Theor Appl Climatol (2014) 117:277–291
DOI 10.1007/s00704-013-1006-x

ORIGINAL PAPER

Diurnal variation in the initiation of rainfall over the Indian subcontinent during two different monsoon seasons of 2008 and 2009

Shouraseni Sen Roy · Soma Sen Roy

Received: 13 November 2012 / Accepted: 27 August 2013 / Published online: 14 September 2013

Abstract In the present study, the diurnal variations in the time of initiation of rainfall, during two contrasting monsoon seasons of 2008 (below normal) and 2009 (normal) over the Indian subcontinent and surrounding oceanic areas has been analyzed. Harmonic analysis was used to detect the spatial variation of the diurnal cycle of the time of initiation of rainfall, as obtained at half-hourly intervals from the Kalpana 1 satellite. In general, the diurnal cycle in the time of initiation is strongest in regions where convective clouds are predominant, while it is weaker in regions where the clouds are predominantly stratiform with long-lived medium to high cloud cover. In the interior of the subcontinent, the time of maximum mainly occurred in the afternoon to evening hours, with a distinct southeast to northwest gradation. Substantial spatial variations were detected in the diurnal patterns between a normal and below normal monsoon years. Spatially, rainfall is initiated later in 2009 compared to 2008 over most of the interior of the Indian subcontinent. The most distinct difference was observed over the core monsoon region in central India, where the diurnal patterns were stronger in 2009 compared to 2008. On the other hand, over the oceans surrounding the Indian subcontinent, the initiation times are generally earlier in 2009.

1 Introduction

The diurnal variation of tropical deep convection and precipitation plays an important role in modulating the energy budget of the climate system (Bergman and Salby 1997). Strong feedbacks exist among diurnal precipitation, shortwave radiation, and longwave radiation, thereby modulating the global energy budget and water cycle. The limited understanding of the mechanism of diurnal variation of precipitation, cloudiness and circulation, and their distribution is a leading cause of errors in rainfall estimation in weather and climate models over India. Deep convection in global climate models tends to be in phase with low-level temperature and atmospheric instability as measured by the convective available potential energy (CAPE), and thus it tends to occur earlier than observed (Chaboureau et al. 2004). Wilson and Mitchell (1986) pointed out that the global climate model cannot simulate the nonlinear feedback process between convection and radiation if the diurnal cycle is not resolved adequately, thus resulting in degraded model simulations. This is especially important in the tropics, where the rainfall is primarily of convective origin (Ebert et al. 2007). For example, the NCEP GFS Global Spectral Model (T80L18) model operational in the National Centre of Medium Range Weather Forecasting (NCMRWF) in India is unable to capture the late night/early morning maximum of rainfall in the foothills of Himalayas over the Indian region (Basu 2007).

In this context a vast amount of literature has been published on the diurnal patterns of weather processes, particularly in the western hemisphere. One of the earliest studies examining spatial patterns in the occurrence of precipitation was by Hann (1901). Since then, there have been multiple studies examining diurnal precipitation patterns in different regions of the world (Wallace 1975; Landin and Bosart 1985; Yang and Slingo 2001). Some of the specific causative processes include the role of cooling and heating processes on the relative concentration of precipitation events during night (Bleeker and Andre 1951) and on warm season rainfall over the Great Plains of the United States (Balling and Brazel 1987). Similar tendency toward nocturnal rainfall was also

S. Sen Roy
Department of Geography and Regional Studies, University of Miami, Florida, Miami, FL, USA

S. Sen Roy (✉)
India Meteorological Department, New Delhi, India
e-mail: senroys@gmail.com

observed in the coastal areas of northeastern United States, which was attributed to the land and sea breeze convergence (Landin and Bosart 1985; Schwartz and Bosart 1979; Chang et al. 1995; Dai 2001). The role of variations in local topography on the diurnal phase and amplitude was found over Africa and South America by Yang and Slingo (2001) and over Panama Bight and Pacific littoral of Colombia by Mapes et al. (2003a and b). The interaction of mountain and valley breeze resulting in a nocturnal high in rainfall activity in the high terrain areas was noted in Brazil's interior by Kousky (1980) and in the form of late evening showers in Sumatra (Mori et al. 2004). Furthermore, Tang and Reiter (1984) also suggested the role of nocturnal surface convergence in low-lying valley areas caused by downslope winds. The role of topography was also highlighted in the form of downward progression of rainfall events on the leeward slopes of high-altitude regions observed in the Rockies (Wallace 1975; Carbone et al. 1990) and the Tibetan Plateau (Asai et al. 1998; Wang et al. 2004).

Majority of the studies on diurnal rainfall patterns focus on summer season rainfall mainly because of the occurrence of greater proportion of rainfall during the warmer months. Initial studies on the diurnal cycle of rainfall activity mostly consisted of analyses of hourly station level rainfall data in the western hemisphere based on the availability of long-term data. Recently, with the availability of hourly scale data over the Indian subcontinent, such studies have become possible. The results of these studies over the Indian region revealed the concentration of peak rainfall activity during the afternoon hours (Basu 2007). However, the coastal areas experienced midnight to predawn showers, as a result of convergence of air masses (Sen Roy and Balling 2007). Furthermore, a recently published study examining the initiation of premonsoon season rainfall over northern India exhibited strong diurnal cycles (Sen Roy and Sen Roy 2010). The existence of a strong north to south gradient, with the first rainfall events occurring close to midnight hours in the foothills of the Himalayas and progressing southwards to over the northern plains toward the late afternoon hours was revealed by Sen Roy and Sen Roy (2010). The spatial variations were mainly a result of localized processes such as convection and regional scale movement of large circulatory processes. However, there are no such studies about the diurnal patterns of rainfall initiation for the monsoon season, when majority of the rainfall occurs over the Indian subcontinent.

Rain-gauge data has, so far, not proved very useful to study extended regions of the Indian subcontinent and the surrounding seas. This is due to the absence of long-term uniform, dense, and good quality rainfall data over the entire south Asian region due to geopolitical constraints and restricted accessibility. The rainfall estimates in the data sparse regions suffer from the biases of interpolation algorithms used to create a spatially continuous data field (e.g., Jeffrey et al. 2001). Given that rainfall is basically a discontinuous observation, most studies of diurnal variation in data sparse regions, therefore, have an inherent source of error due to the methods used to process the data. With the availability of fine temporal and spatial resolution satellite data, it is now possible to extend our understanding of diurnal processes in less explored regions of the world. In the recent past, the diurnal phase of rainfall patterns across the tropics have been more extensively studied as a result of the availability of Tropical Rainfall Measuring Mission (TRMM) data (Nesbitt and Zipser 2003; Mori et al. 2004; Nesbitt et al. 2006; Basu 2007; Romatschke and Houze 2011a, 2011b). For instance, Nesbitt and Zipser (2003) used TRMM data to study the regional level differences in the diurnal activity in the tropics and found the occurrence of maximum frequency of rainfall in the afternoon hours. It was caused by boundary layer destabilization resulting from daytime insolation. The evidence of a stronger diurnal cycle during summertime over the tropical continents relative to the surrounding ocean areas was mentioned by Meisner and Arkin (1987). Another study by Shin et al. (1990) involving the use of infrared brightness temperature data from Geostationary Operational Environmental Satellite (GOES) found early and late afternoon time of maximum in precipitation over oceanic areas. In addition, with the help of satellite microwave data for Amazon, Negri et al. (2000) reported the early morning peaks in rainfall events in coastal areas of northern Brazil. Premonsoon season rainfall is mostly convective and more sensitive to synoptic forcing than during the monsoon season (Romatschke and Houze 2011a). A detailed analysis of the diurnal variation over the Indian subcontinent during the monsoon season reveals that convection from all scales of weather systems over the Himalayan region has twin peaks of occurrence—late evening (about 20 LT) and early morning (about 06 LT) (Romatschke and Houze 2011b). However, the relative values of the two peaks, varies from west to east. While over the western Himalayas, where small, short-lived systems form primarily in the evening when monsoon air is forced upslope, the late evening peak is sharper. Further east, where the downslope winds form larger systems peak in the early morning, the peaks are less sharp and of almost the same frequency of occurrence. They also observed that the Indian and Myanmar west coasts have a less prominent diurnal cycle, where small- and medium-scale systems are present throughout the day, whereas the Bay of Bengal systems have a large diurnal cycle of occurrence. Therefore, with the help satellite data, it is possible to study interannual variations in diurnal patterns of weather processes in less explored regions of the world.

In the present study, we have analyzed the diurnal variation of rainfall initiation over the Indian subcontinent, using global precipitation index (GPI) rainfall estimates from the Kalpana 1 satellite (Kaila et al. 2002). As mentioned above, most studies till date have analyzed the diurnal variation of the most frequent hour of rainfall. However, in this study, the spatial patterns of diurnal variation in the first appearance of clouds with cloud top temperatures below 235 K at a grid point has been examined during two monsoon years of 2008 and 2009.

2 Datasets and methodology

The major source of uniform, continuous data over a large area in the tropics is from geostationary satellites. One of the earliest and most reliable techniques to measure large area rainfall using satellite data was developed by Arkin (1979). He found that radar-estimated precipitation was highly correlated with the fraction of the area covered by satellite infrared channel pixels colder than 235 K. Arkin and Meisner (1987) defined the measure of satellite derived rainfall as the GPI, which is the fractional coverage of brightness temperatures lower than 235 K in a grid box, calibrated to rain-rate units using the constant factor 3 mm/h. This was based on the premise that rainfall is most likely to be associated with deep clouds and thus with cold cloud tops, as observed by an infrared imager. However, when rainfall estimates by this technique for data from different satellites was later validated with ground-based observations over different regions of the world, the Arkin's technique generally overestimates rainfall occurring over land (Arkin and Ardanuy 1989), while over tropical oceans, the bias is near zero (Xie and Arkin 1995). Over the Indian region, the algorithm was observed to underestimate the rainfall amount over the western peninsular India while overestimating the rainfall over other regions, in comparison to rain-gauge data (Arkin et al. 1989). Richards and Arkin (1981) noted that the accuracy of the rainfall estimates generally decreases as the spatial and temporal resolution of the estimates is increased. However, recent studies with the three-hourly GPI estimates from Kalpana 1 satellite show close correlation of rainfall amounts with TRMM 3B42 rainfall estimates as well as surface rain-gauge data (Sen Roy et al. 2012). Furthermore, according to this study, the GPI rainfall was able to capture the general pattern of three hourly rainfall over the Indian region reasonably well. The simplicity of the technique and the small requirement of computer time make it the rainfall estimation technique of choice for most Satellite Meteorology Centres and most readily available. In the present study, the GPI were obtained from the Kalpana 1 satellite at half-hourly temporal resolution and 0.25°. latitude and longitude spatial resolution for the Indian subcontinent and the adjoining oceans (Fig. 1). The temporal extent of the data consisted of the monsoon months (June to September) during 2008 and 2009. The Kalpana 1 satellite did not transmit images at all hours uniformly during this period. There is a data gap at 1800 coordinated universal time (UTC) for much of the period under study when the sun, satellite (Kalpana 1), and the ground station antenna are located in a line. In other words, when the sun and Kalpana 1 are in the same direction as seen from the ground station, then the radio waves from Kalpana 1 are interfered with noises emitted from the sun and normal signal reception becomes impossible. These periods of sun outages occur in most geostationary satellites at a specific time of the day during a period of the year (Perek 1988). However, since at 1730 and 1830 UTC data were generally available from this satellite, any bias due to nonavailability of the 1800 UTC scan will not create a major data gap in the analysis of the timing of rainfall. There are also data gaps at various other times due to the satellite maintenance, but they are intermittent and nonuniform and amount to about 10 % of the total data period. It is worth noting, that our usage of GPI in this study has pushed the utilization of GPI beyond its realm of calibrated validity by calculating rainfall at fine spatial and temporal resolutions at different times of day. However, in conformity with previous studies, the rainfall estimate is only being used as a measure of the variability of cloudiness (Tian et al. 2005; Mapes et al. 2003a). Furthermore, due to the weakness of the GPI technique for rainfall amount estimation, we have confined our analysis to the occurrence of rainfall, rather than actual amount, similar to previous studies using satellite data (Basu 2007).

We also used reanalysis products provided by the NOAA/OAR/ESRL PSD (Boulder, CO; http://www.esrl.noaa.gov/psd/) to analyze the atmospheric variables. The data are available every 6 h on a T62 grid consisting of 192×94 grid points with a spatial resolution of 209 km and 28 vertical levels (Kalnay et al. 1996; Kistler et al. 2001). The wind and moisture fields analyzed for this study have been averaged for the entire summer monsoon period (June to September) separately for 2008 and 2009 to obtain the general features of the season.

One of the methodologies most widely used for analyzing the diurnal cycle of rainfall events consists of harmonic analysis (Landin and Bosart 1985; Balling and Brazel 1987; Basu 2007). In the present study, we compiled the data for the first occurrence of rainfall for each 0.25×0.25° grid point at half-hour resolution in a day for the entire summer monsoon period of 2008 and 2009. These data are available for the spatial domain exending between 40°N–40°S and 40°E–120°E. A table consisting of 1608 rows (one for each grid cell) and 48 columns (one for each half-hour interval) for each monsoon season was created. Next, harmonic analysis was conducted as represented by the following formula:

$$\widehat{\mathrm{P}} = \overline{\mathrm{P}} + \sum_{\mathrm{r}=1}^{\mathrm{N}/2} \mathrm{A_r} cos(r\theta - \Phi_\mathrm{r})$$

wherein the estimated rainfall initiation frequency is $\widehat{P}$; θ is derived as $2\pi X/N$ with X as the time interval, which in this case was half-hour, and N is the number of observations, i.e.,

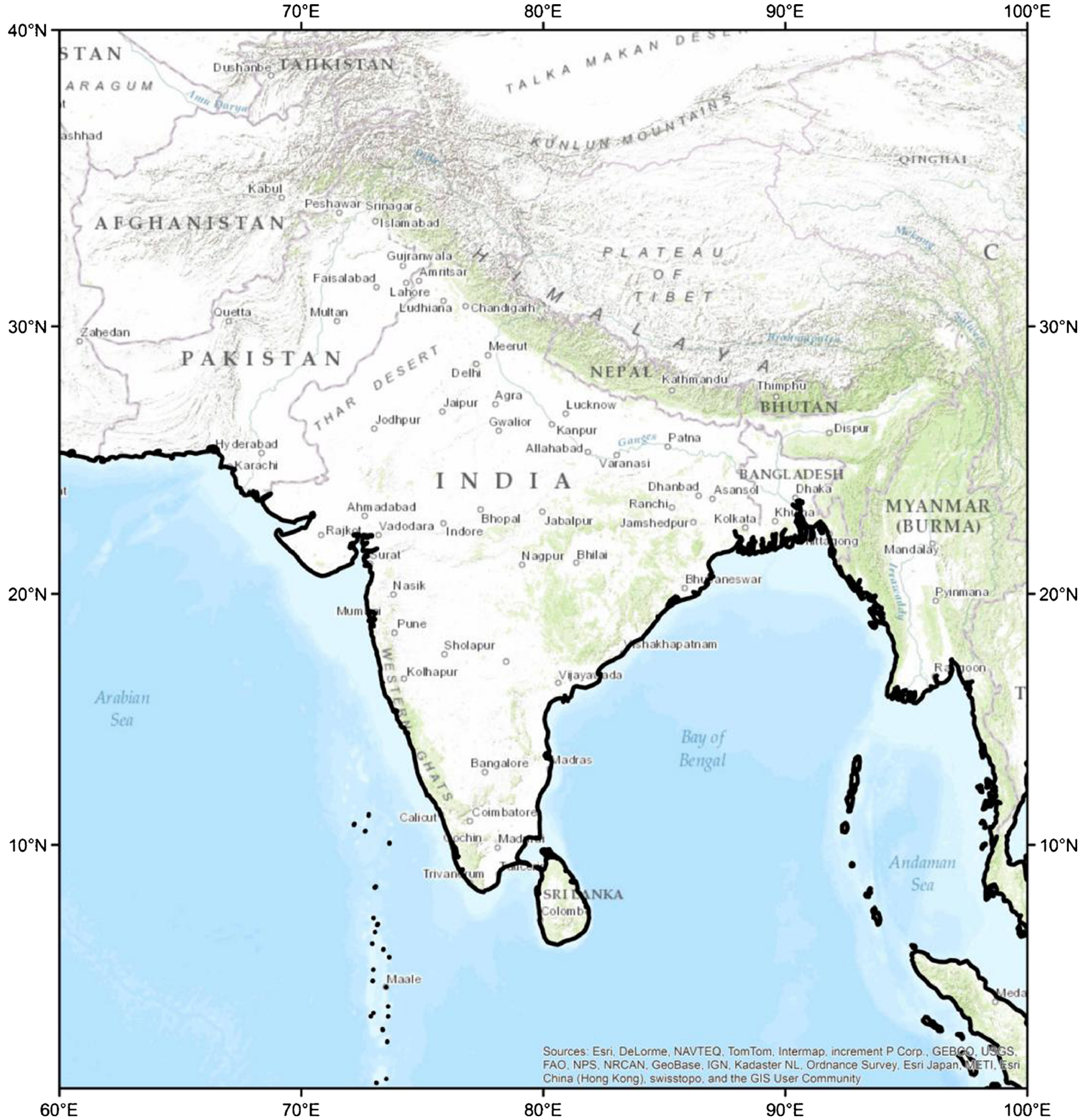


Fig. 1 General layout of the study area, Indian subcontinent

48. Φ is the phase angle of the harmonic curve, also reinterpreted as the time of maximum rainfall. $\overline{P}$ is the average hourly frequency over the N observations, and r is the frequency or number of times the harmonic curve is repeated in a day. The standardized amplitude that mainly reveals the relative importance of a diurnal cycle at each individual grid point was calculated by $A_r/2\overline{P}$. The portion of variance explained was obtained by $V_r=A_r^2/2\sigma^2$, with σ as the standard deviation of the 48 half-hourly rainfall frequency values.

The next stage in the analysis consisted of spatial analysis to reveal the spatial distribution of the variations in the diurnal cycle of monsoon rainfall patterns. Ordinary kriging method of spatial interpolation technique was used to create continuous surfaces to better visualize the spatial patterns, which can be represented by the following equation:

$$z(x) = \mu + \varepsilon'(x)$$

where μ is the fixed unknown mean of the proess, $\varepsilon'(x)$ is the spatially autocorrelated error, and x is the location (ESRI 2004). One of the advantages of ordinary kriging is that it takes ino consideration the local mean, and the actual mean is an unknown constant. A smaller domain of the entire region constituting the Indian subcontinent between 60°–100°E and 0°–40°N was considered for analysis.

3 Results and discussion

The monsoon season is the main period of rainfall over most of the Indian subcontinent. Synoptic scale monsoon systems, such as lows and depressions with lifetime of several days, move across the Indian subcontinent and bring rainfall to large swathes of the region. However, the rainfall amount and its distribution vary from year to year. This is a result of the interannual variation of the spatial and temporal distribution in the moisture and temperature profiles during the season, resulting in variation of the cloud type, organization, and cover as well as interannual scale variability of rainfall systems. The main features of the rainfall during the two monsoon seasons are summarized below.

3.1 Monsoon season of 2008

The monsoon onset over Kerala in south Indian peninsula was very close to the normal date, followed by a rapid progress of monsoon over most parts of the subcontinent. The formation of monsoon low-pressure systems was subdued in comparison with the preceding years. One of the significant features of this season was that the spatial distribution of season rainfall was almost uniform, but there were noticeable large temporal variations. There was also a substantial delay in the withdrawal of monsoon, particularly from northwest India, due to the presence of mid-latitude westerly troughs/cyclonic circulations, availability of moisture, and sporadic rainfall activity (Tyagi et al. 2009). The seasonal rainfall (1st June to 30th September, 2008), as measured by the Kalpana 1 satellite, by the GPI technique is displayed in Fig. 2 a. The intense region of rainfall over the Bay of Bengal and adjacent northeast peninsula (Orissa) is due to the formation of most low-pressure systems over this region, and their movement westward into the subcontinent along the eastwest-oriented monsoon trough from over this region. Rainfall anomaly patterns of the meteorological subdivisions of India (with respect to the 30 year average rainfall of the subdivision) reflect that almost the entire Indian subcontinent received normal rainfall during the season (Fig. 2 c).

3.2 Monsoon season of 2009

The onset of monsoon in 2009 was nearly 1 week ahead of the normal date. However, the monsoon rainfall in 2009 was the third most deficient All India Southwest Monsoon Rainfall (AISMR) year during 1901–2009. The deficient rainfall was caused by a weaker than normal cross equatorial flow during most of the season, except for a brief period between last week of June to third week of July. As a result of these anomalous features, the activity of monsoon low-pressure systems (lows and depressions) in 2009 was greatly subdued compared to previous years. There were only four depressions (two each formed over the Arabian Sea and the Bay of Bengal) and four low-pressure areas formed during the season. Most of these systems had a short life span over land and thus did not help in continuous rainfall activity. The rainfall anomaly was negative for all months and across most of the subcontinent during the season. Furthermore, the rainfall deficiency was more than one standard deviation during all months except July (Tyagi et al. 2010). It is evident from Fig. 2 b that the main rainfall region, seen over the northwest Bay of Bengal in 2008, has shifted further eastwards. As may be seen from the rainfall anomaly patterns of Fig. 2d, almost the entire north Indian subcontinent region is deficient in rainfall in 2009 as compared to 2008.

The AISMR during the monsoon season (1 June to 30 September) was 98 % of its long period average (LPA) in 2008 and 78 % of its LPA in 2009. Hence, 2008 was considered a year with normal monsoon rainfall, while 2009 was a severe drought year for India. In this context, analyzing the initiation cycle of rainfall for 2 years with contrasting monsoon rainfall will enable us to highlight the aspects of the diurnal cycle that are dependent on the dynamic and thermodynamic conditions along with topography.

The results of harmonic analysis for the time of initiation of rainfall are displayed in Fig. 3, for each year separately. The standardized amplitude, which represents the strength of the diurnal cycle, for the initiation time for 2008 and 2009 are displayed in Figs. 3a and 4a, respectively. The diurnal signal in the initiation time for both years show substantial regional variations across the Indian subcontinent. In general, the diurnal signal is strongest over northwest India (0.71 in 2008 and 0.63 in 2009), south peninsular India and neighboring Sri Lanka (0.76 in 2008 and 0.63 in 2009), over the northeast peninsular India adjoining the head of Bay of Bengal (0.60 in 2008 and 0.49 in 2009), over east central Arabian Sea (0.69 in 2008, and 0.64 in 2009), and over central Myanmar (0.70 in 2008 and 0.77 in 2009). The strength of the diurnal cycle is also strong over the Tibetan Plateau. The main difference between 2008 and 2009 is the increased strength of the diurnal cycle along the northwest- to southeast-oriented monsoon trough region over the Indian subcontinent (between about 70–88°E, 20–30°. N) in 2009 (0.37 in 2008 vs. 0.74 in 2009) as well as over central and Northwest Peninsular India (0.35 in 2008 vs. 0.71 in 2009). On the contrary, the diurnal cycle is relatively stronger over northeast peninsular India, adjoining the head Bay of Bengal in 2008. The diurnal cycle is weakest along the Himalayas, especially over northeast India, for the first hour of rainfall occurrence (approximately 0.2 in 2008, and 0.13 in 2009). However, in 2008, the strength of the diurnal cycle was greater over northern Kashmir (0.85) and peninsular India (0.42). In general, there is less preference for any particular initiation time in 2008 as compared to 2009.

Additionally, the dominance of the diurnal cycle in the diurnal variation of gridded rainfall can be assessed by variance explained (Figs. 3b and 4b). The spatial patterns are mostly similar to that observed in the case of standardized amplitude; that is, the regions with a stronger diurnal cycle generally have the most significant contribution to the diurnal variation in rainfall. However, the two zones, which receive the least rainfall during this season, namely the southeast peninsula and the Himalayan region to the north of the subcontinent, have the least significant effect on the diurnal variation in rainfall for both years. Furthermore, the patterns for variance explained had a smoother gradient in 2009, compared to the patterns observed for the standardized amplitude.

Results from previous studies indicate that generally, during monsoon season, deep and wide intense convective echoes develop over northwestern India, where the low-level layer of moist monsoon air from the Arabian Sea meets the raised layer of dry air flow advected from the Afghan and Tibetan Plateau (Houze et al. 2007). Short-lived deep convective clouds with scales of 1–2 km occur over the high terrain of the Tibetan Plateau (Ye and Gao 1979). Moreover, the northeastern peninsular India and the adjacent head of Bay of Bengal is also another zone where large, tall, and long-lived convective cloud systems develop in the afternoon hours during this season due to the strong land–sea interaction (Houze and Churchill 1987; Zuidema 2003). A strong factor for the initiation and growth of all these systems, discussed so far, lies in the upward transport of sensible and latent heat flux from the surface, which in turn is dependent on the diurnal variation of the surface temperature. It has also been widely reported that large diurnal variations in frequency of rainfall are associated with regions of deep convection. Additionally, the more intense the deep convection is, the more it is associated with heavy rainfall systems and stronger the diurnal variation (Gray and Jacobson 1977; Tian et al. 2005). On the other hand, broad stratiform echoes form further eastwards over the eastern end of the Himalayas, over north eastern India and north Bangladesh (Houze et al. 2007; Romatschke and Houze 2011b). These are generally a component of large mesoscale systems that are long-lived and do not show much diurnal variation. This explains the weaker signal in the regions where stratiform rainfall or no rainfall conditions are predominant. The same argument can be extended further to account for the higher values of the diurnal cycle (strength and significance of rainfall initiation) during 2009 as compared to 2008. The increase in the strength of the diurnal cycle in 2009 is particularly significant over central India, which coincides with the monsoon core region (73–82°E, 18–28°N; Mandke et al. 2007). The day-to-day variability of rainfall over this region is closely correlated with the AISMR (Gadgil and Joseph 2003). Their study also demonstrated that the total rainfall of a season is the result of the intraseasonal variability of the monsoon rainfall due to active, break, strong, and weak phases of monsoon rainfall. Choudhury and Krishnan (2011) noted that there is a higher abundance of isolated convective clouds over this monsoon core region during the weak or break phases of the monsoon season, as opposed to a predominance of stratiform-type clouds (mostly nimbostratus) during the active phase of monsoon. The clouds develop with smaller vertical extent (3–5 km) due to the presence of a heat low type of circulation (Gambheer and Bhat 2000). Hence, the increased occurrence of short-lived rainfall systems, driven by the diurnal cycle of the solar insolation, is reflected in the greater strength in the diurnal cycle over this region during 2009 compared to 2008. The diurnal cycle is also stronger north of 30°N and west of 70°E, which is in the domain of extratropical systems during this season. This region is characterized by short-lived weather systems with tall clouds and isolated convection (Sen Roy and Sen Roy 2010). Based on the discussion above, it can be concluded that the diurnal cycle of initiation is strongest and most significant over regions, where there is a predominance of deep convective clouds and abundance of high clouds. It is weaker in regions where large-scale stratiform clouds cause the majority of rainfall or in regions where clouds and rainfall are minimal.

We next mapped the spatial variation in the time of maximum for initiation of rainfall across the subcontinent in 2008 and 2009 (Figs. 3c and 4c). Over the sea, rainfall is initiated mostly in the morning hours, between 0000 UTC and 0600 UTC [0530 and 1130 local standard time (LST)]. The occurrence of maximum rainfall in the early morning over oceans is driven by the cloud-radiation feedback, referred to as the "static radiation-convection" and "dynamic radiation–convection" mechanisms (Gray and Jacobson 1977; Yang and Smith 2006). There is a clear separation in the time of initiation of rainfall over the land and sea during both monsoon seasons, particularly along the west coast of the Indian subcontinent, where the high orographic block in the form of Western Ghats is very close to the coast. This land–sea contrast is not so distinct along the east coast of India, where early morning initiation over the coastal seas penetrates about 100 km inland from the coast due to the low orography (Fig. 5). The most frequent time of initiation gradually shifts in a clockwise direction inland toward later initiation times, which merges with the solar diurnal cycle driven afternoon initiation further inland. Studies using INSAT 1B satellite pixel data have shown that majority of monsoon cloud systems decay within a couple of 100 km from where they form (Gambheer and Bhat 2000). This pattern of initiation at the coast, therefore, points to a propagating component of the diurnal cycle, consisting of convection initiated over the land–sea interface in the early morning and propagating inland to yield an early morning diurnal peak over the coastal landmass. Our findings are in conformity with the findings of previous studies over this region (Zuidema 2003 and Yang and Slingo 2001). Fresh convection is initiated further inland, primarily due to the interaction of large-scale factors with surface heat fluxes, and is closer to the local solar maximum in the afternoon. The afternoon initiation over land is mainly a thermodynamical response of the atmosphere to the diurnal cycle of solar radiative heating. The shorter diurnal lag between the temperature and rainfall in the interior of the subcontinent reflects the much smaller thermal inertia of land and the atmospheric boundary layer (Mapes et al. 2003a). However, the internal land areas do not have the uniform time of initiation throughout and has a southeast to northwest gradation across the subcontinent in both years. Rainfall is initiated earlier (closer to the local maximum temperature at

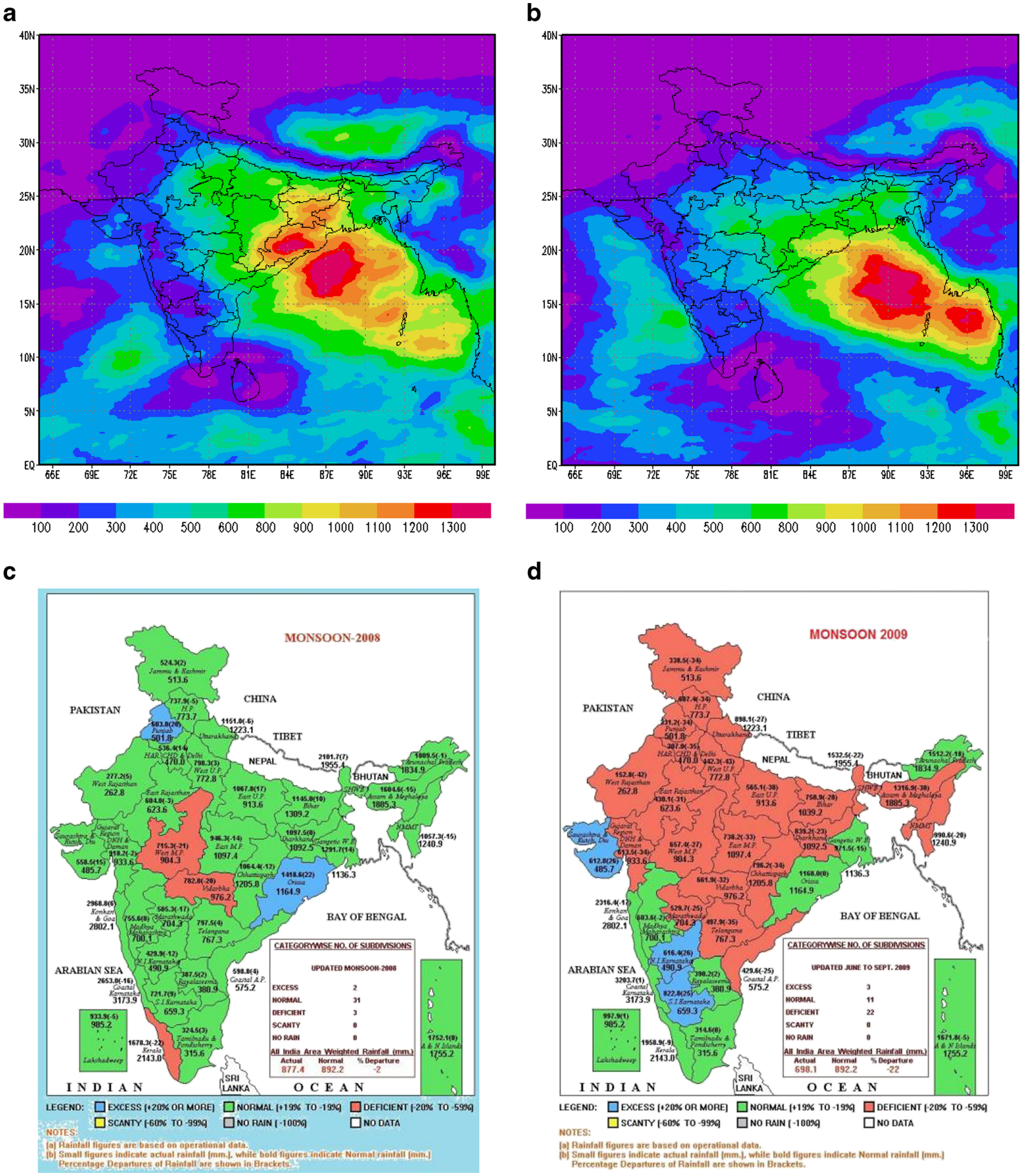


Fig. 2 Spatial distribution of seasonal total rainfall in mm for **a** 2008; **b** 2009 as measured by the GPI technique as applied to Kalpana 1 Satellite data, **c** seasonal anomalies according to daily rain-gauge data in **c** 2008 and **d** 2009. The rainfall amounts are in mm. Adapted from India Meteorological Department Monsoon Report, 2008 and 2009

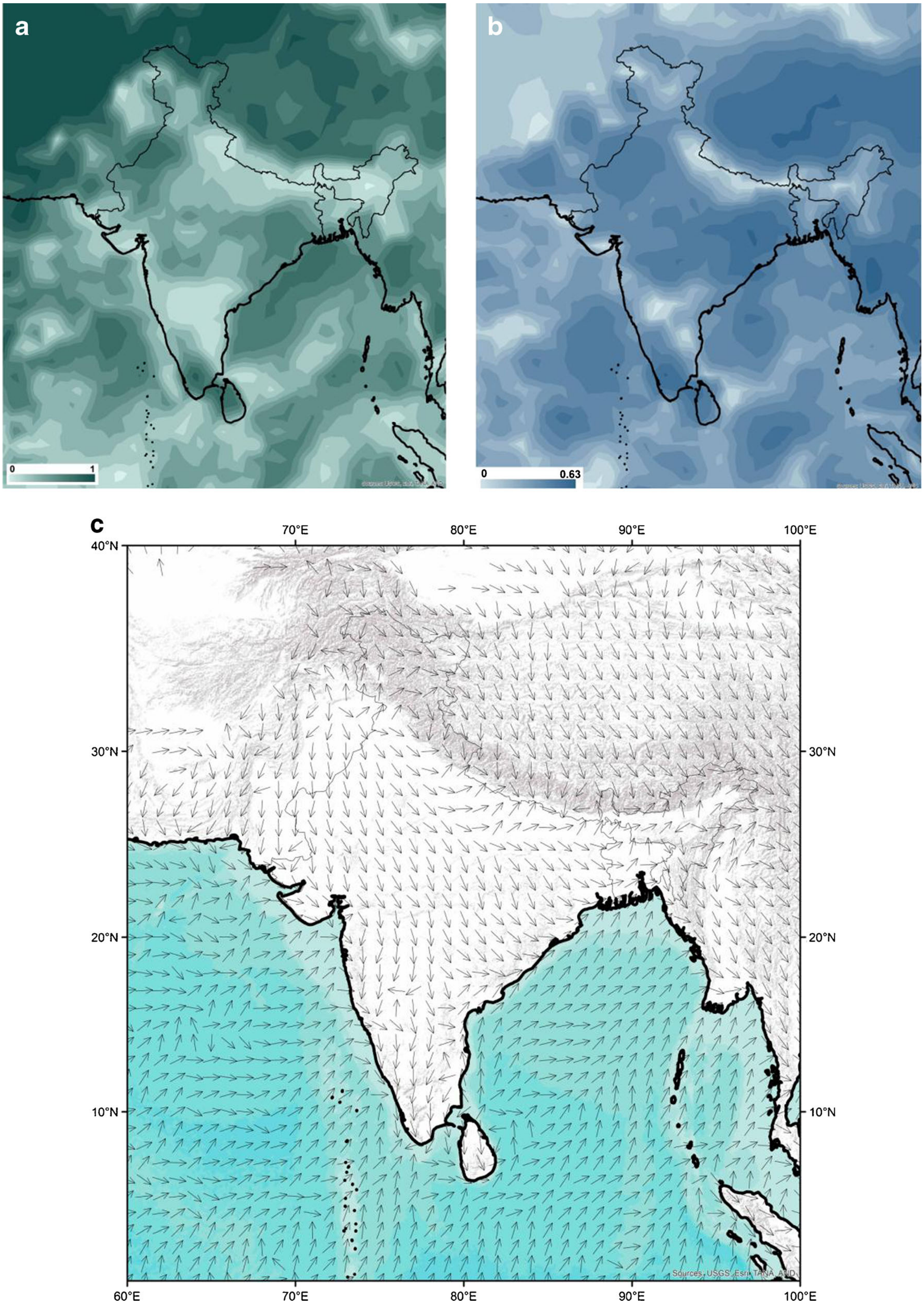


Fig. 3 2008 **a** Standardized amplitude, **b** variance explained, and **c** most frequent time of initiation of convection

about 0900 UTC or 1430 LST) over the east coast of India, with progressively later rainfall initiation times (closer to 1500 UTC or 2030 LST) over northwest India and Pakistan. The gradual clockwise progression in the initiation times

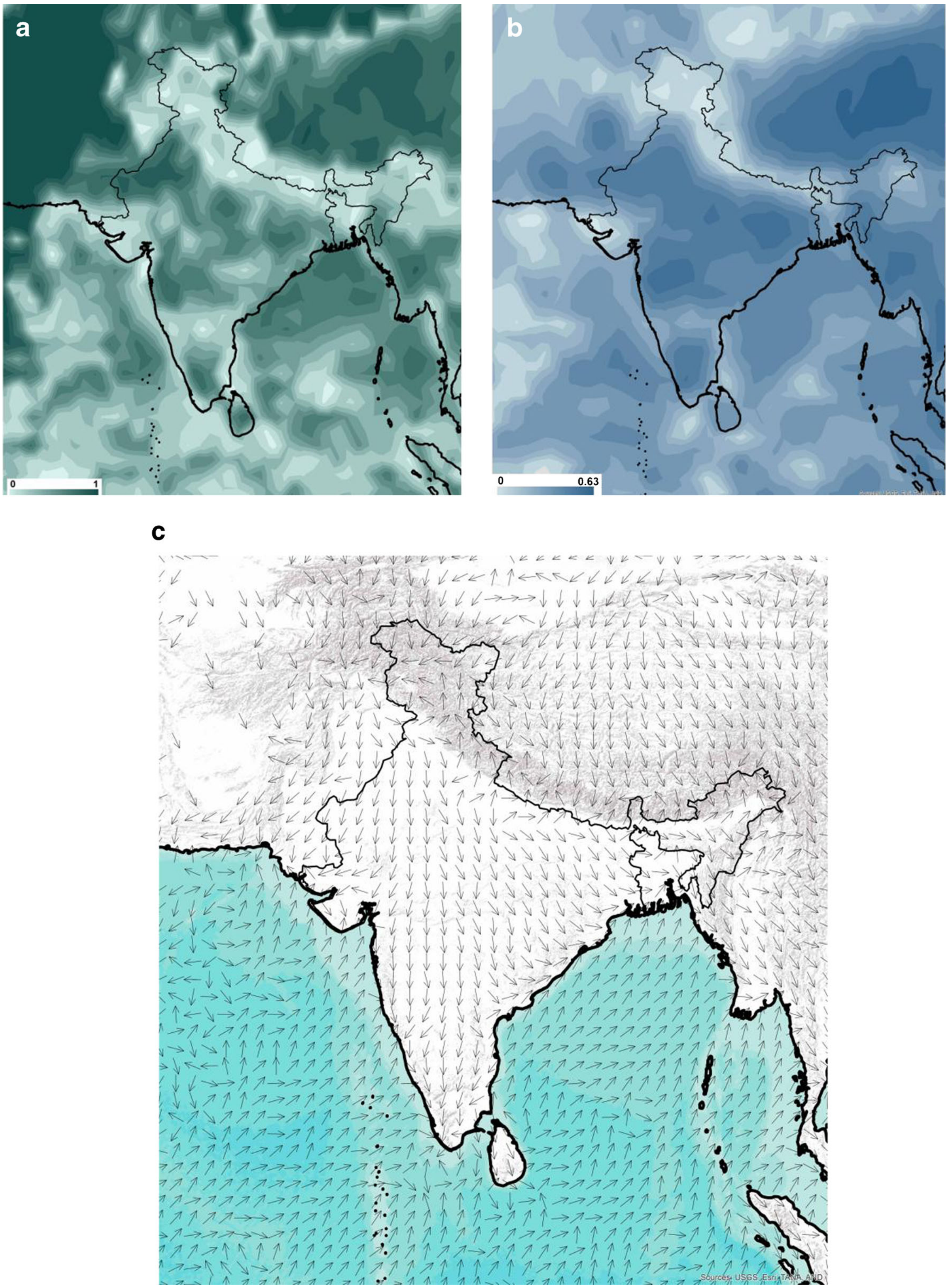


Fig. 4 2009 **a** Standardized amplitude, **b** variance explained, and **c** most frequent time of initiation of convection

toward later hours and northwestwards across the subcontinent needs to be seen in the context of the prevailing wind pattern, the scale of the weather systems, as well as the thermodynamic profile of the atmosphere during this season.

It is evident from Fig. 6a and b that the large-scale mean wind flow (average wind at the 200 to 850 h Pa wind layer) over the Indian subcontinent during 2008 and 2009 is easterly to southeasterly during this season. The large cloud systems,

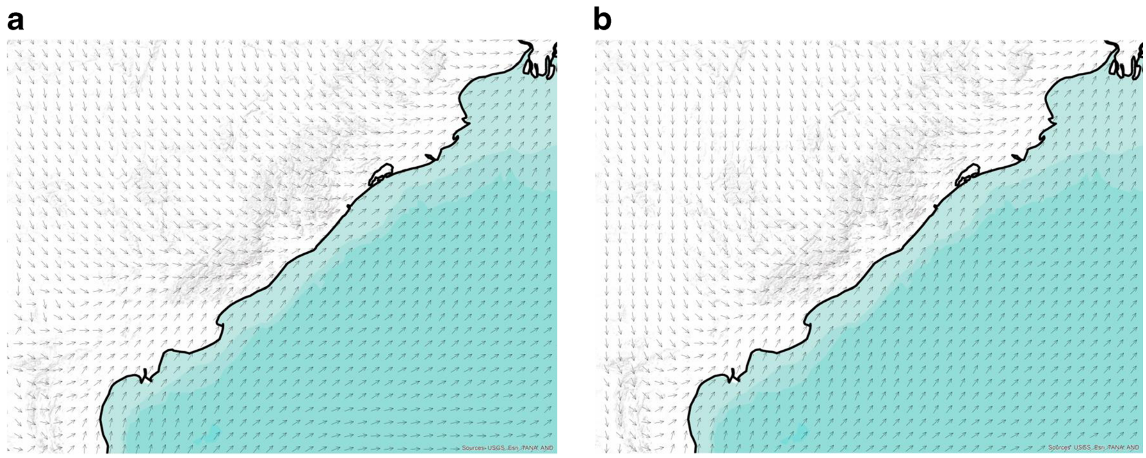
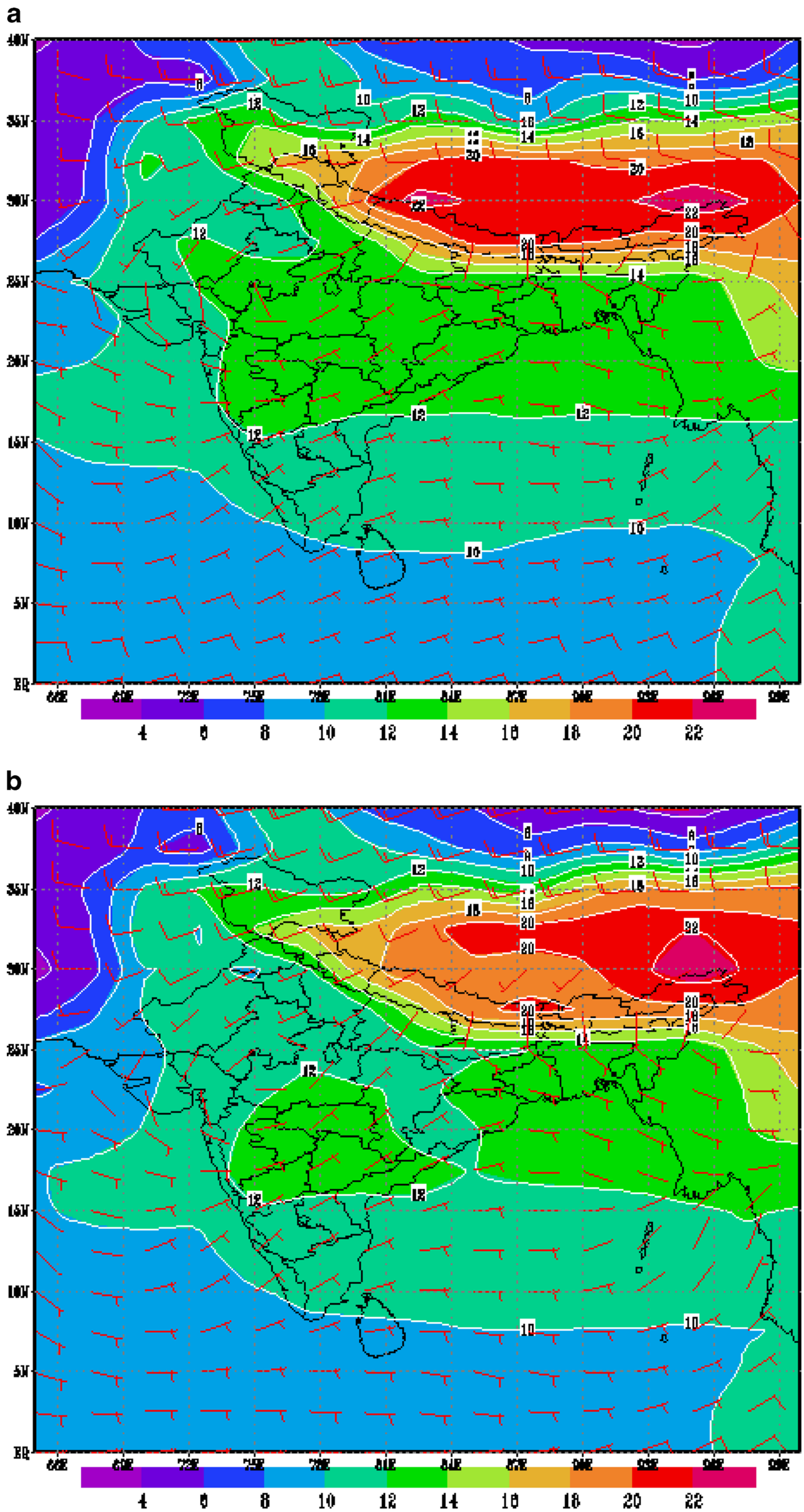


Fig. 5 Most frequent time of initiation of convection over the northeast coast of India in **a** 2008 and **b** 2009

associated with long-lived monsoon depressions originating over the Bay of Bengal, move westwards or generate new systems downstream further westwards. This may be a compelling argument in favor of later initiation downstream of the wind flow. The moisture contours for 2008 and 2009 demonstrate that the moisture in the lowest levels of the atmosphere also decreases northwestwards across the Indian subcontinent (Fig. 6a and b). A recent study using a cloud-resolving model (Meso-NH) has shown that the greater the moisture content of the lower atmosphere, the less is the time lag between the diurnal maximum of the surface temperature and the initiation of convection (Chaboureau et al. 2004). The lag in convection initiation with respect to the diurnal maximum of temperature over the northwest India is therefore sought to be explained on two bases: (a) the diurnal minimum of the level of free convection (LFC) and lifting condensation level between 12 and 18 h of local time and (b) the diurnal maximum of vertical velocity in the night (Zhang 2003). The diurnally varying tropospheric vertical velocity helps the parcel break through the inhibiting energy barrier (if it is small), and convection then occurs. A combination of the two factors therefore produces the lagged maximum in initiation of convection. The more the moisture content of the atmosphere, the lower is the average height of LCL during the day and smaller are the CINE values. Convection is then initiated earlier, with smaller vertical velocity values, closer to the local maximum of CAPE and diurnal maximum of temperature. Observational studies over U.S. Southern Great Plains and modeling studies using the global model (NCAR CCM3) also bear out the above conclusions (Zhang 2003). Since the moisture over northwest India is mostly borne from the Arabian Sea and the Bay of Bengal by the large monsoon synoptic systems, it appears that the large-scale forcing plays a more important role than the local surface heat fluxes in initiating continental convection, although the character of the convective systems is decided by a close interaction of land surface processes, sensible and latent heat fluxes, topography, and large-scale factors (Medina et al. 2010). Further eastwards over Myanmar and in the north over the Tibetan Plateau, the afternoon maxima in convection initiation linked to solar insolation tend to dominate. Along the foothills of the Himalayas, there is a discontinuity in the timing of initiation. The northern slopes of the Himalayas over the Tibetan Plateau are dominated by afternoon initiation of convection. However, the southern slopes of the Himalayas are dominated by an early morning maximum. The initiation time progresses in a clockwise direction downslope of the mountains until it merges with the afternoon maximum in the time of initiation over the north Indian Plains. Previous studies have attributed this pattern of convective initiation to the katabatic downslope wind flow (Basu 2007; Sen Roy and Balling 2007). Once the rainfall activity has begun over the higher elevations, these areas begin to cool due to rainfall. Cool outflow from rainstorms increases the down valley tendency of wind at low levels. The low-level convergence zone over windward slopes, especially in the western Himalayan region, is aided by the interaction between mountain-forced gravity waves and atmospheric instability as suggested by Barros and Lang (2003) and Bhatt and Nakamura (2005).

The overall pattern of initiation time over the entire region has distinct differences between the two years. Rainfall is initiated later in 2009 compared to 2008 over most of the interior of the Indian subcontinent, except in the southern slopes of the Himalayas (Fig. 7). The difference is most pronounced over southern peninsular India. The later initiation times in 2009 monsoon season, over most of the Indian subcontinent, may be caused by decreased levels of moisture in the lower atmosphere over most of the subcontinent (Fig. 6a and b). The decrease in in situ moisture content over most of the subcontinent increases the lag between the daily solar maximum and the initiation time of maximum over the continental region. It also makes the convection less dependent on local processes and more dependent on large-scale processes. On the other hand, over the oceans surrounding the Indian

Fig. 6 The mean wind flow (average between 200 and 850 h Pa) over the Indian subcontinent in knots superposed with contours of specific humidity (g/kg) at 850 h Pa during the monsoon season **a** 2008 and **b** 2009. The colour scale below is with reference to specific humidity values



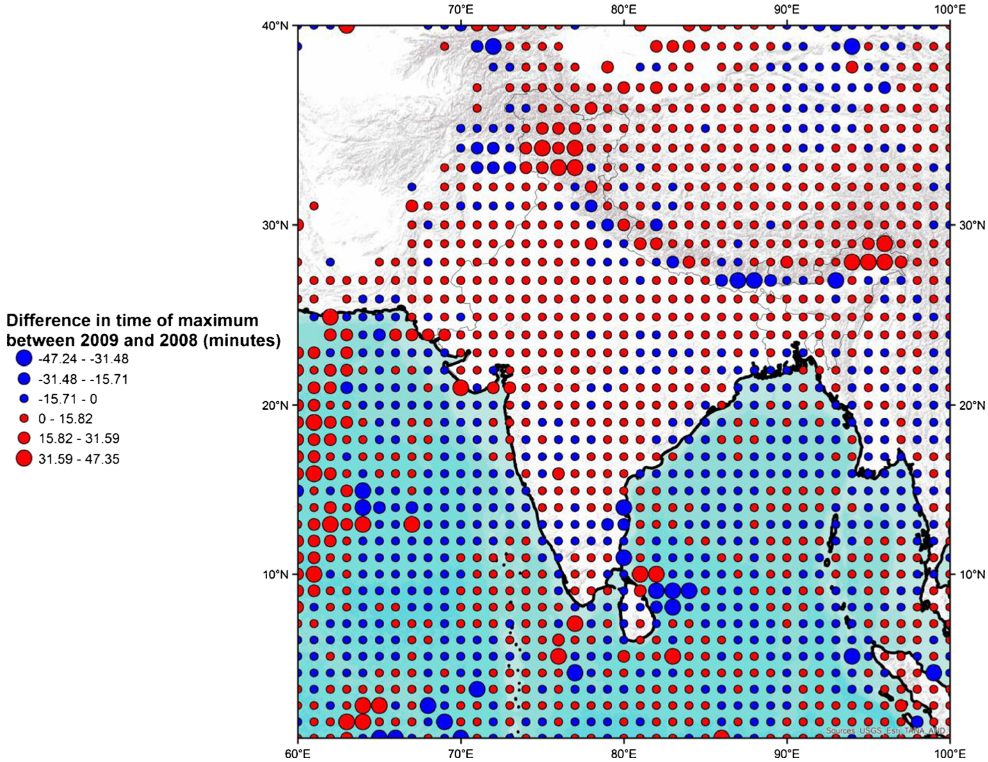


Fig. 7 Difference in initiation time between 2008 and 2009 (2008 minus 2009). The *red circles* indicate grid points, where rainfall was initiated earlier in 2008 as compared to 2009 and vice versa for the *blue circles*

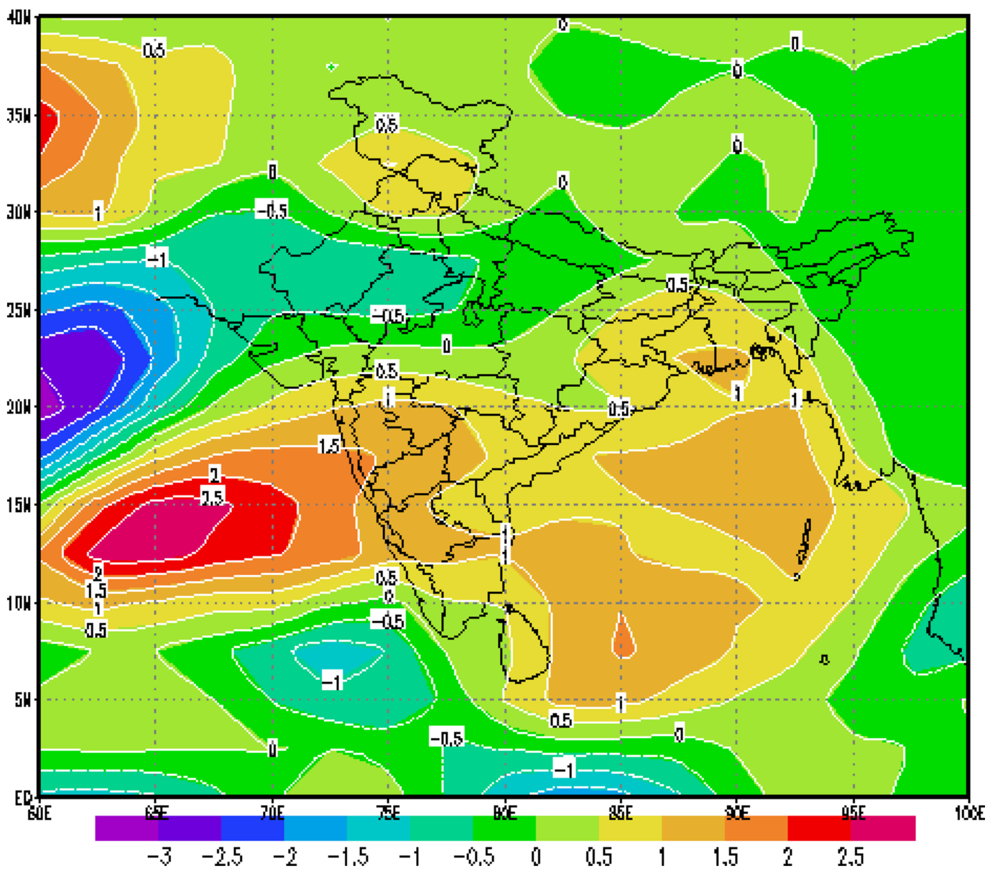


Fig. 8 The difference in the magnitude of shear (m/s) between 2009 and 2008. Shear values are calculated between 700 and 925 h Pa. Values greater than zero indicate regions where the shear was higher in 2009 as compared to 2008 and vice versa

subcontinent, the initiation times are generally earlier in 2009. Gray and Jacobson (1977) first suggested that the diurnal variability over oceanic regions of the tropical east Atlantic was also modulated by the relatively stronger degree of low-level vertical wind shear over the region. This resulted in the frequent appearance of squall lines and other types of line convection over the region. Our observations from the NCEP reanalysis data show that the average shear over the Bay of Bengal as well as the Arabian sea between 925 and 700 h Pa is stronger during 2009 compared to 2008 (Fig. 8). The stronger shear, in all likelihood, triggers the early initiation over the region in 2009.

4 Conclusion

Unlike previous studies, this study focuses on the most frequent time of initiation of rainfall over the Indian subcontinent and the surrounding seas during two different monsoon seasons of 2008 (below normal monsoon season) and 2009 (normal monsoon season). High-resolution half-hourly rainfall data from the Kalpana 1 satellite has been used for this purpose. The results of our analyses, while in conformity with the findings of previous studies examining maximum occurrence of rainfall, provide deeper insights into the seasonal scale variations in the diurnal activity of rainfall initiation over the Indian subcontinent. For instance, there is sparse coverage of station level data particularly in the less accessible high-altitude areas of Kashmir, which is not a hindrance in case of satellite data. Additionally, the results of the study can be used to identify the role of local level vs. regional scale weather processes. The main findings of our study are summarized below.

- The diurnal cycle of variation in the time of initiation is strongest in regions where the clouds are typically convective, which have a preferred time of formation. The patterns over land areas are strongly tied to the diurnal cycle of solar insolation.
- The diurnal cycle is the weakest in regions where the clouds are primarily stratiform and long-lived medium to high clouds dominate. In conformity with the findings of previous studies, these regions overlap with regions where monsoon seasonal accumulated rainfall is highest (Schumacher and Houze 2003). The diurnal cycle is also weaker over the southeast peninsular coast of India, where rainfall is scant during this season and not many weather systems are observed.
- The most distinct difference in the diurnal cycle between the two monsoon seasons is over the core monsoon region in central India, with stronger observed values in 2009. Previous studies have shown that this may coincide with the more frequent development of weak and isolated convective clouds over this region during weak monsoon seasons as compared to more long-lasting cloud systems with stratiform outflows during strong monsoon seasons.
- Over the land regions of the subcontinent, the afternoon to evening initiation times show a clear southeast to northwest gradation, with the initiation timing progressing in a clockwise manner northwestwards. The mean wind flow and the low-level moisture field (850 h Pa) indicate the significance of the large-scale processes vs. local processes in triggering convection. Further eastwards, where the in situ moisture content is higher, the maximum time of initiation is almost in phase with the diurnal cycle of temperature. However, in the western part of the study area, despite the local land surface heating and the resulting pumping of sensible heat flux into the atmosphere, convection is not triggered immediately in the absence of adequate moisture. Large-scale processes coupled with the diurnal maximum of the upward velocity in the late evening initiates convection later. Similar mechanism leads to the later initiation times during 2009, when low-level moisture field is weaker than in 2008.
- Over the southern slopes of the Himalayas, an anomalous midnight convection initiation is observed. The time of initiation progresses clockwise downslope before merging with the continental afternoon maximum. Other studies have attributed this pattern of convective initiation to the katabatic downslope wind flow. The downdraft from these cells increases the down valley tendency of wind at low levels, thereby triggering convection at later times.
- Late night to early morning maximum in initiation is seen over the oceanic regions during this season. It is important to note the observed change in initiation timing over the oceanic regions between 2008 and 2009. The average higher vertical wind shear in the lower levels observed in 2009, over the Bay of Bengal and the Arabian Sea, in the NCEP reanalysis dataset suggests a dynamic support to the convection initiation over this region during this season. Similar studies over the tropical east Atlantic revealed the frequent appearance of squall lines and other types of line convection over the region in association with increase in the vertical shear (Gray and Jacobson 1977).

References

Arkin PA (1979) The relationship between fractional coverage of high cloud and rainfall accumulations during GATE over the B-Scale array. Mon Weather Rev 107:1382–1387

Arkin PA, Ardanuy PE (1989) Estimating climatic-scale precipitation from space: a review. J Clim 2:1229–1238

Arkin PA, Krishna Rao AVR, Kelkar RR (1989) Large scale precipitation and outgoing longwave radiation from INSAT-1B during the 1986 South West Monsoon season. J Climate 2:619–628

Arkin PA, Meisner BN (1987) The relationship between large-scale convective rainfall and cold cloud over the Western hemisphere during 1982-84. Mon Wea Rev 115:51–74. doi:10.1175/1520-0493(1987)115<0051:TRBLSC>2.0.CO;2

Asai T, Ke S, Kodama Y (1998) Diurnal variability of cloudiness over East Asia and the Western Pacific Ocean revealed by GMS during warm season. J Meteorol Soc Jpn 76:675–684

Balling RC Jr, Brazel SW (1987) Diurnal variations in Arizona monsoon precipitation frequencies. Mon Weather Rev 115:342–346

Barros AP, Lang T (2003) Monitoring the monsoon in the Himalayas: Observations in central Nepal, June 2001. Mon Weather Rev 131: 1408–1427

Basu BK (2007) Diurnal variation in precipitation over India during the summer monsoon season: observed and model predicted. Mon Weather Rev 135:2155–2167

Bergman JW, Salby ML (1997) The role of cloud diurnal variations in the time-mean energy budget. J Clim 101114–1124

Bhatt BC, Nakamura K (2005) Characteristics of monsoon rainfall around the Himalayas revealed by TRMM precipitation radar. Mon Weather Rev 133:149–165

Bleeker W, Andre MJ (1951) On the diurnal variation of precipitation, particularly over Central U.S.A., and its relation to large scale orographic circulation systems. Quart J Roy Met Soc 77:260–271

Carbone RE, Conway JW, Crook NA, Moncrieff MW (1990) The generation and propagation of a nocturnal squall line: Part I. Observations and implications for mesoscale predictability. Mon Weather Rev 118:26–49

Chaboureau J-P, Guichard F, Redelsperger J-L, Lafore J-P (2004) The role of stability and moisture in the diurnal cycle of convection over land. Quart J Roy Meteor Soc 130:3105–3117

Chang ATC, Chiu LS, Yang G (1995) Diurnal cycle of oceanic precipitation from SSM/I data. Mon Weather Rev 123:3371–3380

Choudhury AD, Krishnan R (2011) Dynamical response of the South Asian monsoon trough to latent heating from stratiform and convective precipitation. J Atmos Sci 68:1347–1363

Dai A (2001) Global precipitation and thunderstorm frequencies: Part II. Diurnal variations. J Climate 14:1112–1128

Ebert EE, Janowiak JE, Kidd C (2007) Comparison of near-real-time precipitation estimates from satellite observations and numerical models. Bull Am Meteor Soc 88(1):47–64

ESRI, 2004. Using ArcGIS geostatistical analyst: ArcGIS 9. ESRI Press

Gadgil S, Joseph PV (2003) On breaks of the Indian monsoon. Proc Indian Acad Sci Earth Planet Sci 112:529–558

Gambheer AV, Bhat GS (2000) Life cycle characteristics of deep cloud systems over the Indian region using INSAT-1B pixel data. Mon Weather Rev 128:4071–4083

Gray WM, Jacobson RW Jr (1977) Diurnal variation of deep cumulus convection. Mon Weather Rev 105:1171–1188

Hann J (1901) Lehrbuch der Meteorologie, 1st edn. Leipzig Chr Herm Tauchnitz, pp 338–346

Houze RA Jr, Churchill DD (1987) Mesoscale organization and cloud microphysics in a Bay of Bengal depression. J Atmos Sci 44:1845–1867

Houze RA Jr, Wilton DC, Smull BF (2007) Monsoon convection in the Himalayan region as seen by the TRMM precipitation radar. Quart J Roy Meteor Soc 133:1389–1411. doi:10.1002/qj.106

Jeffrey SJ, Carter JO, Moodie KB, Beswick AR (2001) Using spatial interpolation to construct a comprehensive archive of Australian climate data. Environ Model Softw 16:309–330

Kaila VK, Kirankumar AS, Sundaramurthy TK, Ramakrishnan S, Prasad MYS, Desai PS, Jayaraman V, Manikiam B (2002) METSAT—a unique mission for weather and climate. Curr Sci 83:1081–1088

Kalnay E et al (1996) The NCEP/NCAR 40-Year Reanalysis Project. Bull Am Meteor Soc 77:437–471

Kistler R, Kalnay E, Collins W (2001) The NCEP-NCAR 50-year reanalysis: monthly means CD-ROM and documentation. Bull Am Meteor Soc 82:247–268

Kousky VE (1980) Diurnal rainfall variation in northeast Brazill. Mon Weather Rev 108:488–498

Landin MG, Bosart LF (1985) Diurnal variability of precipitation in the northeastern United States. Mon Weather Rev 113:989–1014

Mandke S, Sahai AK, Shinde MA, Joseph S, Chattopadhyay R (2007) Simulated changes in active/break spells during the Indian summer monsoon due to enhanced CO_2 concentrations: assessment from selected coupled atmosphere–ocean global climate models. Int J Climatol 27:837–859

Mapes BE, Warner TT, Xu M (2003a) Diurnal patterns of rainfall in northwestern South America: Part III. Diurnal gravity waves and nocturnal convection offshore. Mon Weather Rev 131:830–844

Mapes BE, Warner TT, Xu M, Negri AJ (2003b) Diurnal patterns of rainfall in northwestern South America: Part I. Observations and context. Mon Weather Rev 131:799–812

Medina S, Houze RA Jr, Kumar A, Niyogi D (2010) Summer monsoon convection in the Himalayan Region: terrain and land cover effects. Quart J Roy Meteor Soc 136:593–616

Meisner BN, Arkin PA (1987) Spatial and annual variations in the diurnal cycle of large-scale tropical convective cloudiness and precipitation. Mon Weather Rev 115:2009–2032

Mori S, Ichi HJ, Tauhid YI, Yamanaka MD, Okamoto N, Murata F, Sakurai N, Sribimawati T (2004) Diurnal land-sea rainfall peak migration over Sumatera island, Indonesian maritime continent, observed by TRMM satellite and intensive rawinsonde soundings. Mon Weather Rev 132:2021–2039. doi:10.1175/1520-0493(2004)132<2021:DLRPMO>2.0.CO;2

Negri AJ, Anagnostou EN, Adler RF (2000) A 10 yr climatology of Amazonian rainfall derived from passive microwave satellite observations. J Appl Meteorol 39:42–56

Nesbitt SW, Zipser EJ (2003) The diurnal cycle of rainfall and convective intensity according to three years of TRMM measurements. J Clim 16:1456–1475

Nesbitt SW, Cifelli R, Rutledge SA (2006) Storm morphology and rainfall characteristics of TRMM precipitation features. Mon Weather Rev 134:2702–2721

Perek L (1988) The scientific and technical aspects of the geostationary orbit. Acta Astron 17:587–598

Richards F, Arkin P (1981) On the relationship between satellite-observed cloud cover and precipitation. Mon Weather Rev 109:1081–1093

Romatschke U, Houze RA Jr (2011a) Characteristics of precipitating convective systems in the South Asian monsoon. J Hydromet 12:3–26

Romatschke U, Houze RA Jr (2011b) Characteristics of precipitating convective systems in the premonsoon season of South Asia. J Hydromet 12:157–180

Schumacher C, Houze RA Jr (2003) Stratiform rain in the tropics as seen by the TRMM precipitation radar. J Clim 16:1739–1756

Schwartz BE, Bosart LF (1979) The diurnal variability of Florida rainfall. Mon Weather Rev 107:1535–1545

Sen Roy S, Balling RC Jr (2007) Diurnal variations in summer season precipitation in India. Int J Climatol 27:969–976. doi:10.1002/joc

Sen Roy S, Sen Roy S (2010) Regional variability of convection over northern India during the pre-monsoon summer season. Theor Appl Climatol 103:145–158. doi:10.1007/s00704-010-0289-4

Sen Roy S, Saha SB, Fatima H, Roy Bhowmik SK, Kundu PK (2012) Evaluation of short-period rainfall estimates from Kalpana-1 Satellite using MET software. J Earth Sys Sci 121:1113–1123

Shin KS, North GR, Ahn YS, Arkin PA (1990) Time scales and variability of area-averaged tropical oceanic rainfall. Mon Weather Rev 118: 1507–1516

Tang M, Reiter ER (1984) Plateau monsoons of the Northern Hemisphere: a comparison between North America and Tibet. Mon Weather Rev 112:617–637

Tian B, Held IM, Lau NC, Soden BJ (2005) Diurnal cycle of summertime deep convection over North America: a satellite perspective. J Geophys Res 110, D08108. doi:10.1029/2004JD005275

Tyagi A, Hatwar HR, Pai DS (2009) Monsoon 2008: a report. IMD Met. Monograph: Synoptic Meteorology No. 7/2009
Tyagi A, Hatwar HR, Pai DS (2010) Monsoon 2009: a report. IMD Met. Monograph Synoptic Meteorology No. 9/2010
Wallace JM (1975) Diurnal variations in precipitation and thunderstorm frequency over the conterminous United States. Mon Weather Rev 103:406–419
Wang B, Kang IS, Lee JY (2004) Ensemble simulations of Asian–Australian monsoon variability by 11 AGCMs. J Clim 17:803–818
Wilson CA, Mitchell JFB (1986) Diurnal variation and cloud in a general circulation model. Quart J Roy Meteor Soc 112:347–369
Xie P, Arkin PA (1995) An intercomparison of gauge observations and satellite estimates of monthly precipitation. J Appl Meteorol 34:1143–1160
Yang G–Y, Slingo J (2001) The diurnal cycle in the tropics. Mon Wea Rev 129:784–801. doi:10.1175/1520-0493(2001)129<0784:TDCITT>2.0.CO;2
Yang S, Smith EA (2006) Mechanisms for diurnal variability of global tropical rainfall observed from TRMM. J Climate 19: 5190–5226
Ye D, Gao Y (1979) Meteorology of the Qinghai-Xizang Plateau, Chapter 1. Beijing, Science Press
Zhang GJ (2003) Roles of tropospheric and boundary layer forcing in the diurnal cycle of convection in the US southern great plains. Geophys Res Lett 30(24)
Zuidema P (2003) Convective clouds over the Bay of Bengal. Mon Weather Rev 131(5), 780–798