Intraseasonal Variation of the Summer Monsoon

I: Observational Aspects

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Abstract

Within the summer monsoon, the circulation and rainfall over the Indian region exhibit large variations over the synoptic scale of 3–7 days and the supersynoptic scales of 10 days and longer. In this paper we discuss some facets of intraseasonal variation on the supersynoptic scale on the basis of existing observational studies and some new analysis. The major variation of the summer monsoon rainfall on this scale is the active-break cycle. The deep convection over the Indian region on a typical day in the active phase is organized over thousands of kilometers in the zonal direction and is associated with a tropical convergence zone (TCZ). The intraseasonal variations on the supersynoptic scale are also coherent on these scales and are related to the space-time variation of the large-scale TCZ. The latitudinal distribution of the occurrence of the TCZ is bimodal with the primary mode over the heated continent and a secondary mode over the ocean. The variation of the continental TCZ is generally out of phase with that of the oceanic TCZ. During the active spells, the TCZ persists over the continent in the monsoon zone. The revival from breaks occurs either by northward propagation of the TCZ over the equatorial Indian Ocean or by genesis of a disturbance in the monsoon zone (often as a result of westward propagations from W. Pacific). The mechanisms governing the fluctuation between active spells and breaks, the interphase transition and the complex interactions of the TCZ over the Indian subcontinent with the TCZ over the equatorial Indian Ocean and the W. Pacific, have yet to be completely understood.

1. Introduction

The rich structure of the subseasonal variation of the summer monsoon has been elucidated by several studies involving analysis of conventional and satellite data. In this paper we do not attempt a comprehensive review of these studies. Rather, we briefly describe some facets of the intraseasonal variation on the supersynoptic scale, identified therein. In part II, we consider model simulations of the variation over the Indian region and discuss the underlying mechanisms.

The monsoon is a manifestation of the seasonal migration of the equatorial trough in response to the seasonal variation of the insolation. Over the Indian subcontinent, in the northern summer, the heat low over the desert of the northwestern part of the near-equatorial trough together comprise the low pressure belt at the surface (Ramage, 1971, p74). The onset phase of the monsoon commences in late May/early June. At the culmination of the onset phase, towards the beginning of July, the near-equatorial trough gets established near its seasonal mean location around 20°N over the heated sub-

continent. During the peak monsoon months, the trough fluctuates in the monsoon zone north of 15°N (Sikka and Gadgil, 1980 and references therein). On the intraseasonal scales, there are also marked variations in the intensity of the shear between the monsoon westerlies to the south of the trough and the easterlies to its north. The variations in the large-scale monsoon rainfall over the monsoon zone occur in association with the variation in the intensity of the trough. Perhaps, the most important feature of the supersynoptic scale variation of the summer monsoon over India is the fluctuation between active spells and the so-called breaks in the monsoon. We consider first the structure of these two extreme phases of the monsoon and the relevance of this subseasonal variation for the interannual variation of the monsoon.

2. Active and break monsoons

Active spells of the summer monsoon over the Indian region are associated with an intense trough with heavy rainfall over the monsoon zone. The circulation at different levels on a typical day in the active phase of the monsoon is shown in Fig. 1. The visible satellite imagery of that day is shown in Fig.

Fig. 1. Circulation pattern at different levels on 8 July 1973.

2 and the rainfall received in the corresponding 48 hours period is shown in Fig. 3. It is seen that in the active phase, the trough is characterized by high cyclonic vorticity above the boundary layer and intense cloudiness and precipitation. M. Murakami (1976) has shown that the temperature and specific humidity profiles over the eastern part of the trough is similar to that for the ITCZ described by Jordan (1958). Thus the trough has the dynamical and the thermodynamical characteristics of the ITCZ (Charney, 1969, Sikka and Gadgil, 1980). Since the convergence in each such system need not be intertropical, we refer to it as a tropical convergence zone (TCZ) henceforth (Gadgil, 1988).

Nearly the opposite situation occurs at the weakest phase of the monsoon during the so-called break. Ramamurthy's (1969) exhaustive survey of the breaks in the monsoon rainfall during 1888–1967 has provided valuable information on the nature of the circulation and rainfall characterising this important phase of the summer monsoon. He defined the break situation as one in which 'the trough of

low pressure was not seen in the surface map and the easterlies present to the north of the trough were practically absent in the lower tropospheric levels.' During the break phase, there is a marked change in the lower tropospheric wind over the monsoon zone (Fig. 4) with the vorticity above the boundary layer becoming anticyclonic (Sikka and Gadgil, 1978). The large-scale rainfall over the monsoon zone is directly related to the meridional shear of the zonal wind just above the boundary layer (Sikka and Gadgil, 1978), and when this vorticity is anticyclonic, there is hardly any rainfall over most of the monsoon zone. During long intense breaks, the surface temperature rapidly increases and a heat low type circulation gets established over the monsoon zone with subsidence over most of the troposphere and westerlies restricted to the lower levels (Raghavan, 1973). The cloud band delineating the TCZ as well as the prominent trough at 700 mb characterizing an active spell disappears (Sikka and Gadgil, 1980).

The phenomenon of breaks in the monsoon has

Fig. 2. NOAA visible satellite imagery on 8 July 1973. a) Northern hemisphere (above) b) Indian region (below).

received considerable attention because the major difference between the so-called good monsoon seasons (with above normal well distributed rainfall over the monsoon zone) and the poor monsoon seasons (with deficient rainfall), is in the duration and intensity of breaks (Krishnamurti and Bhalme, 1976, Sikka, 1980). Sikka (1980) has shown that the average number of break monsoon days per year during July–August is 13.8 days in poor monsoon years (1965, 66, 68, 72, 74) and 3.2 days in good monsoon years (1961, 67, 73, 75, 76).

So far we have considered the variation of the large-scale fields of convection and circulation. However it is important to note that the intensity of the TCZ is not uniform across the zone, being much higher in the synoptic scale vortices embedded in the trough (*e.g.* Fig. 2). There is a strong link between the planetary scale TCZ and the synoptic scale disturbances, with the variation in the intensity and location of the TCZ being generally associated with the genesis and propagation of such monsoon disturbances. Active spells are characterized by a sequence of time-clustering partly overlapping development of monsoon disturbances and wide-spread rainfall (T. Murakami, 1976). On the other hand

Fig. 3. Accumulated daily rainfall (in cms) for 24 hours ending at a) 003Z 8 July 1973 (above) b) 003Z 9 July 1973 (below).

hardly any disturbances are generated in the breaks. A vast majority (about 80 %) of the monsoon disturbances are generated over the head Bay and move westward in the monsoon zone along the monsoon trough (Sikka, 1977). The disturbances range in intensity from a low to a monsoon depression and finally a tropical cyclone. Monsoon depressions have been the subject of several investigations. However, Sikka (1980) has shown that the major difference between the good monsoon years and poor monsoon years is in the frequency of genesis of the monsoon lows with little variation in the number of the monsoon depressions between these two types of years.

There are two distinct modes of the transition from a break to an active phase. In the first, revival of the monsoon from a break occurs with formation of a disturbance in the northern parts of the Bay of Bengal and its subsequent westward movement over the monsoon zone. Krishnamurti *et al.* (1977) showed that the genesis of a disturbance over the Bay often occurs as a result of westward propagation of a disturbance from the eastern regions extending

Fig. 4. Mean upper wind flow pattern during active and break monsoon (from Ramamurthy, 1969).

from Burma to the W. Pacific (Fig. 5). The second mode of revival involves northward movement of the oceanic TCZ onto the subcontinent (*e.g.* Fig. 9). Periodic northward movements of the oceanic TCZ is a feature seen in the variation of the TCZ over the Indian longitudes. Sikka and Dixit (1972) and Sikka and Gadgil (1980) found that the revival by the second mode occurs on a little more than half the occasions.

It is important to note that although the active and weak phases of the summer monsoon have been traditionally identified on the basis of the fluctuation in the rainfall in the monsoon zone, other rainfall indices have also been used. However, the variations of the rainfall over the entire Indian region are not

coherent. Thus during breaks, while there are very large negative anomalies in the rainfall over the monsoon zone, positive anomalies occur over the Himalayan foothills, the northeastern part of India and over the eastern part of the peninsula (Fig. 6, from Ramamurthy, 1969). The rainfall over the west-coast shows even more marked fluctuations than that over central India since the quantum of the rainfall is larger and the variations along the coast coherent (Fig. 7). In T. Murakami's (1976) analysis of intraseasonal variation, the active-spells are identified on the basis of the west coast rainfall. M. Murakami (1976) has shown that in active phases over the monsoon zone, the rainfall over the west coast is also high because of the high intensity of the

Fig. 5. Hovmoller diagram based on global sea level pressure (mb) data at 20°N during July 1965 for the sum of zonal wavenumbers 3–12 (from Krishnamurti *et al.*, 1977).

Fig. 6. Mean departure (%) or rainfall during breaks (from Ramamurthy, 1969).

monsoon westerlies. However, the weak phases in the monsoon zone coincide with weak phases in the west coast on a majority (70 %) but not on all occasions (Ramamurthy, 1969). In particular when the revival of the monsoon occurs through northward movement of the oceanic TCZ (Sikka and Gadgil, 1980), high rainfall occurs over the west coast during the weak phase in the monsoon zone (*e.g.* last week of July in Fig. 7). Hartmann and Michelsen's (1989) detailed analysis of the structure of the 40 day mode also shows that the active spell over central India lags that over the west coast by about 5 days. Since the rainfall variations over the eastern peninsula are out of phase with those of the monsoon zone, the active/weak spells identified by analysis of the variation of the average rainfall of stations dis-

Fig. 7. Variation of rainfall in cms during the summer monsoon of 1979 for the regions of a) Central India (top) b) West Coast (bottom).

tributed throughout the country (Cadet and Daniel, 1988) will also be different from those derived from the traditional definition.

3. Temporal scales of the oscillations of the monsoon

The earlier studies of the break monsoon phenomenon (*e.g.* Dixit and Jones, 1965, Ramamurthy, 1969, *etc.*) focussed on the structure of the associated circulation and rainfall patterns and contrasting these with the opposite phase of the monsoon viz. the active monsoon phase. As the term 'break' implies, the break condition was considered to be associated with a complete destruction of the monsoon flow pattern. The duration of the breaks was found to vary from 2–3 days to as long as 2–3 weeks. The problem of identification of the important time-scales of the intraseasonal variation of the monsoon began to be addressed in the 70s. Ananthakrishnan and Keshavamurthy's (1970) spectral analysis of the data on wind, rainfall and surface pressure at different stations in India during 1965, 67 showed peaks around 4–7 days for the summer season and around 15 days as well as 4–5 days in the winter. T. Murakami (1972) pointed out that the rainfall over the West coast of India varies on a time-scale of about 15 days during the summer monsoon. Krishnamurti *et al.*'s (1973) spectral analysis of the oscillation of the intensity of the Tibetan high revealed a major peak around 12 days and another one around 3 days. Several subsequent studies revealed periodicities around 5 days and 15 days for the variation in the summer (Sikka and Mishra, 1974, Bhalme and Parashnis, 1975 *etc.*). The structure of this quasibiweekly oscillation in all the important fields such as pressure, wind, static stability rainfall over the central part of India *etc.* and the relation to features such as

the Mascarene High and the cross-equatorial jet was elucidated by Krishnamurti and Bhalme (1976). M. Murakami's (1976) spectral analysis of the summer monsoon fluctuations over India using the data of 1962 showed clear peaks around the synoptic scale of 5 days and the supersynoptic scale of about 15 days. He showed that the 15 days mode represented an oscillation of the zonal wind at the eastern end of the monsoon trough at Calcutta which in turn were related to the active-weak spells of the rainfall on the West Coast. He also pointed out that it may be more appropriate to view the break monsoon as one stage (albeit more intense!) of the active/weak cycle which occurs in almost every season, rather than a destruction of the monsoon flow. M. Murakami (1977) has reviewed the earlier work on the 15 day mode. This mode has been shown to be a westward propagating mode by Krishnamurti and Ardanuy (1980). From an analysis of the breaks over a 40 year period they also showed that on a majority of cases the break commenced when a westward propagating ridge arrived over central India. The in-situ revival of the monsoon from breaks by genesis of disturbances within the monsoon zone (Sikka and Gadgil, 1980) also appears to be associated with this westward propagating mode.

When more sophisticated methods capable of detecting periodicities of $1/2$ or $1/3$ the data length came into vogue, the Madden-Julian 30–50 day mode over the Asian monsoon zone was discovered. Yasunari (1979) found periodicities around 15 and 40 days in the cloudiness fluctuations during June–September 1973. The 40-day mode was found to involve northward propagations (Yasunari, 1979, 1980). The structure of this mode is described in detail in the review by M. Murakami (1987). This mode is clearly associated with the revival of the continental TCZ by northward propagation of the oceanic TCZ about one month after the culmination of the onset phase (Sikka and Gadgil, 1980). In the last decade, there has been a large number of studies of the structure of this 30–50 day mode (*e.g.* Lau and Chan, 1986, Murakami *et al.*, 1984, Murakami and Nakazawa, 1985), and attempts have been made to predict it (Cadet and Daniel, 1988).

Hartmann and Michelsen's (1989) in-depth analysis of the daily rainfall at 3700 stations in the Indian region during the summer half years in 1901–1970 revealed only two statistically significant peaks—one for the synoptic time-scale of 5–7 days and another at the Madden-Julian time-scale of 30–50 days over most of the Indian region south of 23°N . Only a part of the region in the far north of India along the southern flank of the Himalayas exhibit spectral peaks near 15 days. Significant variance at periods of about two weeks was found at a few locations around India, but where the 40–50 day oscillation is strong the two week variance was found to blend into

the red background spectrum. Therefore, over much of India, a distinct two week oscillation could not be identified. The discrepancy between these and earlier results arises because of the presence of a large variation in the variance in the different frequencies from year to year (Hartmann and Michelsen, 1989). Thus although the maximum variance associated with the biweekly mode and 40 day modes in the region $21\text{--}22^\circ\text{N}$ and $72\text{--}73^\circ\text{E}$ are about 50 % and 70 %, the average variance in these modes is only about 7 % and 12 % respectively.

The relationship of the active-break cycle to the amplitude of these modes and their relative phases in a given year is rather complex. For example, in the summer of 1979, even though the 40 day mode was prominent (Kondragunta, 1990, Krishnamurti and Subramanian, 1982), the long breaks resulted in a poor monsoon. Yasunari's (1981) analysis geopotential height at 850 mb during the monsoon of 1969 showed that all the monsoon depressions occurred in the negative pressure anomaly phases of the bi-weekly as well as the 40 day mode. If the frequency of genesis of the monsoon lows also increases significantly in such phases, we expect the duration and intensity of the active spells (and hence the total monsoon rainfall during the summer) to depend on the power in each of these modes as well as the relative phases.

4. Spatial scales

We have seen that the important time-scales for the intraseasonal variation of the summer monsoons are the synoptic scale of 3–7 days and the supersynoptic scales of 10–20 days and 30–50 days. The dynamical system associated with the 3–7 day scale is the monsoon disturbance. We have seen that on a typical day in the active phase (Figs. 1–3) these synoptic scale disturbances are organized into a larger-scale system. In the seasonal mean picture also the region with low values of outgoing long-wave radiation stretches across at least 50° in longitude (Janowiak *et al.*, 1985). Are the intraseasonal variations on time scales longer than 10 days, also characterized by large spatial-scales?

This problem was first addressed by T. Murakami (1976). He analysed the daily data on satellite-derived cloudiness during 1 June to 15 September 1967–72. In order to remove the synoptic scale variation, 5-day running means were taken and correlations related to a point in Central India were calculated using a grid with 30° in longitude and 5° in latitude (Fig. 8a). A rather coarse grid was taken because it was found that for finer grids the cross-correlation exhibited oscillations between neighbouring grids. It is seen that the variations in cloudiness are in phase with Central India for a region over 90° in longitude, extending southeastward upto the equatorial Western Pacific. The variations

Fig. 8a. Latitude-longitude distribution of correlation coefficients between cloudiness changes at a reference point (80°E, 15°N) and those at every 5°x30° grid during the period from June 1–September 15 for 8 years from 1968–1972. Hatching indicates regions with correlation coefficients >0.2 and dashed hatching indicates regions with correlation coefficients <-0.2 (from T. Murakami, 1976).

Fig. 8b. Spatial distribution of correlation coefficients of cloudiness with a reference point over Central India (17.5°N, 78°E). The values greater than 0.4 are shaded and those less than -0.4 are dotted (from Yasunari, 1979).

over the equatorial Indian Ocean are seen to be out of phase relative to those over Central India. This is consistent with see-saw in the activity of the continental and oceanic TCZs discussed in Section 3.

Yasunari's (1979) computation of the cross-correlations relative to 72–86°E, 15–20°N using the satellite-derived cloudiness data for June–September 1973 with a grid of 12° in longitude and 5° in latitude is shown (Fig. 8b). The region over which cloudiness fluctuations vary coherently is found to be a zonal belt extending from 60°E to about 140°E. As in Fig. 8a, a zone of negative correlation appears over the equatorial Indian Ocean. The absence of the southeastward tilt of the coherent zone found by T. Murakami (1976) may be due to difference in the quality of the data over the two periods.

We computed cross-correlations relative to 80°E, 17.5°N using 5-day running mean of the 2.5° daily data of OLR for the summer monsoon (June–September) during 1974–77 (Fig. 8c). With the finer grid of this standard data set also, the region over

which fluctuations are in phase with the Indian region is a zonal belt extending over about 50° in longitude and 20° in latitude as suggested by Yasunari's (1979) analysis. The out-of-phase oscillation between the continental and oceanic convection indicated by a region of negative correlation over the equatorial Indian Ocean is also seen. The correlation relative to 0°, 80°E (Fig. 8d) also shows spatial scales of about 50° in longitude and 20° in latitude and is similar to the first EOF in Lau and Chan's (1986) analysis of the 40–50 day oscillation in the northern summer based on OLR data. It is seen that the variation over the equatorial Indian ocean is negatively correlated with that over the equatorial West Pacific.

5. Intraseasonal variation of the TCZ

We have seen that the deep convection associated with the summer monsoon is characterized by large zonal scales on the daily, intraseasonal (supersynoptic) as well as seasonal scales. The dynamical system associated with this large-scale convection/rainfall

Fig. 8c. Spatial distribution of correlation coefficients (X10) of the 5-day running mean OLR for July–August during 1974–77. Contours of correlation are marked at values $-1, 0, 2, 4, 6, 8$.

Fig. 8d. Same as Fig. 8c except relative to $(0^\circ, 80^\circ\text{E})$.

is the tropical convergence zone, which appears as a prominent cloud band in the satellite imagery (*e.g.* Fig. 2). Sikka and Gadgil's (1980) detailed analysis of satellite imagery has revealed many features of the seasonal transitions and intraseasonal variation of the TCZ over the Indian longitudes. The variation of the TCZ at the central longitude of 80°E during the monsoon seasons 1975, 1979 is shown in Fig. 9. The basic features of the seasonal transitions and intraseasonal variation which occur every summer are seen to be:

- (i) Northward movement of the TCZ from the equatorial Indian Ocean onto the continent during the onset phase of the monsoon in late May or June; sometimes there is a multiple onset involving 2–3 propagations in rapid succession (*e.g.* 1975).
- (ii) Persistence of the TCZ in the monsoon zone for 10–20 days after the establishment of the continental TCZ at the culmination of the onset phase (*e.g.* 5–25 July 1975, 7–17 July 1979 in Fig. 9).
- (iii) Disappearance of the TCZ from the monsoon zone for periods ranging from few days to few weeks.
- (iv) revival of the TCZ from phase (iii) either by genesis within the monsoon zone or by northward

Fig. 9. Variation of the maximum cloudiness zone (Sikka and Gadgil, 1980) at 80°E during the summer monsoons of 1975 and 1979.

propagation of the oceanic TCZ. This is followed by another cycle of events (ii) to (iv).

The variation of the daily rainfall over central India and the fractional area covered by the TCZ at 80°E between $15^\circ\text{--}25^\circ\text{N}$, for the summer of 1979 are depicted in Fig. 10. It is seen that the persistence phase (ii) above (*e.g.* 7–14 July, 27 July–9 August

Fig. 10. Variation of the daily rainfall over Central India for 1979 (above). Fractional area covered by the TCZ at 80°E (below).

1979) correspond to the active spells and the disappearance phase (iii) (e.g. 15–25 July 1979) above to the breaks.

One prominent feature of the intraseasonal variation of the TCZ over the Indian longitudes is the repeated northward propagations of the oceanic TCZ onto the continent at a rate of about 1° per day (Gadgil and Srinivasan, 1990). T. Murakami's (1976) time-lag correlation analysis of satellite data had revealed northward propagation of cloudiness at the same rate. Murakami *et al.* (1984) and M. Murakami (1987) also found propagations at a similar rate. The phenomenon of repeated poleward propagations is a basic feature of the variation of the TCZ which is observed year after year only over this part of the tropics (Gadgil and Srinivasan, 1990). The period between successive propagations varies from 2 to 6 weeks (Sikka and Gadgil, 1980, Webster, 1983).

The total rainfall received in the summer monsoon season depends upon the duration and intensity of the active spells/breaks. Hence the major difference between the good monsoon season of 1975 and the poor one of 1979 is the longer active spells (breaks) in the former (latter). For example the active spell commencing at the end of the onset phase in 1975 lasted for about 20 days while that in 1979 for only a week. The deficient rainfall in the MONEX season of 1979 was a consequence of the long breaks in July, August.

Fig. 11. Frequency of occurrence of the TCZ at 80°E during June, July, August (from Sikka and Gadgil, 1980).

It is seen from Fig. 9 that, even after the establishment of the continental TCZ over the monsoon zone during the onset phase towards end of June, a TCZ appears intermittently over the warm waters of the equatorial Indian Ocean. Thus the distribution of the occurrence of the TCZ is bimodal with the primary mode over the ocean in June and the primary mode over the monsoon zone during the peak monsoon months of July–August (Fig. 11). Sikka and Gadgil (1980) observed that there is a see-saw in the activity of the continental and the oceanic TCZs with active spells of one generally coinciding with weak spells of the other. M. Murakami (1984) also found that in the break monsoon period in the summer of 1979 there was enhanced convection over the equatorial ocean. This is consistent with the negative region of correlation vis-a-vis 80°E, 17.5°N (Figs. 8b, 8c). We expect such a competition between two TCZs over the same longitudes. This is because the air ascending in each TCZ descends in the surrounding latitudinal belt (which is therefore cloudfree) and this subsidence prevents convection in the vicinity of the TCZ. There is considerable interannual variation in the duration of the active spells of the oceanic TCZ. In poor monsoon seasons

with shorter active spells of the continental TCZ relative to the normal, the oceanic TCZ is characterized by relatively longer spells. Yet the oceanic TCZ plays an important role in maintaining the continental TCZ by northward propagations.

The TCZ over the Indian longitudes also has an interesting relationship with the TCZ over the West Pacific. We have seen that revival of the former often occurs by westward propagation of the disturbances from the latter. Yet, there is competition between the TCZ over the Indian longitudes and that over the West Pacific as suggested by (Fig. 8d), and also by Joseph (1990). Thus eastward and equatorward of the continental TCZ over the Indian region are systems which contribute to its maintenance by propagation onto the region; but also compete with it and can suppress its activity. Understanding the nature of interactions between these systems is one of the central problems of intraseasonal variation.

We consider the simulation of the important features of the observed variation of the TCZ over the Indian longitudes viz. the northward propagations, active-weak spells and bimodality of occurrence with a see-saw between the oceanic and continental TCZs, in part II.

6. Concluding remarks

The studies over the last two decades have given considerable insight into the nature of the supersynoptic scale variation of the monsoon over India during summer. The active-break cycle has been shown to be an intrinsic feature of the variation of the planetary scale monsoon on this scale. The role of the repeated northward propagations from the equatorial Indian Ocean and the westward propagations from the W. Pacific, in revival from breaks has been recognized. The variation of the Indian monsoon has been shown to be related to that of the convection over the equatorial Indian Ocean and the W. Pacific. The physical mechanisms leading to the observed variation of the summer monsoon over India and its links with convection over the other parts of tropics are yet to be completely understood. The theoretical investigations of a few facets of this problem are considered in Part II.

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夏のモンスーンの季節内変動 第1部：観測的側面

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夏のモンスーン期におけるインドの大気循環と降雨活動は、3–7日スケールの総観規模、10日スケール及びより長い時間スケールの超総観規模で大きな変動を示す。この論文では超総観規模の季節内変動の側面について、既存の解析研究及び新たな解析に基づいて議論する。このスケールでの夏のモンスーン降雨の主要な変動は活発・休止のサイクルである。活発期におけるインド地域での深い積雲対流活動は東西方向に数千kmにわたって組織化されており、熱帯収束帯(TCZ)に伴っている。超総観規模の季節内変動もこのスケールと一体となっており、大規模なTCZの時空間変動と関連している。TCZの緯度分布は、主要なピークが熱大陸上、二次的なピークが海洋上に現れるバイモーダルな特徴を示す。大陸上のTCZの変動は、海洋上の変動と通常は逆位相の関係にある。活発期におけるTCZは大陸上のモンスーン域に位置している。休止期から活発期への移行は、TCZの赤道インド洋からの北進か、またはモンスーン域における擾乱の発生(しばしば西太平洋からの西進の結果として)に伴っている。活発・休止の変動を支配する機構、位相の移行、インド亜大陸上のTCZと赤道インド洋・西太平洋上のそれとの複雑な相互作用などは未だ十分に理解されていない。