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THE INDIAN MONSOON AND ITS VARIABILITY

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■ **Abstract** For over 300 years, the monsoon has been viewed as a gigantic land-sea breeze. It is shown in this paper that satellite and conventional observations support an alternative hypothesis, which considers the monsoon as a manifestation of seasonal migration of the intertropical convergence zone (ITCZ). With the focus on the Indian monsoon, the mean seasonal pattern is described, and why it is difficult to simulate it is discussed. Some facets of the intraseasonal variation, such as active-weak cycles; break monsoon; and a special feature of intraseasonal variation over the region, namely, poleward propagations of the ITCZ at intervals of 2–6 weeks, are considered. Vertical moist stability is shown to be a key parameter in the variation of monthly convection over ocean and land as well as poleward propagations. Special features of the Bay of Bengal and the monsoon brought out by observations during a national observational experiment in 1999 are briefly described.

1. INTRODUCTION

The word “monsoon” is derived from the Arabic word for season, and the distinguishing attribute of the monsoonal regions of the world is considered to be the seasonal reversal in the direction of the wind. The monsoonal region delineated on the basis of significant change in the wind direction between winter and summer (with the direction of the prevailing wind within each season being reasonably steady) extends over a large part of the tropics, namely, 25°S to 35°N, 30°W to 170°E (Ramage 1971).

Near the center of this monsoonal region is the Indian subcontinent and the ocean surrounding it, which experiences large seasonal variation in wind direction (Figure 1). In this paper, I focus on the monsoon over this region, specifically, the Indian monsoon; elucidate some facets of the mean seasonal patterns and the variability; endeavour to take stock of the present understanding of the underlying mechanisms; and identify the challenging problems that need to be addressed for further insight. I do not, however, attempt a comprehensive review, such as by Webster et al. (1998), but discuss only a few aspects that I consider important.

For the millions inhabiting monsoonal regions, the seasonal variation of the rainfall associated with the monsoon system is of far greater importance than the seasonal variation of wind, which was critical knowledge for Arab sailors. The space-time variation of rainfall has such a large impact on the resources of the region that it has been said that the Indian economy is a gamble on the monsoon rains. Over a large part of the Indian region, most of the rainfall occurs in the months of June to September during the summer monsoon (or the so-called southwest monsoon, named after the direction of prevailing surface wind). The exception is the east coast of the southern peninsula, where most of the rainfall occurs during October–November. The spatial variation of the mean June–September rainfall over the Indian region is shown in Figure 2.

A useful index of the summer monsoon rainfall over the Indian region in any year is the all-India summer monsoon rainfall (ISMR), which is a weighted average

(*a*)

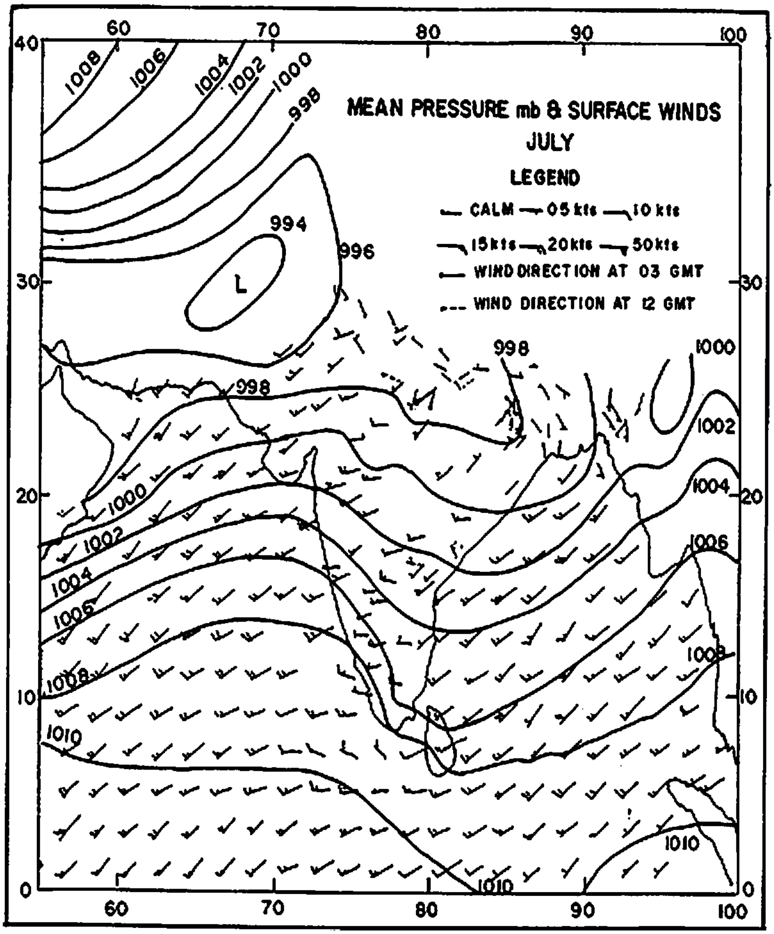
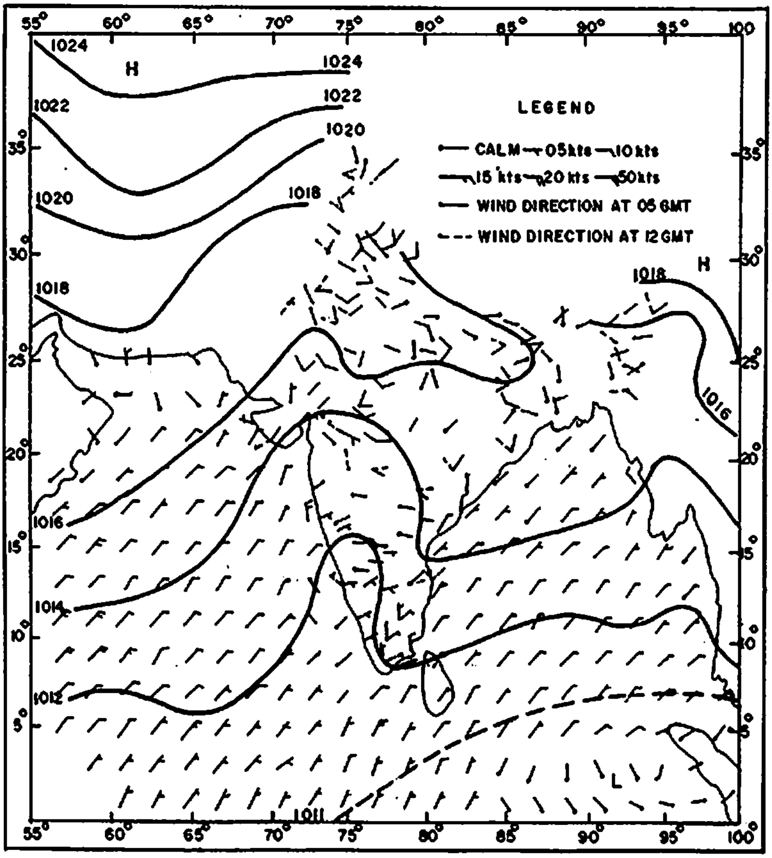


Figure 1 Mean surface winds and pressure for July (*a*) and January (*b*) (after Rao 1976).

(*b*)



MEAN PRESSURE (mb) AND SURFACE WINDS JANUARY

Figure 1 (*Continued*)

of the June–September rainfall at 306 well-distributed rain guage stations across India (Parthasarathy et al. 1992, 1995). The variation of ISMR during the last century is shown in Figure 3*a*. The interannual variation is not very large, with the standard deviation being only about 10% of the mean. Yet it has a very large impact on agricultural production (Figure 3*b*). Note that over the past decade, the growth rates of the production have decreased in association with fatigue of the green revolution (Gadgil et al. 1999); hence, an even larger impact is expected in the future. As shown in Figure 3, there is a close correspondence between deficit monsoon rainfall and El Niño (Sikka 1980, Pant & Parthasarathy 1981, Rasmusson & Carpenter 1983). However, Kumar et al. (1999) have suggested that the link with El Niño has weakened in the last decade, and in fact the ISMR anomaly was positive in the recent intense warm event of 1997. Mechanisms leading to the interannual

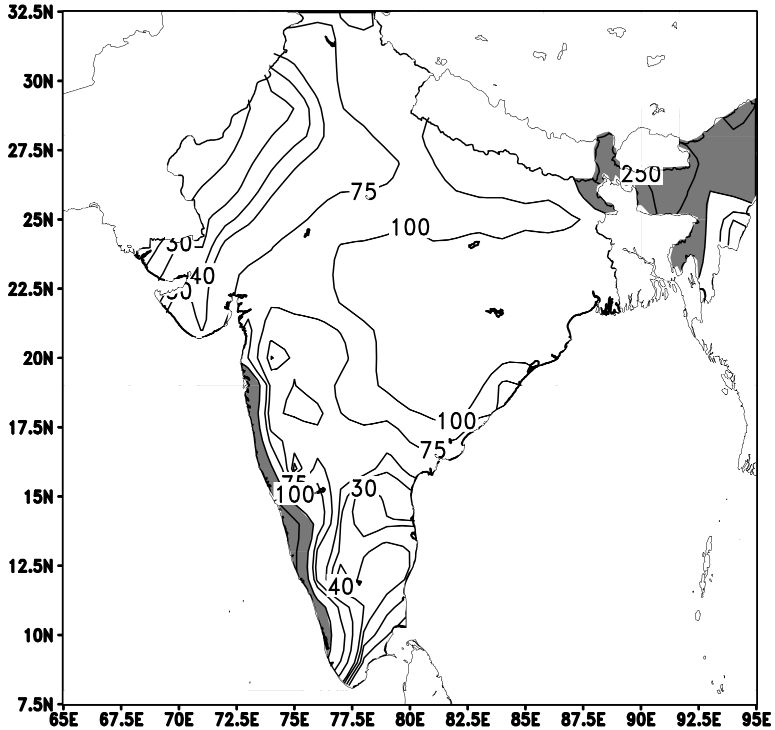


Figure 2 Mean June–September rainfall over the Indian region south of 30° N.

variation of ISMR, including teleconnections with other phenomena (such as the El Niño) are yet to be unravelled.

During the summer monsoon, high rainfall occurs along the West Coast of the peninsula (associated with the orography parallel to the coast) and over the northeastern regions. In addition, there is a broad zone around 20°N, stretching northwestward from the head of the Bay of Bengal, which receives significant rainfall (Figure 4*a*). In fact, the variations of all-India summer monsoon rainfall are highly correlated with the variation of the rainfall over this zone (Figure 4*b*), which has been called the monsoon zone (Sikka & Gadgil 1980). Hence, the central problem is to identify the system associated with the rainfall over the monsoon zone, which tends to be organized on synoptic and supersynoptic scales.

There is considerable variation of rainfall over subseasonal scales between active and weak spells, which Blandford (1886) described as the fluctuation between spells "during height of rains" and "intervals of droughts" (e.g., Figure 5 for daily variation of rainfall over central India during 1972 and 1975). The major difference between the rainfall variation in the good monsoon season of 1975 and the poor

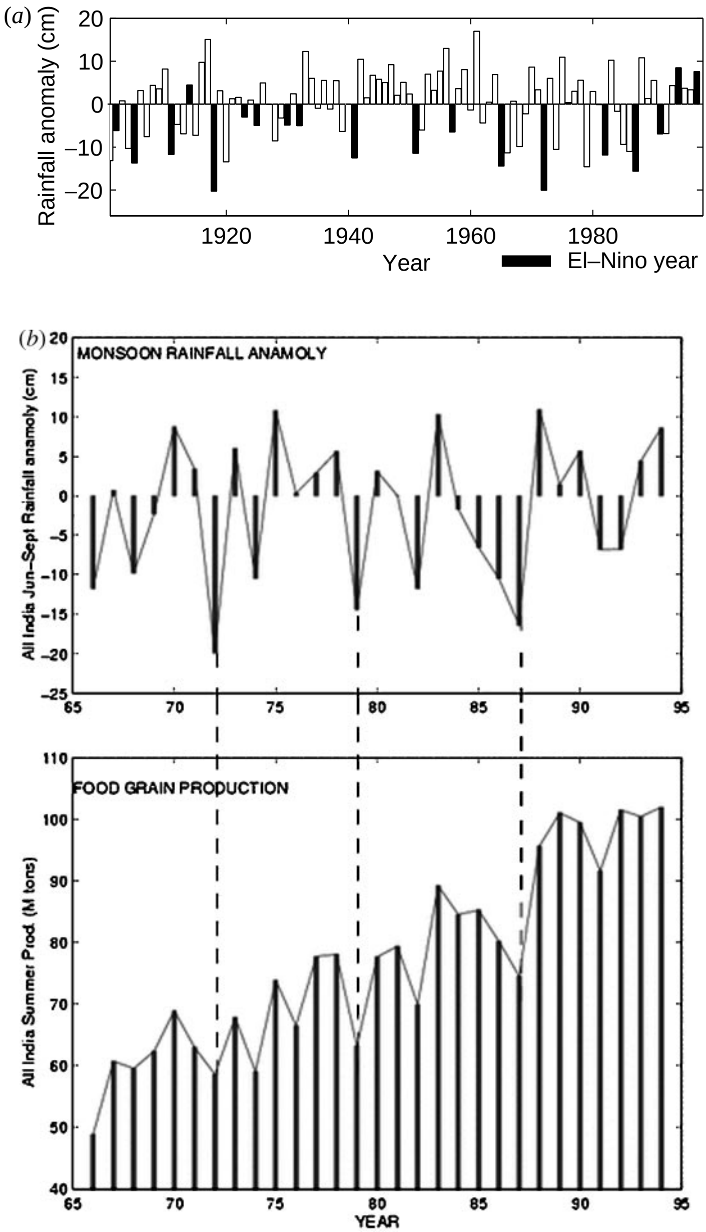


Figure 3 (*a*) Interannual variation of the all-India summer monsoon rainfall (ISMR) during 1901–1998; the El Niño years are shaded. (*b*) Variation of ISMR anomaly (*top*) and the Indian summer foodgrain production.

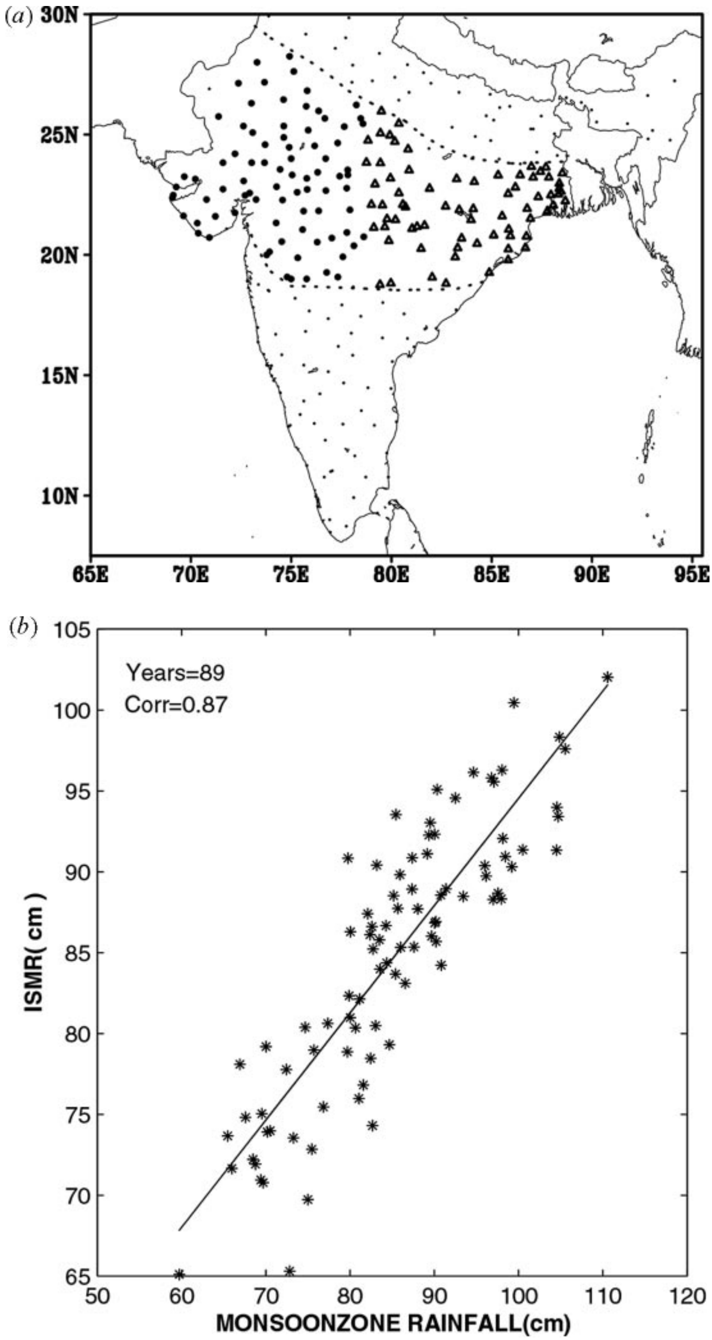


Figure 4 (*a*) The Indian monsoon zone (limits indicated by *dashed lines*) and the network of stations. Stations in the western and eastern parts of the monsoon zone indicated by circles and triangles. (*b*) ISMR versus June–September rainfall over the monsoon zone; sample size (89 years) and correlation (0.87) are indicated.

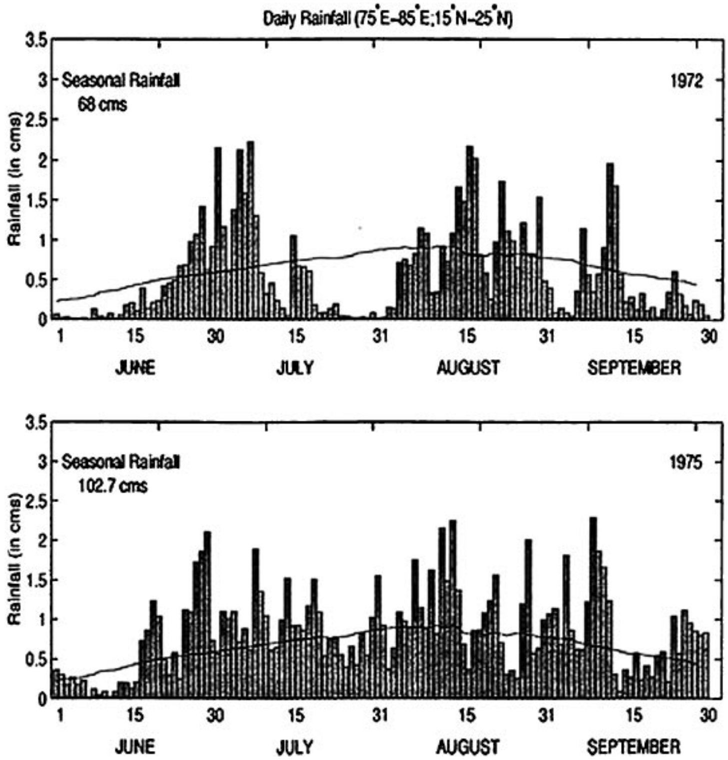


Figure 5 Variation of daily rainfall over central India during 1972 and 1975.

monsoon season of 1972 is the occurrence of a long dry spell or "break" in 1972. There has been considerable work on documenting the nature of intraseasonal and interannual variation of the monsoon and the interrelationship [e.g., Ramamurthy 1969, Krishnamurti & Bhalme 1976, Sikka & Gadgil 1980, Ferranti et al. 1997, Goswami et al. 1998, Lawrence & Webster 2001, Sperber et al. 2000]. However, processes that lead to the fluctuation between active and weak spells or breaks, and in particular, factors that trigger the transition between the two, are yet to be adequately understood.

It is important to note that most of the rainfall over the Indian region during the summer monsoon occurs in association with synoptic-scale convective systems generated over the warm seas/ocean surrounding the subcontinent that propagate onto the subcontinent. Propagation of systems from the Bay of Bengal along the monsoon zone contributes substantially to the rainfall during the summer monsoon (e.g., Figure 6 for the season of 1999). Active spells of the monsoon are characterized by a sequence of time-clustering partly overlapping development of such disturbances (Murakami 1976), whereas no such systems occur over the

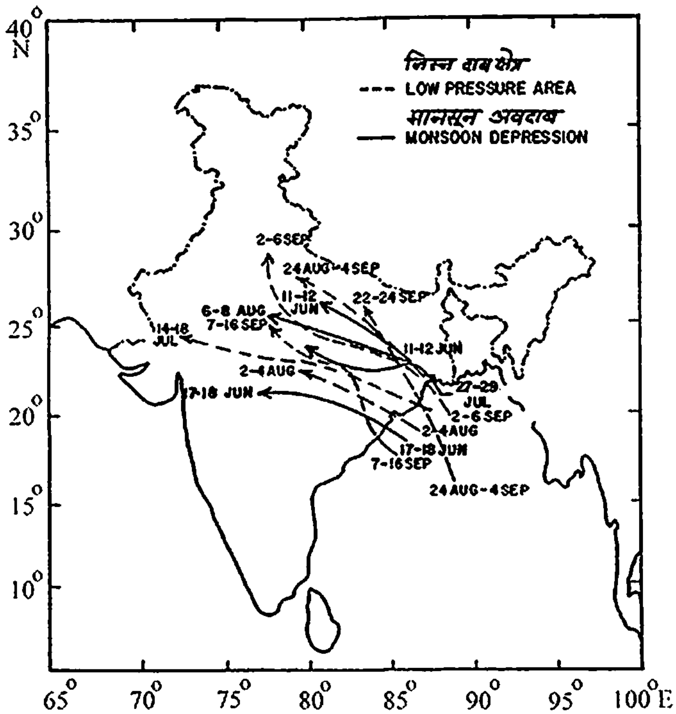


Figure 6 Paths of low pressure systems during the summer monsoon season of 1999.

monsoon zone during breaks. The spatial variation of the summer monsoon rainfall over the Indian region is linked to that of the convection over the Bay (Gadgil 2000). For example, during the summer monsoon of 1998, the convection over the northern Bay was anomalously low, with occurrence of relatively few systems, whereas a large number of systems formed over the southern parts of the Bay, implying positive (negative) anomalies of the outgoing longwave radiation (OLR) over the northern (southern) Bay. Most of these systems moved from the southern Bay onto and across the southern peninsula, and the rainfall anomalies were positive (negative) over southern peninsula (monsoon zone). One of the most challenging problems is understanding the links of monsoon variability to the convection over the surrounding ocean as well as the Pacific.

In this paper, I consider first the theories proposed for the basic system of the monsoon and discuss what can be deduced from the available observations about it. Intraseasonal variation is discussed in Section 3, and variation of convection over the oceans in Section 4. In Section 5, what we can surmise about the mechanisms associated with variability is summarized, and recent developments are briefly described in Section 6.

2. MONSOON—THE BASIC SYSTEM

Over 300 years ago, Halley (1686) suggested that the primary cause of the monsoon was the differential heating between ocean and land. Thus, in this first-ever model proposed, the monsoon was considered to be a gigantic land-sea breeze. Differential heating of land and sea is still considered as the basic mechanism for the monsoon by several scientists (e.g., Webster 1987). There is an alternative hypothesis in which the monsoon is considered as a manifestation of the seasonal migration of the intertropical convergence zone (ITCZ; Charney 1969) or the equatorial trough (Riehl 1954, 1979) in response to the seasonal variation of the latitude of maximum insolation. Whereas the first hypothesis associates the monsoon with a system special to the monsoonal region, in the second, the monsoonal regions differ only in the amplitude of the seasonal migration of the basic system (ITCZ/equatorial trough). The two hypotheses have very different implications for variability of the monsoon. For example, in the first case, we expect the intensity of the monsoon to be directly related to the land-ocean temperature contrast.

Simpson (1921) pointed out that the observations of the space-time variations of the monsoon over the Indian region are not consistent with the first hypothesis. "I believe very few educated people would have any difficulties in giving an answer to the question—what is the cause of the monsoon? They would refer to the high temperature over the land compared with that over the surrounding seas; would speak of ascending currents of air causing an indraft of sea-air towards the interior of the country. It is only when one points out that India is much hotter in May before the monsoon sets in than in July when it is at its heights—or draws attention to the fact that the hottest part of India—the northwest gets no rain at all during the monsoon—or shows by statistics that the average temperature is much greater in years of bad rains than in years of good rains, that they begin to doubt whether they know the real cause of the monsoon."

The alternative hypothesis in which the monsoon is considered to be a manifestation of the seasonal migration of the ITCZ (Gadgil 1988) is also not acceptable to several scientists. Murakami (1987) pointed out that the latitudinal extent of the region characterized by low values of (OLR) is much larger over the Indian longitudes in the Northern Hemispheric summer than that characterizing the ITCZ over the Atlantic and Pacific (Figure 7) and suggested that "the ITCZ over the Indian Ocean changes its existence drastically from winter to summer."

The first study of the daily satellite imagery over the Indian longitudes (Sikka & Gadgil 1980) showed that the cloud band over the Indian subcontinent on an active monsoon day is strikingly similar to that characterizing the ITCZ over other parts of the tropics. Further, Sikka & Gadgil (1980) showed that, dynamically, the system has all the important characteristics of the ITCZ (Charney 1969), including low level convergence, intense cyclonic vorticity above the boundary layer, and organized deep convection. Sikka & Gadgil (1978) showed that the large-scale rainfall over the Indian monsoon zone is directly related to the meridional shear of the zonal wind just above the boundary layer. Sikka & Gadgil's (1980)

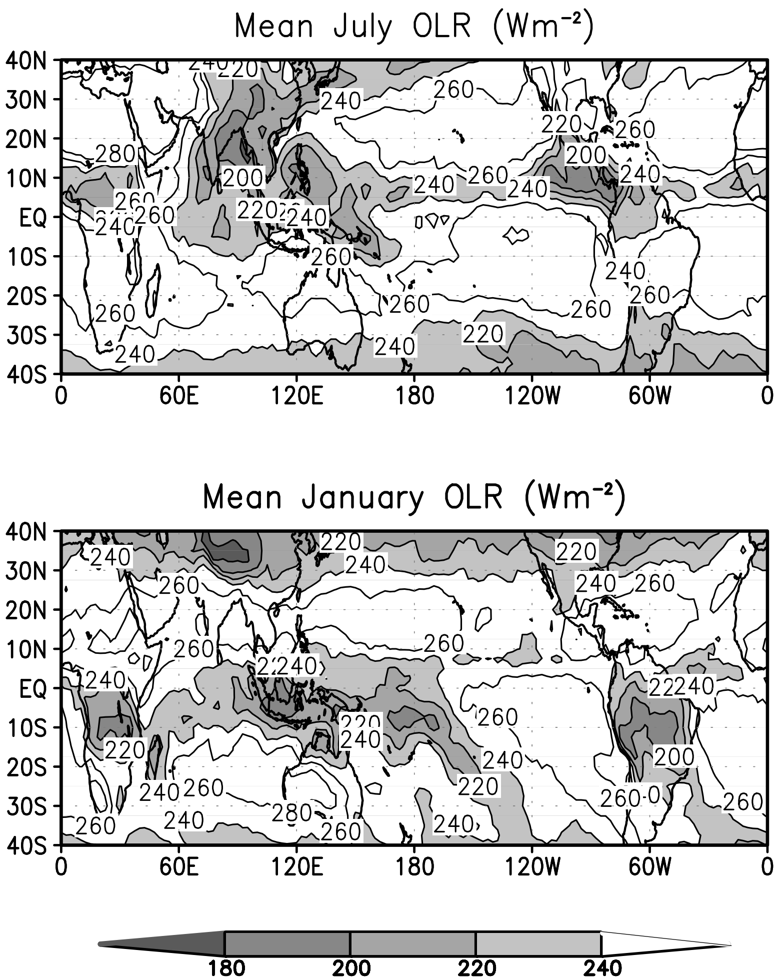


Figure 7 Mean OLR (wm^{-2}) for January and July from data of Walisser & Zhou (1997).

study also pointed out two important features of the variation of the cloudbands over the Indian longitudes during the summer. First, there are two favorable locations for cloudbands/ITCZ: one over the heated subcontinent and another over the warm waters of the equatorial Indian ocean (Figure 8). Because the convergence in both the zones cannot be intertropical, the term tropical convergence zone, TCZ (rather than the ITCZ), is used when referring to the Indian longitudes (Gadgil 1988). A prominent feature of the intraseasonal variation is northward propagations of the cloudbands from the equatorial Indian Ocean onto the Indian monsoon zone at intervals of 2–6 weeks (Figure 9). The seasonal migration of the ITCZ

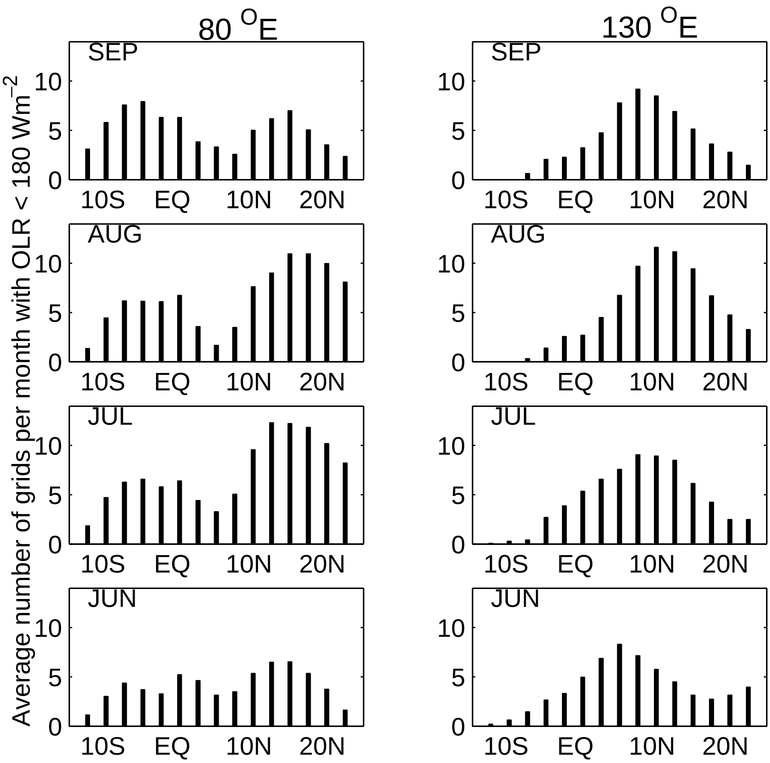


Figure 8 Variation of the occurrence of days in a month with OLR less than 180 wm^{-2}, with latitude at 80°E (central longitude of the Indian region) and 130°E over the western Pacific for June to September.

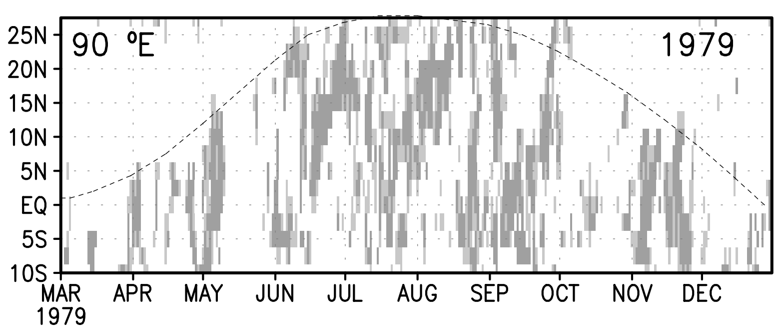


Figure 9 Variation of the grids with OLR less than 200 and 180 wm^{-2} at 90°E during March–December 1979. The seasonal envelope is also indicated.

is accomplished by such northward propagations. These propagations culminate farther and farther northward in the onset phase and more and more southward in the retreat phase. Thus, the retreat of the monsoon is not the mirror image of onset as believed earlier (e.g., Riehl 1979, p. 369). Clearly, the large latitudinal extent of the low OLR region noted by Murakami (1987) does not arise from any difference in the basic system, but rather from the special nature of its intraseasonal fluctuations.

These observations lend support to the second hypothesis considering the monsoon as a seasonal migration of the ITCZ. In the steady state, it is expected that the ITCZ over the oceans will be located over the latitude at which the sea surface temperature (SST) is maximum (Charney 1969, Pike 1972). However, Hastenrath (1990, p. 161–65) pointed out that, on the monthly scale over the tropical Atlantic and Pacific, the latitude of maximum convergence and cloudiness/precipitation is 3°–4° equatorward of the latitude of confluence, i.e., the surface trough, and suggested that "the near-equatorial trough over both continents and oceans has a heat low structure and appears, above all, thermally induced." A heat low is characterized with ascent only upto about 2–3 km from the surface, in contradistinction to the ITCZ with ascent throughout the troposphere. Ramage (1971, p. 74), recognizing the importance of heat lows, pointed out that "in a climatological sense, heat lows and near equatorial troughs (read ITCZ) together comprise a continuous low pressure belt." In fact, this is also clearly seen in the surface wind and pressure pattern for July over the Indian region (Figure 1). The well-marked low over the northwestern region is a heat low, which together with the low pressure belt extending westward from the head of the Bay of Bengal (which is associated with organized convection and rainfall), makes up the surface trough zone over the Indian region.

It is interesting that Trenberth et al.'s (2000) analysis of the global divergent circulations derived from NCEP and ECMWF reanalysis data (Kalnay et al. 1996, Gibson et al. 1997) shows that the vertical structure of the first complex empirical orthogonal function (CEOF1, which explains about 60% of the variance) corresponds to that of the ITCZ or near equatorial trough, whereas the vertical structure of the second one, (CEOF2, which explains about 20% of the variance) corresponds to that of a heat low with convergence restricted to the lowest 2–3 km (Figure 1 of Trenberth et al. 2000). The spatial patterns of these two leading CEOFs (Figures 2 and 3 of Trenberth et al. 2000) are consistent with Ramage's (1971) observation of the two systems together making up the surface low pressure belt.

It appears that the response of a tropical atmosphere to heating from the lower boundary is either (*a*) a heat low type circulation with ascent over the zone of maximum heating (or SST) restricted to the lowest 2–3 km or (*b*) ascent throughout the troposphere, i.e., of the ITCZ type. Unravelling the factors that determine the transition from a heat low type circulation in space and time is an important problem of great relevance to the Indian monsoon. Before the onset of the monsoon, the Indian monsoon zone is characterized by the presence of a heat low. At the end of the onset phase, a tropical convergence zone gets established over the region (Sikka & Gadgil 1980). During prolonged breaks (such as the one in 1972 in Figure 5), a heat trough gets established over the monsoon zone (Raghavan 1973). Revival from such breaks again involves a transition from a heat trough to a moist convective regime. On the seasonal scale, the boundary between the heat low regions over the northwestern part of India and the moist convective regions over the eastern part of the monsoon zone also exhibit significant variation from year to year. In poor monsoon years, the boundary between the moist convective regime and the heat low retreats eastward (Parthasarathy et al. 1995).

Because the observations have lent credence to considering the monsoon as a manifestation of the seasonal migration of the ITCZ, investigations of the ITCZ and Hadley cell become relevant for understanding the monsoon. The basic problem is the derivation of the response of a tropical atmosphere to heating (or SST) specified at the lower boundary, which varies with latitude. Schneider & Lindzen's (1976) analysis of a simple axisymmetric model showed that the response is a heat low type circulation with ascent over the latitude of maximum heating restricted to the lower 2–3 km. In order to simulate a more realistic ITCZ, they imposed mid-tropospheric heating, which would occur in the presence of moist convection. Most of the later studies in this genre (e.g., Held & Hou 1980, Lindzen & Hou 1988) assume diabatic heating, which is maximum in the midtroposphere. Lindzen & Hou (1988) showed that when the peak heating is not at the equator, the latitude separating the winter and summer cells (i.e., surface trough), does not occur at the latitude of maximum ascent, i.e., cloudiness and precipitation. Thus, Hastenrath's (1990) observations are not necessarily inconsistent with the structure of the ITCZ away from the equator. The monsoon can, thus, be considered as a manifestation of the seasonal migration of the ITCZ. Chao & Chen's (2001) experiments with general circulation models (GCMs) have also supported this hypothesis. Hence, the central problem in monsoon meteorology is understanding mechanisms leading to the variability of the continental TCZ.

The simulation of the continental TCZ over the Indian monsoon zone by atmospheric GCMs has proved to be a difficult problem. Gadgil & Sajani's (1998) analysis of the simulation of precipitation by thirty GCMs participating in the atmospheric model intercomparison project (AMIP; Gates 1992) for a specified observed SST during 1979–1988 showed that several models simulated only the precipitation over the equatorial Indian ocean and not the precipitation over the Indian monsoon zone. Whereas some models simulated the monsoonal rainbelt over India as a manifestation of seasonal migration of the planetary scale tropical rainbelt, a few simulated it as a more localized system. On the whole, the skill in simulation of the interannual variation of monsoon rainfall over the Indian region was found to be much larger for the first set of models, lending further support to the second hypothesis for the monsoon.

Magnitude of the systematic error in simulation of the seasonal mean monsoon, with largest errors in the vicinity of the TCZ, was a major contributing factor to poor predictability exhibited by the integrations performed under the European Union PROVOST project (Brankovic & Palmer 2000). Thus, a reasonable simulation of the mean monsoon is required for understanding as well as prediction. Sperber et al. (2001) showed that a high degree of fidelity is also required in the simulation of the dominant modes of subseasonal variability.

3. INTRASEASONAL VARIATION

The important role of the special nature of the variations of the TCZ within the season in determining the nature of the seasonal pattern of OLR over Indian longitudes has been discussed above. In this section, I consider two features of the intraseasonal variation of the monsoon rainfall on the supersynoptic scale, that is, the fluctuation between active and weak spells/breaks and northward propagations of the TCZ/rainbelt at intervals of 2–6 weeks throughout the summer monsoon. Several important facets, such as the onset of the monsoon (Krishnamurti & Surgi 1987) are not considered for want of space.

3.1. Active-Weak Spells, Breaks

Large intraseasonal fluctuations between active and weak spells of the large-scale rainfall over the monsoon zone (Figure 4) have been known for over a century. Blanford (1886) first described the variation in the circulation between active spells and "intervals of drought," i.e., prolonged dry spells. In his words: "In intervals of drought, when northwesterly and westerly winds interrupt the monsoon in North-Western and Central India, it (the trough of low pressure) is pushed northward to the foot of the hills. On the other hand, during the height of the rains, at certain intervals, true cyclonic vortices, with closed isobars (barometric minima) are formed on or in the immediate neighborhood of this trough."

Blanford's (1886) "intervals of droughts," during which the large-scale rainfall over the monsoon zone is interrupted for several days in the peak monsoon months of July–August, have been called "breaks" in the monsoon by Indian meteorologists for over five decades (e.g., Malurkar 1950, Ramamurthy 1969, Alexander et al. 1978, Raghavan 1973, Krishnamurti & Bhalme 1976, Sikka 1980). Although interruption of monsoon rainfall is recognized as the most important feature of the break, the criterion used by the India Meteorological Department (IMD) and by several meteorologists for identifying a break is the synoptic situation associated with such a rainfall anomaly, rather than the rainfall distribution itself (Rao 1976). In Ramamurthy's (1969) comprehensive study of breaks during 1888–1967, a break situation was defined as one in which the surface trough is located close to the foothills (similar to the situation described by Blanford 1886). Subsequent to Ramamurthy's (1969) classic work, De et al. (1998) have identified the breaks during 1968–1997.

Active spells are characterized by a sequence of time-clustering, partly overlapping development of monsoon disturbances (Murakami 1976) and cyclonic vorticity above the boundary layer (Sikka & Gadgil 1978). These spells correspond to active spells of the continental TCZ in which these synoptic-scale disturbances are embedded. The break phase is characterized by a marked change in the lower tropospheric circulation over the monsoon zone, with the vorticity above the boundary layer becoming anticyclonic (Ramamurthy 1969, Sikka & Gadgil 1978). During long intense breaks, the surface temperature increases rapidly and a heat low type circulation gets established over the monsoon zone with subsidence over most of the troposphere, no disturbances are generated, and the prominent trough at 700 hPa (associated with large-scale monsoon rainfall) disappears (Raghavan 1973). According to Koteswaram (1950), low pressure systems are generated in the equatorial region during breaks. Whereas active-weak cycles in fluctuations of the monsoon rainfall occur every year, breaks do not, and long breaks such as the one in 1972 occur only in a few years. The active-weak cycles in rainfall over the monsoon zone are associated with fluctuations in the intensity of the continental TCZ. Sikka & Gadgil (1980) showed that active (weak) spells of the continental TCZ are associated with weak (active) spells of the TCZ over the equatorial Indian ocean.

As noted above, during intense breaks a heat trough develops over the monsoon zone. Air ascending over the equatorial TCZ descends above the shallow cell associated with the heat trough (Raghavan 1973). Revival of the monsoon occurs with a transition to a moist convective regime, either with northward propagation of the equatorial TCZ (Sikka & Gadgil 1980) or with westward propagation of synoptic-scale systems generated over the warm waters of the Bay of Bengal (Sikka & Dixit 1972, Sikka & Gadgil 1980). Genesis of synoptic scale systems over the Bay often occurs as a result of westward propagation of a disturbance from the western Pacific (Krishnamurti et al. 1977).

Gadgil & Joseph (2003) defined breaks (and active spells) of the Indian monsoon in terms of the rainfall over the western and eastern parts of the monsoon zone (Figure 4). The basic rainfall data analyzed were the daily rainfall during June to September at 273 well-distributed stations over the Indian region for the period 1901–1989 obtained from IMD. A day on which the rainfall is below specified thresholds for western and eastern parts of the monsoon zone was defined as a break day. The thresholds were chosen so as to have maximum possible overlap with breaks, identified by Ramamurthy (1969) and De et al. (1998) on the basis of the synoptic situation as per the IMD definition (Rao 1976). The rainfall anomaly pattern for breaks comprises large negative anomalies over the monsoon zone with negative anomalies of magnitude of about 75% of the mean (50% of the mean) over the western (eastern) part (Ramamurthy 1969). Because the mean July–August rainfall for the western part (6.6 mm/day) is less than that of the eastern part (11.2 mm/day) and the magnitude of the anomalies (expressed as a fraction of the mean) is larger over the western part, the threshold for the western part has to be smaller than that for the eastern part. Gadgil & Joseph (2003) found that if the thresholds are chosen as 2.5 mm/day for the western part (i.e., 66% of the mean) and 7.5 mm/day for the eastern part (i.e., 33% of the mean) maximum overlap with breaks identified by Ramamurthy (1969) and De et al. (1998) is obtained. During 1901–1989, rainfall below the threshold occurred on 33% (38%) of days over the western (eastern) part of the monsoon zone, whereas rainfall was simultaneously below the thresholds over the two parts on 16% of the days.

The composite patterns of rainfall and rainfall anomaly for these breaks during 1901–1989 are shown in Figure 10. Because there is a systematic bias in the OLR derived from the NOAA satellite series (Gadgil et al. 1992), daily OLR data from which the bias was removed by Lucas. et al. (2000) using the method of Waliser & Zhou (1997) was used by Gadgil & Joseph (2003). The see-saw between the two TCZs over Indian longitudes is manifested as a band of significant negative correlation with respect to the OLR of the monsoon zone (Figure 11*a*).

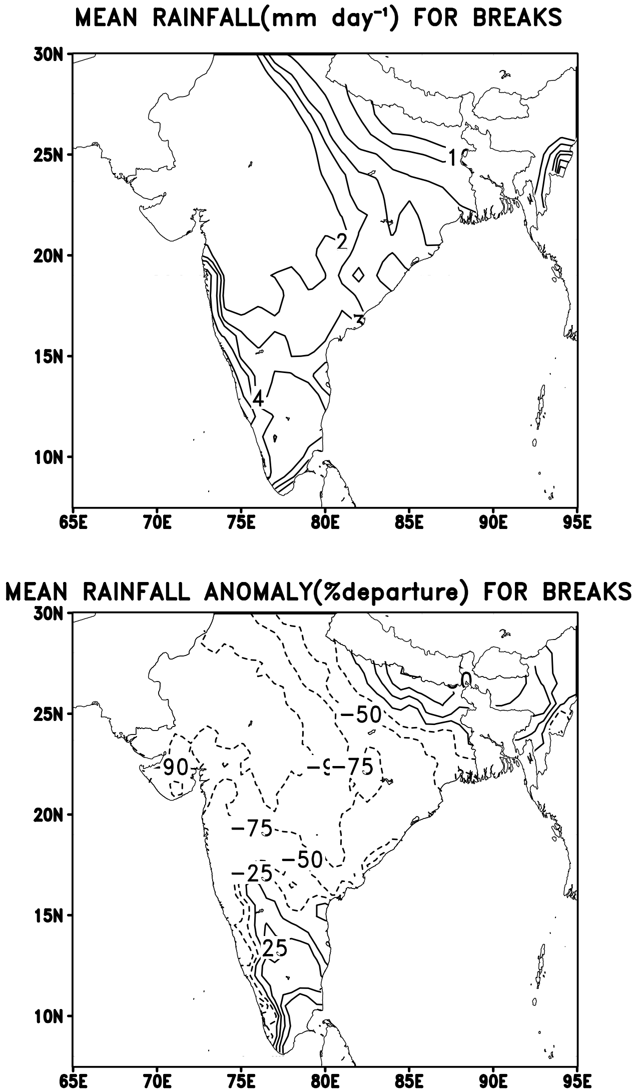


Figure 10 Break composite patterns of rainfall and rainfall anomaly (after Gadgil & Joseph 2003).

Convection over the region 95°–140°E north of 25°N is also seen to vary in opposite phase to that over the Indian monsoon zone. However, over most of the western Pacific, between the equator and 20°N, the correlation is not significant.

The composite OLR anomaly pattern (with respect to the July–August mean) for all the breaks larger than five days during 1974–1989 is shown in Figure 11*b*. In the break composite, the OLR is higher than 240 wm^{-2}, and OLR anomalies are positive and larger than 30 wm^{-2} over the most of the Indian monsoon zone, and over almost the entire Indian region. A prominent feature is the low OLR (below 200 wm^{-2}) with large negative anomalies over the equatorial region of the Indian longitudes associated with anomalously active oceanic TCZ. The occurrence of low pressure systems in the equatorial region reported by Koteswaram (1950) is thus a manifestation of an active spell of the oceanic TCZ during breaks.

Active spells are characterized by well-distributed rainfall over the monsoon zone, with rainfall more than the mean over western and eastern parts. Gadgil & Joseph (2003) used thresholds of 8 mm/day (12.5 mm/day) for the western (eastern) part of the monsoon zone to identify days of active spells. During 1901–1989, the rainfall is higher than the threshold over western or eastern parts on 33% of days, whereas it is simultaneously higher than the thresholds over the two parts on 14% of the days. The composite OLR anomaly pattern for active spells of 4 to 7 days during 1974–1989 duration are shown in Figure 11*c*. As expected, the OLR active composite comprises low OLR over the monsoon zone (with negative anomalies) and high OLR over the equatorial region between 70–90°E with positive anomalies. This corresponds to an active phase of the continental TCZ and weak phase of the oceanic TCZ over the equatorial region. Thus, over the Indian longitudes, the sign of the north–south dipole for the active composite is opposite to that for the break composite. However, in the active composite there is band with low OLR (as well as negative OLR anomalies) that is almost continuous from 70°E to 150°E, whereas in the break composite the equatorial band is seen only over the Indian longitudes. Hence, over the western Pacific, the active composite is not the mirror image of the break composite, and analysis of active minus break anomalies, in the expectation of doubling the signals (Annamalai & Slingo 2001, Webster et al. 1998) may be misleading for the patterns over the western Pacific.

An interesting feature of Figure 11 is the association of what happens over the Indian monsoon zone with convection over the eastern Pacific between 10°–20°N. The OLR over this part of the eastern Pacific is negatively correlated with the OLR of the monsoon zone (Figure 11*a*) and active (break) spells of the Indian monsoon are associated with positive (negetive) OLR anomalies over this region. In fact, negative OLR anomalies over this region characterize each of the long breaks, including those in seasons such as of 1979 in which there was no El Niño. Thus, what happens over the Indian monsoon zone appears to be linked to convection over the eastern Pacific not only on the interannual scale (as evinced by the link between El Niño and the monsoon) but on the intraseasonal scale as well. More work is needed to determine whether convection anomalies over one region consistently lead the convection anomalies over the other region, or both evolve simulteneously.

Recently "breaks in the monsoon" has become a very popular topic of research. However, different scientists have used the same term to denote different features of convection and/or circulation over different regions. Thus, Webster et al. (1998) use the term break, as defined by Magana & Webster (1996), to denote weak spells of convection and 850 hpa zonal winds over a large-scale region (65°–95°E, 10°–20° N) because it is believed that breaks in the Indian monsoon are on a scale that is much larger than India or even South Asia. On the other hand, Goswami & Mohan (2001) define breaks on the basis of the strength of the 850-hpa wind at the single grid-point 15°N, 90°E. Even when the so-called breaks are identified in terms of rainfall or convection over the Indian region, a variety of definitions are used. Rodwell (1997) and Annamalai & Slingo (2001) use the term break to denote weak spells of the daily all-India average rainfall (calculated operationally by the IMD), whereas Cadet & Daniel (1988) denote the weak spells of the average rainfall at 15 stations distributed throughout India as breaks in the monsoon. Krishnan et al. (2000) define break days as days with positive OLR anomalies over northwest and central India (i.e., only over the western part of the monsoon zone), provided the average OLR anomaly over 73°–82°E, 18°–28°N exceeds 10 wm^{-2}.

It should be also be noted that in the traditional approach, break is defined as an event with specified characteristics, e.g., the surface pressure pattern, whereas in several studies, such as Annamalai & Slingo (2001) and Goswami & Mohan (2001), active spells and breaks are identified for a specific mode of a particular field. In some cases, the break or active minus break composites for such modes have a well-defined structure, e.g., the 20–60 day mode of Annamalai & Slingo (2001). However, the active minus break OLR composite of their 10–20 day mode is characterized by very small values of OLR, generally less than 4 wm^{-2}. The relevance of such a mode is not clear.

Because the definition of breaks varies in different studies, there are differences in the breaks identified in any season, their duration, and their frequency of occurrence. Gadgil & Joseph (2003) showed that there is hardly any overlap between the breaks identified by them and those of Webster et al. (1998). Webster et al. (1998) state that within each summer monsoon, there are three or four active break sequences, whereas for the breaks identified by Ramamurthy (1969) during 1901–1967, there were no breaks in 10 of the years. Although the duration of the breaks identified by Ramamurthy (1969) varies from 3 days (the minimum possible duration considered) to over 15 days, with over 30% of the breaks of duration 7 days or longer, the duration of the breaks identified by Webster et al. (1998) is generally short, varying from 1 to 7 days with 90% of the breaks of duration 3–5 days.

There are also differences in the patterns of circulation, anomaly of the circulation, OLR, and OLR anomaly associated with breaks based on different definitions. Thus, whereas an anomaly in the cross-equatorial flow over the western Arabian Sea is an important feature of the active minus break patterns of Annamalai & Slingo (2001) and Webster et al. (1998) as well as breaks of Krishnan et al. (2000), there is hardly any anomaly of the meridional circulation over this region for the Goswami & Mohan (2001) breaks. The negative OLR anomaly corresponding to active minus break of the 20–60 day mode in Annamalai & Slingo (2001) is maximum around 20°N over the Indian region, whereas that of the active minus break of Webster et al. (1998) is maximum near 10°N over the Indian peninsula.

It is interesting that, consistent with Figure 11, depressed convection over the equatorial region is seen in the composite of active minus break for the Webster et al. (1998) study as well as the 20–60 day mode of Annamalai & Slingo (2001). For the break composite, large negative OLR anomalies are seen over the region of the western Pacific north of 15°N in the break composite (Figure 11). The OLR anomaly pattern involving the north-south dipole over the Indian region and another (opposite in phase) over the western Pacific obtained by Gadgil & Joseph (2003) is similar to the quadrapole in the active minus break pattern for the 20–60 day mode of Annamalai & Slingo (2001). However, note that a pole with reverse sign is not seen over the western Pacific in the active composite (Figure 11). Thus, although the north-south dipole over the Indian longitudes appears to be a basic feature of the intraseasonal variation, the quadrapole present in the active minus break composite by Annamalai & Slingo (2001) probably arises only from the break phase.

The relationship of interannual variation of the monsoon rainfall to occurrence of breaks has been known for at least five decades since Ramdas's comparison of the weekly rainfall over central India during the excess monsoon season of 1917 when the deficit year of 1918 (Figure 1 in Normand 1953) showed the large impact of the dry spell in the latter. Krishnamurti & Bhalme (1976) pointed out that the major difference between the rainfall variation in good and poor monsoon seasons is the occurrence of a long dry spell (break) in the latter (e.g., Figure 5). Sikka (1980) also showed that the average number of break days in poor monsoon seasons is much larger than in good monsoon seasons. Gadgil & Joseph (2003) found that the ISMR is negatively correlated (correlation coefficient $= -0.58$) with the number of break days and positively correlated (0.47) with the number of active days. It is interesting that the basic feature of breaks, that is, the anomalously active TCZ over the equatorial Indian ocean, is also seen in the anomaly patterns of poor monsoon seasons such as 1979, 1987, etc. Recently, the relationship between intraseasonal and interannual variation has been investigated by Ferranti et al. (1997), Goswami et al. (1998), Sperber et al. (2000), and Lawrence & Webster (2001), but as yet there is no consensus about the nature of the links.

3.2. Intraseasonal Variation of TCZ

The first step in understanding the intraseasonal variation of the TCZ over the Indian longitudes is to ascertain from observations which among the major features, namely, (*a*) the fluctuations between active and weak spells, (*b*) existence of two favourable zones—latitudinal belts (Figure 8), and (*c*) poleward propagations (Figure 9), are basic features of the TCZ and which are special to the Indian monsoonal region. Gadgil & Srinivasan (1990) addressed this problem with analysis of TCZ variation over different parts of the tropics using a bispectral algorithm based on OLR and albedo data developed by Gadgil & Guruprasad (1990). They found that active-weak cycles are an ubiquitous feature of the TCZ, although the timescales vary in different regions. For example, active spells tend to be of shorter duration over the African region (10°E) than the Indian monsoon zone during the Northern Hemispheric summer. Bimodality (existence of two favorable zones) was found to be a special feature of the Asian summer and winter monsoon regimes. Although poleward propagations of the TCZ are occasionally seen over other regions, they appear to be a basic feature of the TCZ variation only over the Asian summer monsoon zone. Thus, we expect the mechanisms leading to active-weak cycles to be related to the intrinsic features/dynamics of the TCZ, such as cloud-radiation feedbacks. On the other hand, special features of the Asian monsoon zone are likely to be important in poleward propagations of the TCZ.

A class of zonally symmetric models of increasing complexity have yielded increasingly realistic simulations of the intraseasonal variation of the TCZ over the Indian longitudes during the summer monsoon. The first model to simulate poleward propagations of the TCZ was the simple, two-level, zonally symmetric climate model developed by Webster & Chou (1980a,b). However, the simulated propagations were restricted to the region over the continent (whereas the observed propagations are across the equatorial ocean and the continent) and were far more frequent than observed. Gadgil & Srinivasan (1990) and Srinivasan et al. (1993) modified the model to incorporate thermal inertia of land and a realistic SST distribution and simulated more realistic propagations across ocean and continent. In all of these simulations, at the culmination of each propagation, the TCZ over the continent disappeared and another propagation commenced. Thus, the active phase of the monsoon at the end of each poleward propagation in which the TCZ fluctuates over the monsoon zone (around 20°N) was not simulated. A realistic simulation (Figure 12) of all the important features was obtained by the model developed by Nanjundiah et al. (1992).

Webster (1983) had suggested that hydrological feedbacks leading to cooling of the land surface beneath the TCZ, i.e., to a perturbation in sensible heat flux, played an important role in poleward propagations. Goswami & Shukla (1984) also suggested that hydrological feedbacks gave rise to propagations simulated in the all-land case of the symmetric GLAS model. Obviously, mechanisms based on hydrological feedbacks cannot explain the propagations over the ocean. The mechanism of propagation over land and ocean was identified by Gadgil & Srinivasan (1990) and Srinivasan et al. (1993). They showed that the north-south differential in total heating arising from the north-south gradient in the convective stability and moisture availability led to the profiles of convective heating having a maximum northward of the maximum in the profile of vertical ascent and hence propagation. The same mechanism was shown to operate for the propagations simulated in the model of Nanjundiah et al. (1992). The simulation of the active phase with TCZ persisting over the continent at the culmination of the

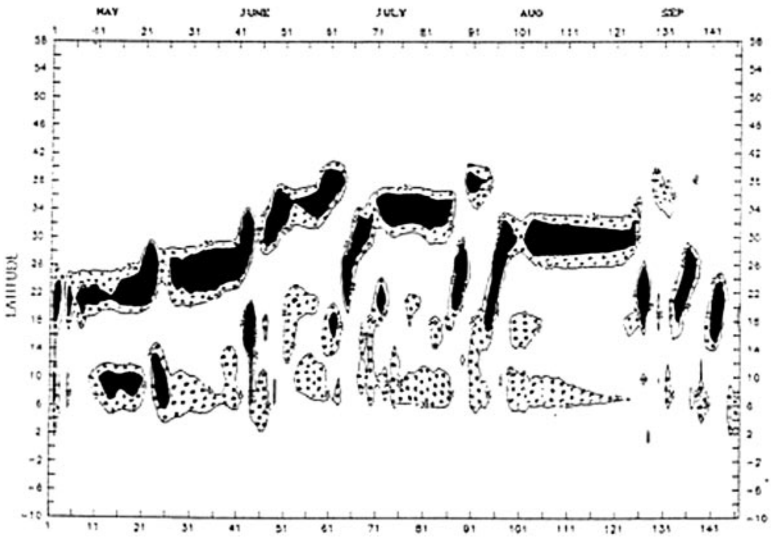


Figure 12 Simulation of intraseasonal variation of TCZ with a simple axi-symmetric model by Nanjundiah et al. (1992).

poleward propagation by Nanjundiah et al. (1992) was found to be associated with anchoring of the TCZ in the surface trough. The duration of the active spell was shown to depend on hydrological processes over the region equatorward of this trough.

3.3. Approach

Two distinct approaches have been adopted in the study of intraseasonal variation. In the first, the more traditional approach, the focus is on special events, such as breaks in the monsoon, or special features of some systems, such as the propagation of the TCZ. The major features of these events are studied with analysis of data and attempts are made at their simulation by models in order to understand the underlying mechanisms. In the second approach, the variation is viewed as a superposition of waves/modes or spatial patterns. Structure and evolution of these modes is investigated in detail for further understanding and generating predictions.

The important timescales of variation of the intraseasonal variation have been identified in the past three decades as the 10–20 day and the 30–50 day scales, in addition to the synoptic scale (Sikka & Mishra 1974; Krishnamurti & Ardanuy 1980; Krishnamurti & Bhalme 1976; Bhalme & Parasnis 1975; Yasunari 1979, 1980; Lau & Chan 1986). The Madden & Julian (1972) 30–50 day mode is believed to be associated with northward propagations from the equatorial Indian ocean (Wang & Rui 1990), whereas the 10–20 day mode with westward propagation from the west Pacific (Krishanamurti & Ardanuy 1980).

Prominent features of variation, such as northward propagations, are seen irrespective of the approach. In that case, the first approach reveals that although the dominant timescale for the interval between successive propagations is 30–50 days, there is considerable variation in this period within the same season or between year to year. On the other hand, when fields such as the 850-hpa wind field are filtered for the dominant timescales, the variation appears more like a simple oscillation with northward propagations of troughs and ridges at regular intervals (Krishnamurti & Subrahmanyan 1982).

A complementary approach to analysis of fields corresponding to a specific temporal scale is a description of the fields at each time interval in terms of superposition of waves or empirical orthogonal functions (EOFs), which are optimum for economical description of variation (Lorenz 1956). One of the problems with this approach for intraseasonal variation is that only a small fraction of the variance is explained by the leading few EOFs. For example, Sperber et al.'s (2000) analysis of the daily 850-hPa flow during 1958–1997 revealed that the first four EOFs explained only 13.0%, 7.6%, 6.6%, and 4.9% of the variance, respectively.

Two distinct approaches have also been used in description of turbulence (Narasimha 1989). Classical description in the language of generalized harmonic analysis yielded important insights in the 1930s and 1940s, particularly on homogeneous, isotropic turbulence. In the 1970s, it was recognized that turbulence is extremely anisotropic and intermittent, occurring in sporadic bursts that produce most of the fluid's Reynold's stress in less than 10% of the time (Mollo-Christensen 1976). Over the past 25 years, evidence has accumulated that in turbulent shear flows, there is considerable degree or order, as revealed by the spectacular pictures of coherent structures by Brown & Roshko (1974). Ramage (1976) suggested that turbulence "bursts" are the most important events of atmospheric change on all space- and timescales. It has now been shown that an episodic description is both feasible and natural for the fluxes in the atmospheric boundary layer (Narasimha 1989, Narasimha & Kailas 1990). In atmospheric science and in monsoon meteorology, the trend seems to be in the opposite direction from an episodic description (which was natural because the scales of the systems were so large that coherent structures were prominent in weather maps/satellite imagery) to one in terms of modes of different temporal scales or EOFs. Which of the two approaches (or combination thereof) is most useful for providing insight into the mechanisms will become clear in the years to come.

4. CONVECTION OVER OCEANS

4.1. Monsoon and the Convection Over Oceans

The TCZ over the Indian monsoon zone often extends eastward up to the western Pacific and beyond. The continental TCZ over the Indian region is maintained by westward propagation of synoptic systems from the Bay of Bengal and by northward propagation of the TCZ from the equatorial Indian Ocean. Genesis of synoptic-scale systems over the Bay often occurs as a result of westward propagation of a disturbance from the western Pacific (Krishnamurti et al. 1977). Thus, the TCZs over the western Pacific and the equatorial Indian Ocean play an important role in maintaining the continental TCZ. However, the relationship of the continental TCZ to these TCZs is complex. Because the TCZ over the equatorial Indian Ocean occurs over the same longitudinal belt as the continental TCZ, there is competition between the two, leading to active spells of one being associated with weak spells of the other. An active TCZ over the equatorial Indian Ocean is, therefore, a prominent feature of the break composite (Figure 11*b*). When the systems over the western Pacific move along a northward track instead of a westward one, the convection over the Indian monsoon zone becomes weaker. In the break composite of OLR anomaly, negative anomalies are seen over the western Pacific north of 20°N (Figure 11*b*).

The strong link between El Niño Southern Oscillation (ENSO, Philander 1990) and the Indian summer monsoon (Sikka 1980, Pant & Parthasarathy 1981, Rasmusson & Carpenter 1983, etc.) is a manifestation of the role of the TCZ over the Pacific Ocean in the interannual variation of the monsoon. During the period since 1974, for which OLR data are available, there have been severe droughts (with ISMR deficit above 15%) in the seasons of 1979, 1987, and 2002. In July 2002, the deficit of all-India rainfall was an unprecedented 49%. The OLR anomaly patterns over the tropical region from 40°E to 60°W for July 1979, 1987, 2002 are shown along with the break composite in Figure 13. There is considerable similarity between the OLR anomaly patterns of the break composite and those for July of 1979, 1987, and 2002. In particular, two features of the break composite, namely, enhanced convection over the western Pacific north of 20°N and eastern Pacific between 10° and 20°N, are seen in the July patterns of each of the three years. In the El Niño years of 1987 and 2002, a coherent band of negative OLR anomalies stretches across the Pacific. Such a band is not seen in the OLR anomaly pattern of July 1979. The spatial extent of the region with negative OLR anomalies over the equatorial Indian Ocean (which is a prominent feature of the break pattern) varies considerably between the three droughts. It is more prominent in the patterns of 2002 and 1979 than in 1987. In July 1979, the band with negative OLR anomalies extends from the Indian Ocean up to about 150°E in the western Pacific.

It appears that there is a difference in the relative contributions of the TCZs over the equatorial Indian Ocean and the Pacific Ocean in the deficit monsoon over the Indian region between the season of 1979 and the El Niño seasons of 1987 and 2002. In the former case, the largest anomalies occur over equatorial parts of the Indian and western Pacific Ocean, suggesting the importance of factors local to the Asia–western Pacific region. During El Niño seasons, global factors may be more important. This needs to be investigated further.

Given the links of the variability of the Indian monsoon to the convection over the oceans on intraseasonal and interannual scales, it is clear that identification of the factors that determine variability of convection over the oceans is important for understanding the variability of the monsoon.

4.2. Convection Over the Tropical Oceans and SST

It is well known that SST plays an important role in determining the variability of convection over the oceans (Palmen 1948, Bjerknes 1969). Systematic investigation of the variation of convection and its relationship with SST became possible only after the availability of satellite data. Studies of the variation of organized deep convection over tropical oceans with SST based on different measures of convection, such as cloudiness intensity (Gadgil et al. 1984), OLR, (Graham & Barnet 1987), and the frequency of highly reflective clouds (HRC) (Waliser et al. 1993), all suggest a similar and highly nonlinear relationship (Figures 14 and 15). Over cold oceans, with SST below the Palmen threshold of 26.5°C, OLR is above 240 wm^{-2}, suggesting that no organized convection occurs. The propensity for convection increases with SST and around 27.5°C, and the mode shifts to less than 240 wm^{-2}, implying the occurrence of organized deep convection. From this, Gadgil et al. (1984) concluded that there is a threshold value of SST of about 27.5°C for the Indian ocean, above which the propensity for organized deep convection is high. Graham & Barnett (1987) showed that the nonlinear relationship between SST and convection (first discovered by analysis of data over the Indian Ocean) is a basic feature of organized deep convection over the tropics, by analysis of a better data set on convection, namely, OLR. However, the threshold value of SST varies a little from one ocean to another (Waliser et al. 1993). The mean convection (mean OLR) increases (decreases) rapidly in a range of about 1°C around the threshold (Figure 14). When the SST is above the threshold, there is considerable variation in convection. Hence, SST being above the threshold is a necessary but not a sufficient condition for organized deep convection. This is not surprising because tropical convection is also known to depend on other factors, such as large-scale convergence. Graham & Barnett (1987) attributed the variation in convection for SST above the threshold to the variation in the surface divergence.

A detailed study by Bony et al. (1997a) has clearly demonstrated that the variation of precipitation with SST is maximum around the threshold of 28°C when the entire tropical region is considered. Analysis of the important dynamical fields, such as vertical velocity at 500 hpa and upper level divergence, from the NCEP reanalysis data set by Bony et al. (1997b) and Lau et al. (1997) has suggested that the relationship between the zones with ascent of air and regions with high SST is much stronger than supposed thus far. The variation of the fraction of the tropical region (between 20°N and 20°S), with upward velocity of different magnitudes, with SST was analyzed. It was shown that 90% of the region with SST between 29°C–30°C is characterized by ascent. For SST above 30°C, the fraction of the ascending zone decreases to about 75%, indicating cloud-free conditions over a quarter of the warmest oceans. Whether a heat low type circulation occurs in the atmosphere over such warm seas needs to be investigated.

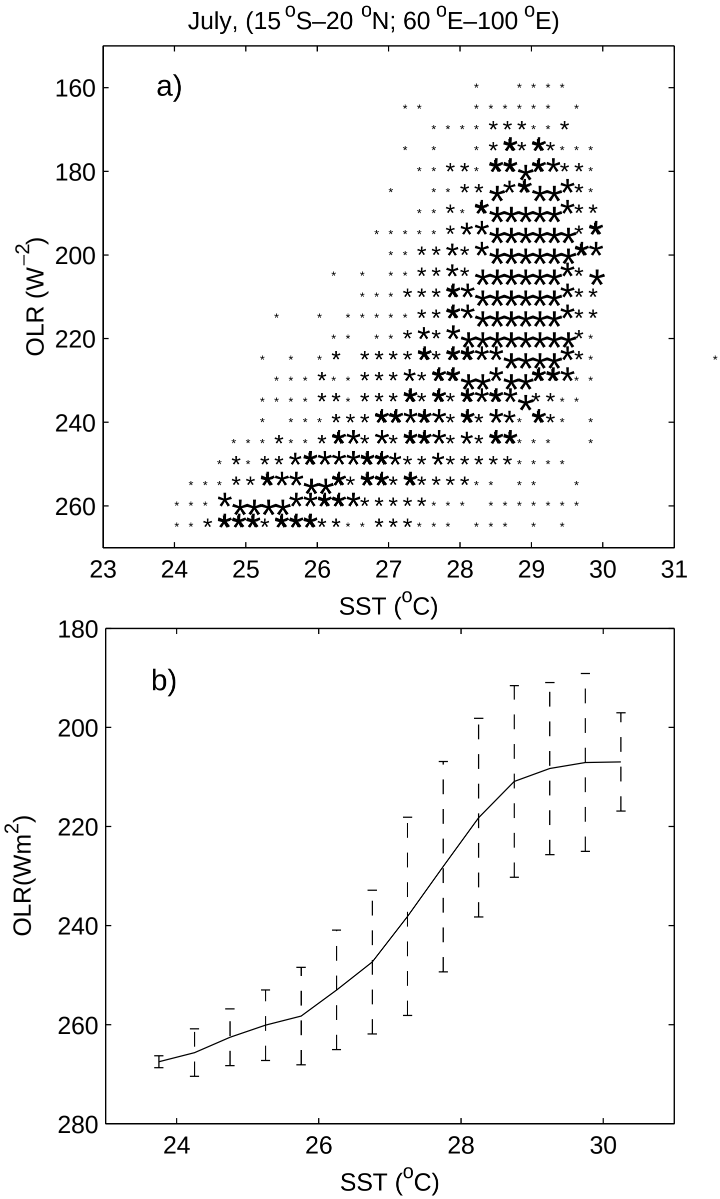


Figure 14 For July (1982–1998) over the Indian Ocean (15°S–20°N, 60°E–100°E). (*a*) Variation of OLR and SST. (*b*) Variation of the mean and standard deviation of OLR with SST.

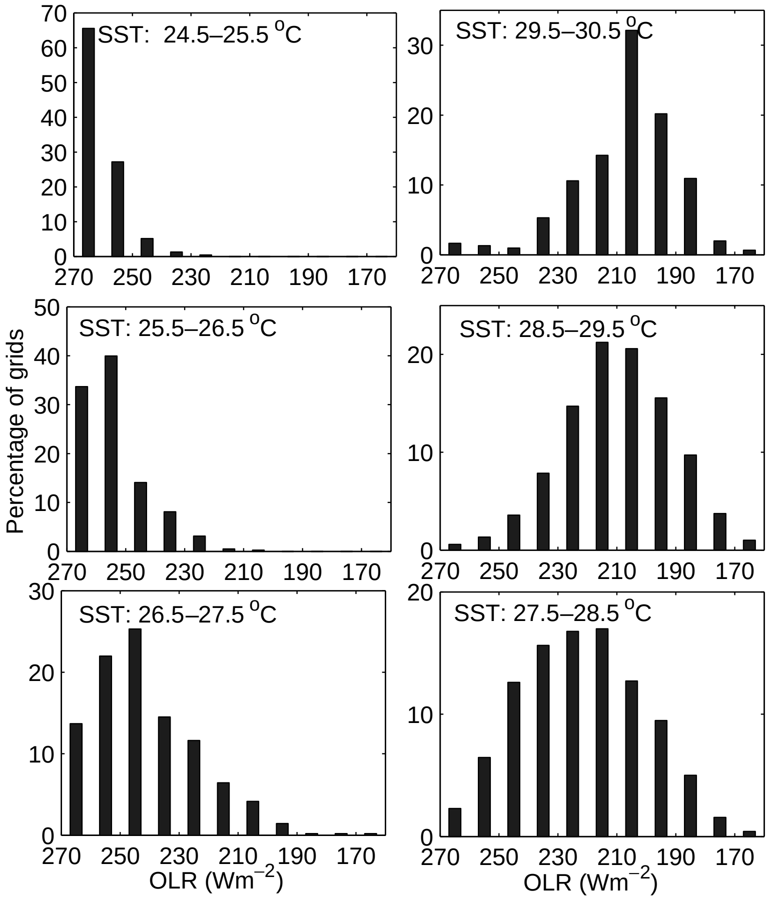


Figure 15 For July (1982–1998) over the Indian Ocean, percentage of grids in different OLR ranges for specified SST ranges.

Variations of SST appear to play an important role in determining the variability in convection when the SST is around the threshold, as is the case for the central Pacific, and not when the SST is maintained well above the threshold, as for the Indian Ocean (Figure 16). Thus, the relationship between convection over warm oceans (with SST above the threshold) to oceanic conditions is more complex. Hence, although it is clear that the variability of the monsoon is linked to the convection over the oceans, the correlations between monsoon rainfall over India and SST of the Indian seas on the interannual scale are poor (Shukla & Misra 1977, Rao & Goswami 1988). In fact, the impact of the atmosphere on the ocean comes out as the stronger signal, with SST in the seasons following a good monsoon being lower (Joseph & Pillai 1984, Shukla 1987).

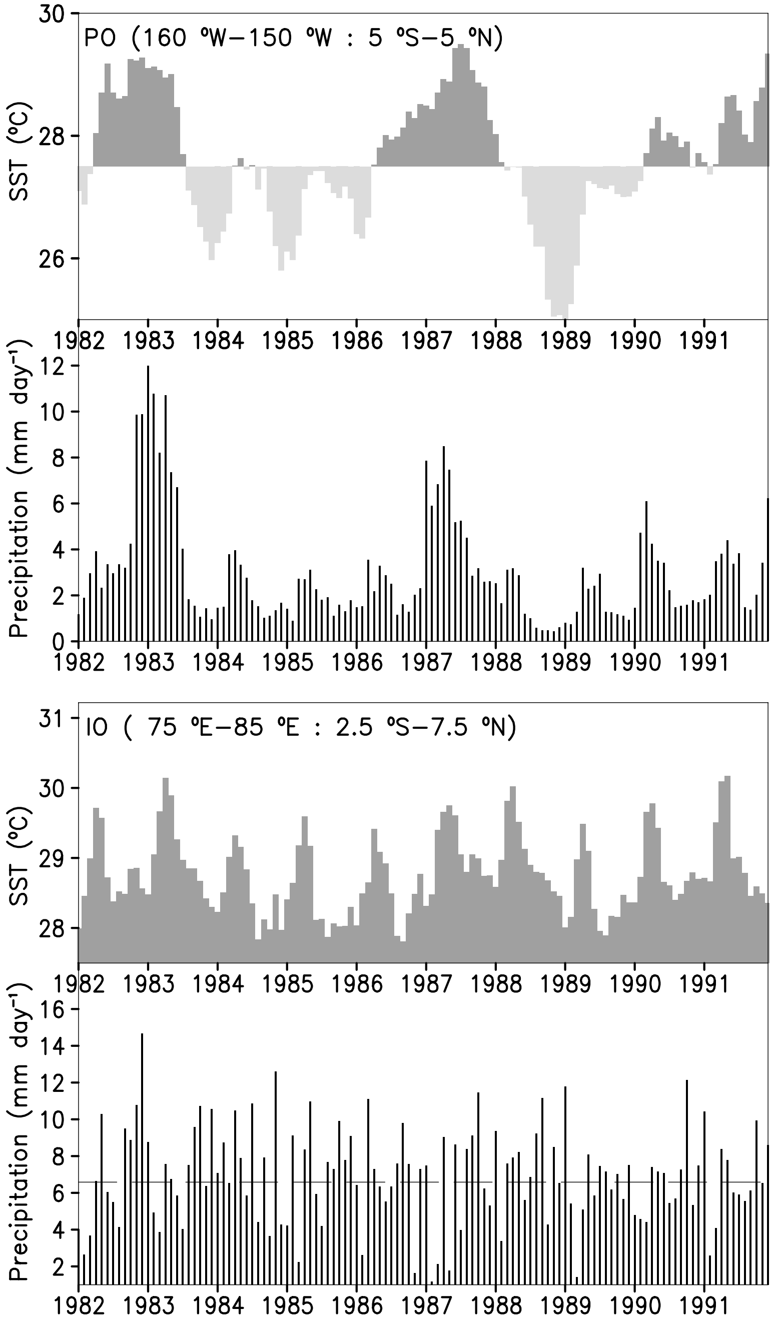


Figure 16 Variation of monthly SST and MSU precipitation over parts of the Pacific and Indian Oceans.

5. MECHANISMS

5.1. Convection Over the Ocean and SST

The first possibility suggested for explaining the relationship of convection over the ocean and SST is the nonlinearity between the saturation water vapour pressure and the temperature (Clausius–Claperon relation). However, it has been shown that this nonlinearity is not adequate to explain the observed nonlinearity of the SST–convection relationship (Neelin & Held 1987, Lau & Shen 1988). Deep convection in the tropics can be maintained only when there is a positive cloud buoyancy over a large depth of the atmosphere. The buoyancy of the cloud depends not only on the properties of the air near the surface of the ocean, but also on those of the upper air through which it rises. Hence, we expect a parameter that incorporates the effects of surface and the air above to be better correlated with deep convection than one that includes only the properties of surface air. One such parameter is the convective available potential energy (CAPE), which is a measure of the vertical instability of the atmosphere when accompanied by a phase change of water (Moncrief & Miller 1976), and hence, determines whether the surface air can ascend. It is known that, as in the case of deep convection, CAPE is zero below a certain SST; positive CAPE is necessary for occurrence of deep convection and CAPE plays an important role in systems ranging from thunderstorms (Williams & Renno 1993) to mesoscale systems (Moncrief & Miller 1976). CAPE decreases during periods of active convection and builds up when convection weakens or disappears. Bhat et al. (1996) showed that the frequency of deep convection (HRC) is highly correlated with CAPE, determined from monthly mean profiles, and that the relationship between this CAPE and SST is strikingly similar to that between frequency of convection and SST.

CAPE represents the work done by the buoyancy force acting on a parcel under moist adiabatic ascent in the atmosphere only when it is based on individual soundings. Bhat et al. (1996) demonstrated that CAPE based on monthly mean soundings represents the work potential of the atmospheric heat engine with ascent in the region of deep clouds and descent in cloud-free regions. Because the number of days with deep convection is typically 4 days/month, with a maximum of about 10 days/month in a few regions such as the Bay of Bengal during the summer monsoon (Garcia et al. 1985), the monthly mean sounding at any station represents the average thermal structure with greater weightage to nonconvective periods. Thus, average convective instability of the region as a whole (including ascending and descending limbs) has been shown to be relevant for convection over the oceans.

5.2. Convection Over Land

The relationship of convection over land to the surface temperature differs markedly from that of convection over ocean to SST (Srinivasan & Smith 1996) because of the large variations in relative humidity over land. A parameter that takes into account the surface temperature as well as specific humidity is the surface moist static energy. Srinivasan & Smith (1996) showed that the relationship between OLR and surface moist static energy for land or ocean is similar to that between OLR and SST. However, the threshold surface moist static energy for land (340 kJ/kg) is somewhat lower than that for ocean (350 kJ/kg) because for the same value at the surface, the moist static energy over land tends to be higher than over the ocean, up to about 600 hPa (Riehl 1979, p. 611). Srinivasan & Smith (1996) showed that there is a universal relationship between OLR and the integral of the moist static energy over the lower troposphere from surface to 400 hPa (Figure 17). The difference between moist static energy of the lower troposphere considered here and the upper troposphere is a major factor in determining the vertical stability of the atmosphere (Neelin & Held 1987).

Srinivasan (1997) found that deep convection occurs in regions with low values of the vertical moist stability, i.e., when the atmospheric profiles are near neutral stability. He showed that occurrence of low values of the vertical moist stability is also a necessary, but not sufficient, condition for deep convection/precipitation. He further demonstrated that there is a close correspondence between regions characterized by low values of the vertical moist stability and those with high values of CAPE.

Zhang (1994) showed that the differences between two simulations of precipitation by a GCM can be related to the differences between the vertical moist

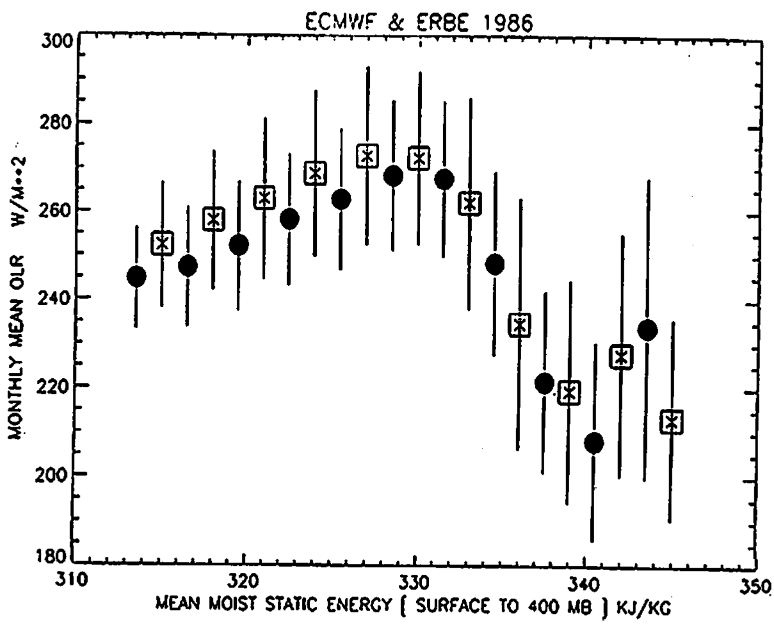


Figure 17 OLR versus lower tropospheric moist static energy for grids over land (*squares*) and ocean (*solid circles*), after Srinivasan & Smith (1996).

stability. Nanjundiah (2000) found that the difference in the simulated precipitation patterns between two different schemes for moisture transport in the NCAR (National Centre for Atmospheric Research) GCM was associated with changes in the vertical moist stability. Nanjundiah & Srinivasan (1999) showed that the vertical moist stability plays an important role in the evolution of El Niño.

5.3. Propagations of the TCZ Over Oceans and Continents

In the set of models that simulated poleward propagations of the TCZ (Section 3.2), the cumulus parameterization scheme based on Ooyama (1969) and Anthes (1977) was used. This convective heating depends on the moist static stability of the atmosphere, which for the two layer models is simply the difference between the moist static energy at two levels. During the summer, the moist static stability generally increases with latitude, up to the latitude of the seasonal surface trough. Whenever a TCZ is generated at a latitude equatorward of this region, the convective heating to the north of the TCZ is larger than that under the TCZ. Hence, the peak in the meridional profile of heating is northward of the peak in the profile of ascent/precipitation, leading to northward propagation of the TCZ. Gadgil & Srinivasan (1990) demonstrated the critical role of the meridional gradient of the moist static stability by showing that when the gradient was assumed to be zero, the propagations vanished.

Thus, the lower tropospheric moist static energy and the vertical stability (which is related to it) play an important role not only in determining regions of organized convection but also in their temporal variation during the season. Detailed analysis of the nature of intraseasonal variation of the TCZ simulated by GCMs to assess the skill in simulating the important features and identify the mechanisms have not, as yet, been undertaken to test the hypotheses proposed by these investigations with simple models. However Rajendran et al. (2002) have shown that the NCAR model does simulate active-weak fluctuations over the Indian region, as well as poleward propagations (particularly when the moist convective adjustment scheme is used for cumulus parameterization). These propagations occur over the region of positive meridional gradient of moist static energy of the lower troposphere, which is consistent with the mechanisms proposed by the simple models.

5.4. Convection and Vertical Moist Stability

Thus, CAPE (based on monthly mean soundings) or a related parameter, namely the vertical moist stability, play an important role in determining the location of organized convection over ocean or land on seasonal scales. The variation of vertical moist stability has also been shown to be important in generating poleward propagations. The mechanisms proposed for the fluctuation between active and weak spells, such as hydrological feedbacks or impact of the mid-tropospheric heating associated with convection (Krishnamurti & Bhalme 1976), also imply changes in the vertical moist stability. It appears that these parameters contain the essential elements of thermodynamics as well as the dynamics that govern the variation of organized convection and precipitation over tropical continents and ocean and hence the variation of the monsoon as well.

6. RECENT DEVELOPMENTS

Understanding and predicting the variability of the Indian monsoon is extremely important for the well-being of over one billion people and the diverse flora and fauna inhabiting the region. Naturally, the major aim of the Indian Climate Research Programme (ICRP), launched in 1995 (Department of Science and Technology 1996), is the study of the variability of the monsoon on subseasonal, interannual, and longer timescales with observational experiments and empirical and theoretical investigations. An important component of the ICRP is observational experiments, particularly over the Indian seas and land mass, for obtaining high quality data to yield new insights into the variation of organized convection and air-sea and land-air interactions and to test the hypotheses put forth for mechanisms.

The major international monsoon experiment MONEX (Krishnamurti 1985) in 1979 was followed after about a decade by a national observational experiment on the monsoon boundary layer (MONTBLEX) (Goel & Srivastava 1990, Narasimha et al. 1997). The first observational experiment over the Indian seas under the ICRP was the Bay of Bengal Monsoon Experiment (BOBMEX), conducted during July–August 1999 (Bhat et al. 2001). During BOBMEX, high-resolution measurements of the vertical profiles of temperature and humidity from which reasonable estimates of CAPE/vertical moist stability could be obtained were made over the north Bay. It is well known that CAPE builds up in the weak phases and gets depleted in the active phases. The composite profiles of temperature and specific humidity for the three active and weak phases of convection from these observations (Figure 18; Bhat et al. 2002) clearly show the large variation in vertical moist stability in association with active-weak fluctuations. It was found that CAPE decreased by 2–3 kJ/kg^{-1} during convective episodes but recovered in 2–3 days (Bhat 2001). The quick recovery of CAPE suggests that the thermodynamic conditions become favorable for convection within two days of its cessation. After that, dynamical conditions determine when the next active spell commences.

The relationship of the convection over the Bay to SST is complex because SST is maintained above the threshold throughout the summer monsoon (Vinayachandran & Shetye 1991). On the intraseasonal scale, observations from moored buoys (Figure 19) suggest that the SST varies primarily in response to variation in convection (OLR), decreasing in active spells and increasing in cloud-free conditions (Premkumar et al. 2000). BOBMEX observations also showed that SST decreased during active phases of convection (Bhat 2001). Sengupta & Ravichandran (2001) found that the net surface heat flux in active phases of convection over the Bay is large and negative (about 80 wm^{-2}) and is large and positive (about 100 wm^{-2}) when convection is suppressed. The rate of increase (decrease) in SST during suppressed (active) phases is consistent with these changes in the surface heat flux. However, from direct measurements of air-sea fluxes made

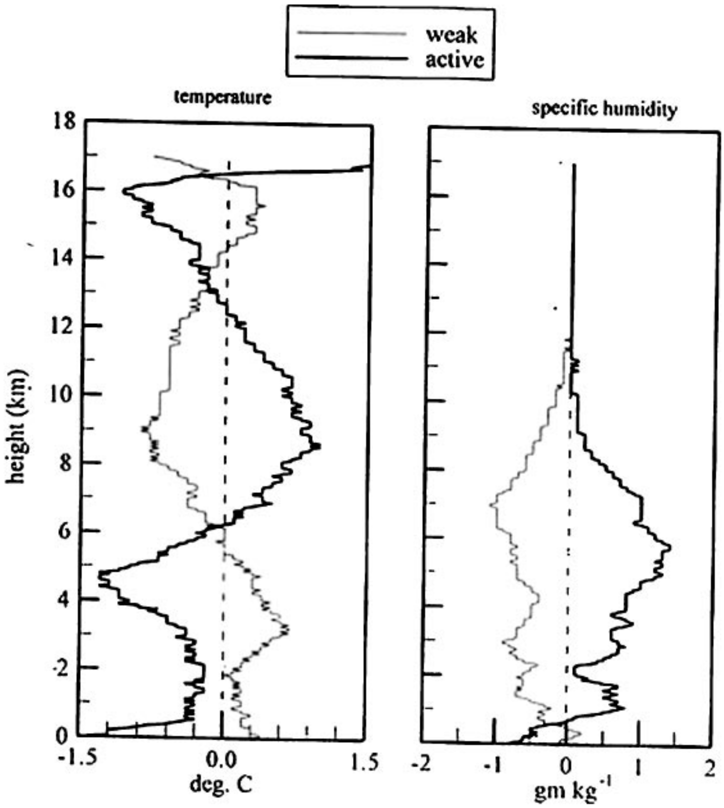


Figure 18 Composite vertical profiles of temperature and specific humidity anomalies for active and weak spells over the Bay of Bengal during BOBMEX (after Bhat et al. 2002).

during BOBMEX, Bhat (2002) showed that the observed variation in SST during part of the period could not be explained without invoking horizontal advection. Thus, dynamics of the circulation of the Bay (Shankar et al. 2002) is also important in determining the temporal variation of the SST. Bhat (2002) also showed that the air-sea gradients in temperature and specific humidity over the north Bay are significantly smaller than over other tropical oceans, including the tropical western Pacific, which has very important implications for estimation of air-sea fluxes.

The special characteristics of the thermohaline structure of the northern Bay with very low salinity of the upper layer, strong halocline at the base of the surface mixed layer, and a barrier layer that persists throughout the monsoon (Bhat et al. 2001) have a large impact on the nature of the air-sea interaction. During 1999, at the BOBMEX measurement site in northern Bay, the formation of the barrier

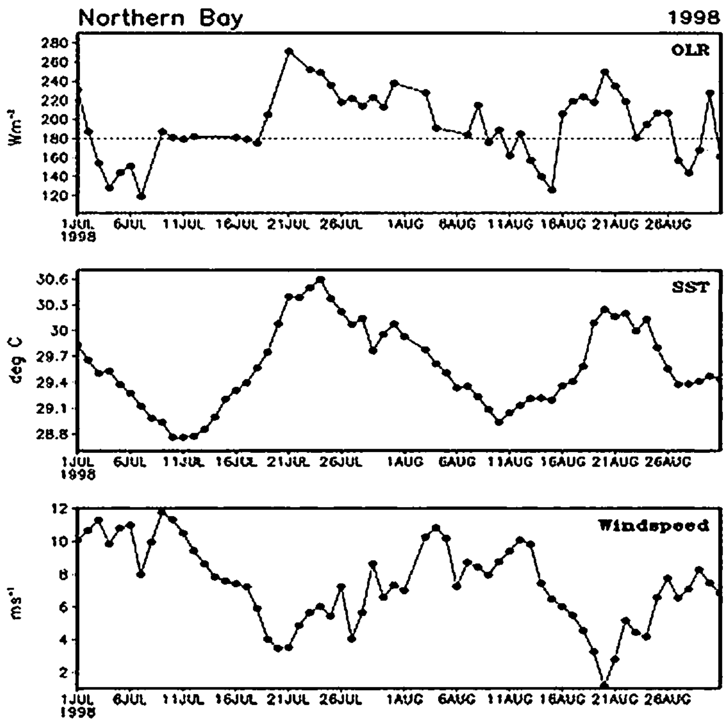


Figure 19 Variation of OLR, windstress, and SST (from moored buoy) in the Bay of Bengal during summer monsoon of 1998 (after Premkumar et al. 2000).

layer occurred with the arrival of a freshwater plume whose source was traced to freshwater from river discharge and rainfall (Vinayachandran et al. 2002).

It is important to note that there is considerable interannual variation in the amplitude of variation of SST of the Bay of Bengal. For example, whereas the amplitude over the northern Bay was about 3°C in July–August 1998, in the same period in 1999, it was less than 1°C (Bhat et al. 2001). During the season of 1999, several disturbances occurred, and the OLR was low for a large number of days. The prolonged active phases may have been the key factor leading to the smaller amplitude of SST variation. Further investigations are required to understand the interannual variation and its links to intraseasonal variation.

The second experiment under ICRP—the Arabian sea and the monsoon experiment (ARMEX)—aimed at studying the convective systems over the Arabian sea, which are responsible for intense rainfall events over the west coast of the Indian peninsula (Department of Science and Technology 2001) was conducted during June–July 2002. It is expected that analysis of new observations from the observational experiments under ICRP, as well as international programs such as JASMINE conducted during the premonsoon season of 1999 (Webster et al. 2002), will provide new insights into the nature of the coupling of the monsoon to the ocean.

In the past two decades, considerable insight into monsoon variability has been gained, with elucidation of important features on intraseasonal and interannual scales, with analysis of data from conventional observatories, satellites, and special observational experiments on one hand and studies of underlying mechanisms with models on the other. However, simulation of the rainfall patterns over the Indian region by GCMs on the seasonal scale and its variation on intraseasonal and interannual scales remains a challenging problem despite the recent rapid developments in GCMs. It is, therefore, not surprising that the major drought of the season of 2002 could not be predicted by models (Gadgil et al. 2002).

I hope that the phenomenal success in understanding the interactions between the tropical Pacific and the atmosphere and unraveling the dynamics of ENSO (Philander 1990, Anderson et al. 1998) will be followed by conquering this next frontier in tropical oceans and global atmosphere in the near future.

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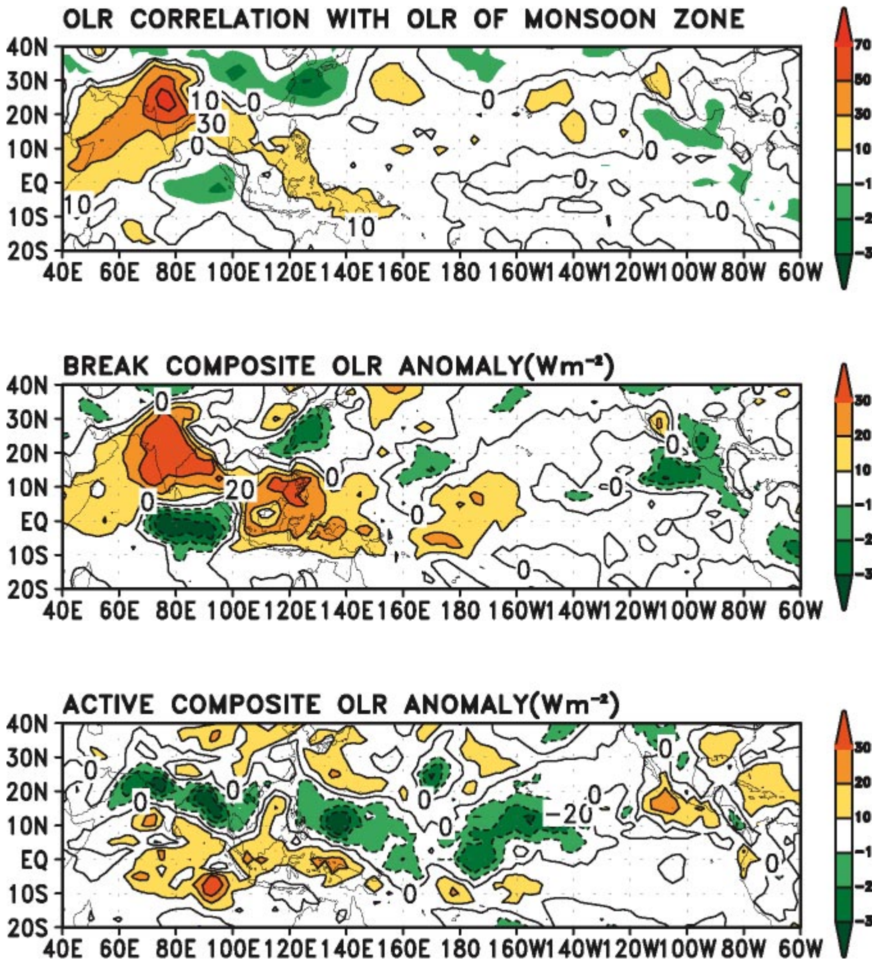

Figure 11 (*a*) Correlation of three-day running mean OLR with respect to that of the OLR of the monsoon zone (Figure 4*a*) for July–August. (*b*) Break composite OLR anomaly patterns (after Gadgil & Joseph 2003). (*c*) Active composite patterns of OLR and OLR anomaly patterns (after Gadgil & Joseph 2003).

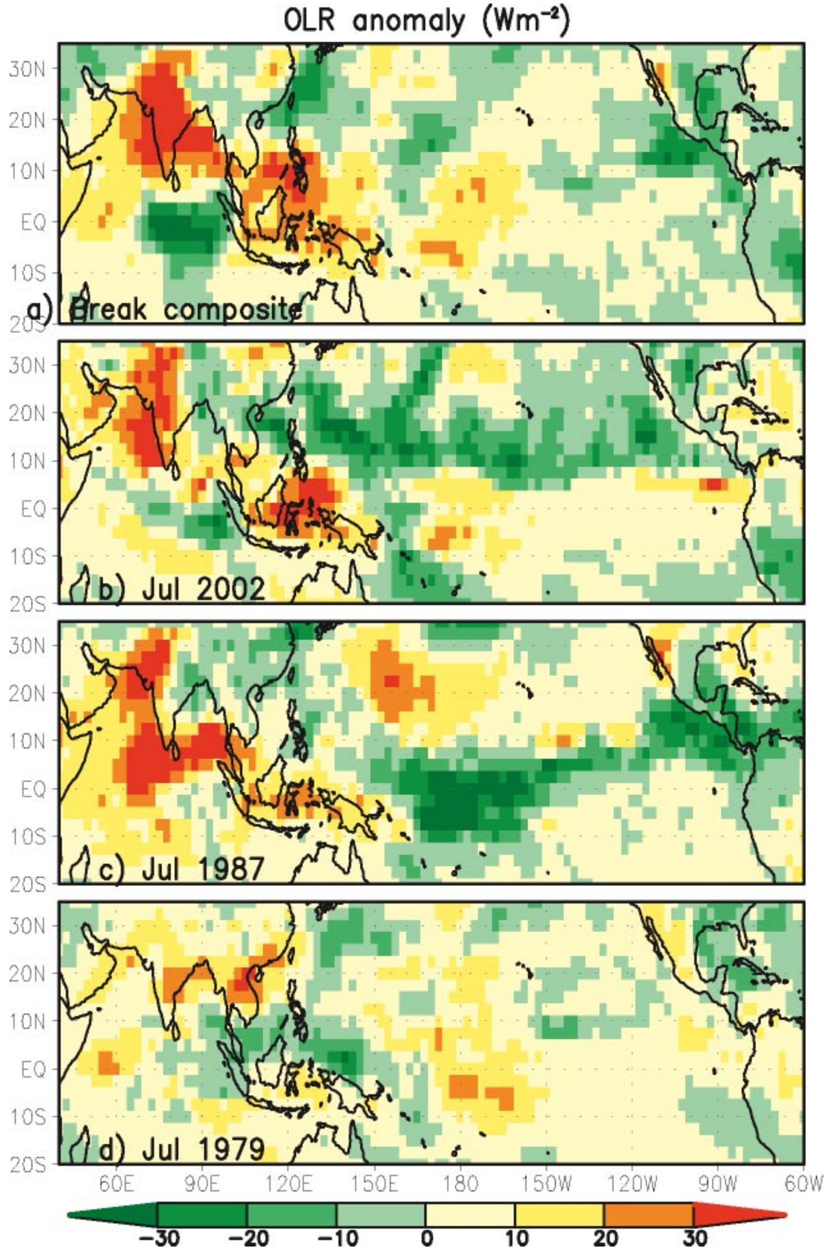

Figure 13 OLR anomaly patterns for (*a*) the break composite, (*b*) July 2002, (*c*) July 1987, and (*d*) July 1979.