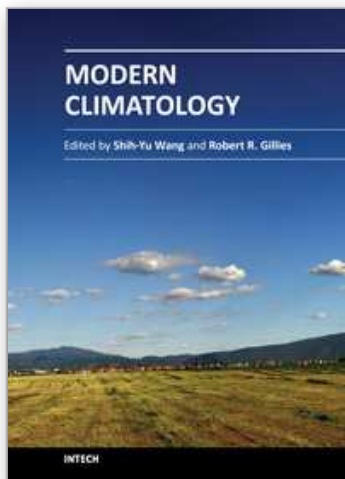
Modern Climatology

Edited by Dr Shih-Yu Wang

Climatology, the study of climate, is no longer regarded as a single discipline that treats climate as something that fluctuates only within the unchanging boundaries described by historical statistics. The field has recognized that climate is something that changes continually under the influence of physical and biological forces and so, cannot be understood in isolation but rather, is one that includes diverse scientific disciplines that play their role in understanding a highly complex coupled "whole system" that is the earth's climate. The modern era of climatology is echoed in this book. On the one hand it offers a broad synoptic perspective but also considers the regional standpoint, as it is this that affects what people need from climatology. Aspects on the topic of climate change - what is often considered a contradiction in terms - is also addressed. It is all too evident these days that what recent work in climatology has revealed carries profound implications for economic and social policy; it is with these in mind that the final chapters consider acumens as to the application of what has been learned to date.

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