Dynamics of Breaks in the Indian Summer Monsoon

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ABSTRACT

In this paper the authors present results of diagnostic analysis of observations and complementary experiments with a simple numerical model that enable them to synthesize the morphology and dynamics of “breaks” in the Indian summer monsoon (ISM). Almost one week ahead of the onset of a break spell over India, a monotonically decreasing trend in convective activity is found to occur over the Bay of Bengal in response to a steady eastward spreading of dry convectively stable anomalies from the equatorial Indian Ocean. A major intensification of the convectively stable anomalies over the Bay of Bengal is seen about 2–3 days prior to commencement of a monsoon break. Both observations and modeling experiments reveal that rapid northwest propagating Rossby waves are triggered in response to such a large strengthening of the convectively stable anomalies. It is shown that an abrupt movement of anomalous Rossby waves from the Bay of Bengal into northwest and central India marks the initiation of a break monsoon spell. Typically the Rossby waves are found to traverse from the central Bay of Bengal to northwest India in about 2–3 days’ time. With the establishment of a break phase, the eastward spreading low-latitude anomaly decouples from the rapid northwest propagating anomaly. This decoupling effect paves the way for the emergence of a convectively unstable anomaly over the equatorial Indian Ocean. It is proposed that the dynamics of the rapid northwest propagating anomalous Rossby waves from the central Bay of Bengal toward northwest India and decoupling of the eastward propagating anomaly are two extremely vital elements that determine the transition from an above normal phase to a break phase of the ISM and also help maintain the mutual competition between convection over the Indian subcontinent and that over the equatorial Indian Ocean. Through modeling experiments it is demonstrated that low-latitude Rossby wave dynamics in the presence of a monsoon basic flow, which is driven by a steady north–south differential heating, is a primary physical mechanism that controls the so-called monsoon breaks.

1. Introduction

On the subseasonal timescale, the Indian summer monsoon (ISM) undergoes periods of enhanced and reduced rainfall activity over a large region in central and north India and these intraseasonal variations are termed “active” and “break” monsoon phases. During an inactive or break phase, the monsoon trough is found to shift northward from its normal position to the foothills of the Himalaya (Ramamurthy 1969; Rao 1976) resulting in above normal pressures, of the order of 4 hPa, over northern and central parts of the country. Associated with the northward shift of the monsoon trough, there is excess rainfall over the foothills of the Himalaya and northeast India during a break monsoon phase. Additionally, the rainfall activity increases over south–eastern peninsular India during break spells. An exhaustive survey of the observed characteristics of breaks in the Indian summer monsoon was carried out by Ramamurthy (1969) using 80 years (1888–1967) of Indian rainfall data. He noted that most of the breaks, in general, had a duration of about 3–5 days and occurred most frequently during the month of August. He also identified very long break epochs that lasted for 17–20 days. Our specific interest to understand monsoon

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breaks stems from the fact that break spells can indeed significantly contribute to droughts in the region. Without breaks, the monsoon can turn out to be a nearly continuous rainy season as it is sometimes. Based on this thinking, we feel that it is important to understand breaks in the monsoon rather than treating active and break spells as simply opposite phases of the same intraseasonal variability.

a. Intraseasonal variations of the summer monsoon

Numerous observational studies during the past few decades have been of considerable use in understanding active and break monsoon spells in terms of subseasonal oscillations of the summer monsoon. The early works of Keshavamurty (1973) and Murakami (1977) indicated the existence of a 10–20-day oscillation in the meridional wind data over north Indian stations. Krishnamurti et al. (1973) for the first time pointed out that the spectrum of the Tibetan high showed a dominant periodicity around 13 days. Krishnamurti and Bhalme (1976) made a substantial contribution by identifying a quasi-biweekly oscillation in most of the salient elements of the summer monsoon system. They postulated that the alternations between active and break spells could result from cloud feedback and large-scale radiative effects. Fluctuations between active and break monsoon spells on a quasi-biweekly timescale have also been noted by Murakami and Frydrych (1974) and Murakami (1976).

The existence of longer period oscillations of about 30 days over the summer monsoon region was noted by Dakshinamurty and Keshavamurty (1976). It was Yasunari (1979, 1980) who first emphasized the significance of meridionally propagating cloud bands, over the Indian subcontinent, on the timescale of 30–50 days. Similar features were also identified by Sikka and Gadgil (1980) based on satellite cloudiness data. Krishnamurti and Subrahmanyan (1982) showed that the circulation anomalies on this timescale manifest in the form of meridionally propagating trains of lower-tropospheric cyclonic and anticyclonic flow patterns from the equatorial regions toward the Himalaya. They observed that the passage of these cyclonic and anticyclonic wind anomalies coincided with the occurrence of wet and dry spells over India and neighboring Indochina suggesting a strong coherence between the wind oscillations and the precipitating patterns. Hartmann and Michelsen (1989) noted a spectral peak with periodicity around 40 days by examining 70 yr of daily precipitation data over the Indian subcontinent. In a comprehensive review, Madden and Julian (1994) have discussed several other important studies related to 40–50-day fluctuations over the summer monsoon region.

b. Observations of monsoon breaks

In an important paper, Ramaswamy (1962) noted that breaks in the ISM were influenced by the intrusion of midlatitude troughs into the Indo–Pakistan region in the middle and upper troposphere. He observed that such an intrusion of large-amplitude extratropical Rossby waves can weaken the Tibetan high and the upper-level easterlies. Ramaswamy and Pareek (1978) further noted that the atmospheric general circulation in both the hemispheres was locked in a low-index Rossby regime during intensely weak phases of the summer monsoon. Unninayar and Murakami (1978) have noted a bifurcation of the Tibetan high under the influence of a midlatitude trough near the Indo–Pakistan region during weak monsoon periods. Another prominent change associated with weak monsoons is the occurrence of a stagnant blocking ridge in the upper troposphere between 90° and 115°E over East Asia as noted by Raman and Rao (1981). Ramamurthy (1969) carefully cataloged the breaks in July and August from 1888 to 1967. The criterion for break days that he adopted was to identify persistent synoptic patterns, of at least two-days’ length, in which the monsoon trough was absent in the sea level chart as well as up to 850 hPa. His analysis showed that the month of August was more susceptible to breaks. From a systematic analysis of daily surface pressure data for a 40-yr period (1933–1972), Krishnamurti and Ardanuy (1980) found that break monsoon spells were associated with westward propagating trough–ridge systems that had a periodicity of about 10–20 days. They found that a large contribution to this quasi-biweekly mode was from zonal wave-numbers 3–6. They also noted that extrapolation of the phase information of the steady westward propagating quasi-biweekly mode had fairly good skill in predicting monsoon breaks.

c. Theoretical and modeling studies

There have been quite a few important theoretical studies on the mechanisms of intraseasonal oscillations over the summer monsoon region. Webster (1983) was able to simulate a northward propagating biweekly oscillation, using a zonally symmetric model that he attributed to land surface hydrological feedback. Goswami and Shukla (1984) showed that an interactive feedback between moist convection and the monsoon circulation is important for the generation of intraseasonal oscillations over the subcontinent. The studies Lau and Peng (1990) and Krishnan and Kasture (1996) suggest that a convective feedback between the monsoon large-scale flow and the equatorial 30–50-day oscillation can trigger northward propagating modes over the monsoon region. However, there are relatively few modeling attempts that have explicitly addressed the dynamical causes for monsoon breaks. In a recent work, Rodwell (1997) showed that breaks in the Indian summer monsoon could be triggered by extratropical weather systems in the Southern Hemisphere (SH). He noted that the passage of ridges over the eastern coast of South Africa resulted in injection of dry, high negative po-
tential vorticity (PV) air from the SH midlatitudes into
the monsoon low-level flow. He argued that the effect
of this change in PV causes the low-level air to turn in
such a way so as to “avoid” India. The central issue
which Rodwell (1997) emphasized was on the modifi-
cation in the trajectory of the low-level monsoon flow
resulting due to movement of high pressure weather
systems in the SH extratropics.

d. Objectives of the present study

Despite the above-mentioned studies, there is still a
major need to understand the basic dynamical mecha-
nism that controls the transition from an active to a break
phase of the ISM. The previous works of Yasunari
(1979) and Sikka and Gadgil (1980) had reported that
intraseasonal variations of the ISM are linked to the
occurrence of a mutual competition between convection
over the Indian subcontinent and that over the equatorial
Indian Ocean. However, the dynamical details of how
such a competitive interaction explicitly operates are
still not adequately clear. The specific question of how
the dynamics of transition from an active phase to a
break phase of the summer monsoon is connected with
the competition between the two convection zones still
needs to be resolved.

The primary objective of this study is to improve our
current understanding of the dynamical mechanism that
controls the evolution of monsoon breaks and the com-
petition between the two convection zones. A twofold
strategy, of combining data analysis and simple mod-
ing experiments, is adopted in this study. The data
diagnostic examination involves a careful and detailed
analysis of monsoon breaks using 17 yr (1979–95) of
daily observed outgoing longwave radiation (OLR) and
daily wind data from National Center for Environmental
Prediction–National Center for Atmospheric Research
(NCEP–NCAR) Reanalysis (see Kalnay et al. 1996).
One purpose of the data analysis is to understand wheth-
er breaks in the regional scale ISM are associated with
any preferred large-scale pattern of convection and cir-
culation anomalies. Another reason for carrying out the
data analysis is to identify major synoptic-scale features
that are relevant for tracking the spatiotemporal evolu-
tion of breaks in the ISM. It should be pointed out
that significant strength for this paper comes from the
availability of a fairly long-period data of daily observed
OLR and wind fields, which have allowed us to objec-
tively determine the evolution of break monsoon anom-
alies. The data analysis component is supplemented by
experiments using a shallow-water model so as to pro-
mote our understanding of the dynamics of monsoon
breaks.

2. Data diagnostics

a. Identification of break monsoon days

The spatial pattern of rainfall anomaly during mon-
soon breaks, obtained by Ramamurthy (1969), is an ex-
tremely useful guiding tool for identifying break spells.
The mean percentage departure of rainfall during breaks
obtained by Ramamurthy (1969) is shown in Fig. 1.
This diagram has also been presented in a paper by
Gadgil and Asha (1992). The most conspicuous feature
in Fig. 1 is the large negative rainfall anomaly over the
plains of northwest and central India. The positive rain-
fall anomaly over northeast India is in association with
the northward shift of the monsoon trough to the foot-
hills of the Himalaya during break spells. There is also
a small pocket of positive rainfall anomaly over Tam-
ilnadu in southeastern peninsular India.

Given the inhomogenous nature of the Indian mon-
soon rainfall and the existence of a wide range of spatial
and temporal scales, the process of determining a unified
break monsoon criterion is in fact quite involved. Also
the availability of long-period precipitation data on a
daily basis, both over land and ocean, has been a serious
problem. Hence, we had to rely on the daily OLR ob-
servations in order to study the spatial and temporal
evolution of convective anomalies associated with mon-
soon breaks. Daily OLR data measured from Advanced
Very High Resolution Radiometers aboard National
Oceanic and Atmospheric Administration polar orbiting
spacecraft (see Gruber and Krueger 1984) are a vital
source of information for studying intraseasonal vari-
ability of tropical convection. Deep convection in the
Tropics is characterized by low cloud-top temperatures
and small OLR values. Likewise, regions having large
OLR values indicate scarcity or absence of cloud cover.
Thus OLR is a good proxy for inferring the rainfall
activity associated with tropical convection. Usually
weekly or monthly averaged OLR data are considered
a good surrogate for rainfall. However, considering the
fact that monsoon breaks represent an intensely sup-
pressed phase of convection across a wide region, during
the core rainfall season, it is quite plausible to make use of the daily OLR data in order to detect these robust anomalies. In other words, the identification of monsoon breaks from the daily OLR data is greatly facilitated by the conspicuous nature of the break anomalies. It will be seen later that the break monsoon convective anomalies identified using daily OLR data are remarkably consistent with the wind anomalies obtained independently from the NCEP reanalysis. The basic approach adopted for identifying break days is to isolate consecutive days in the summer season having high OLR values (suppressed convection) over a broad region of northwest and central India that is representative of the area of negative rainfall anomaly obtained by Ramamurthy (1969) (see Fig. 1). The details of the identification procedure are outlined below.

- The first step is to construct a daily climatological OLR time series, using 17 yr (1979–95) of daily OLR data. This daily climatological time series is used for computing the daily anomalies during each year. It is realistic to calculate daily anomalies by subtracting the daily climatology instead of the monthly mean climatology because of the strong seasonality seen over the Indian subcontinent from May to September. We have verified that this daily climatological OLR data captures the annual cycle of tropical convection quite realistically.

- We focus our inspection of monsoon breaks during the core period from 15 June to 15 September. It was felt that early June and late September were not suitable for inferring breaks because ambiguities such as delayed onset and early withdrawal of the summer monsoon can often lead to misinterpretation of breaks in the monsoon. In order to minimize such conflicts, we restrict the analysis of monsoon breaks from 15 June to 15 September for each year.

- In the next step we isolate periods having significantly large positive OLR anomalies for at least four consecutive days or more over a wide region covering northwest and central India. In addition to isolating the days of suppressed convection, we apply an objective criterion that the OLR anomaly averaged over (18°–28°N, 73–82°E) should exceed a threshold value of +10 W m⁻² during all the days of a break period.

- We noted that the magnitude of OLR anomaly exhibits variations from one break epoch to another. For instance, the break anomalies during severe droughts years (e.g., 1982 and 1987) were considerably stronger than during normal monsoon years. Even within a single break epoch, the magnitude of OLR anomalies varies from one day to another. Based on all these considerations, the cutoff value of +10 W m⁻² was decided.

- The onset and withdrawal dates of breaks correspond to the starting and ending days of each break spell.

By applying the above criteria, we could clearly identify 25 break epochs during the 17 yr (1979–95). The total number of break days for these 25 cases is 191. The 25 break epochs are listed in Table 1. More than one break spell was seen, from our analysis, during 1979, 1985, 1987, 1991, 1993, and 1995. Although during 1988, there were a few weak monsoon spells of duration less than 4 days, we could not identify any major break monsoon signal. Also from Table 1, it can be noted that one of the break spells of 1991 commenced on 14 June. For the purpose of continuity, this exception had to be retained in our analysis although it was 1 day before the border date of 15 June. It is verified that this break spell is not a case of delayed onset of summer monsoon during 1991. It should be pointed out that the procedure adopted in our analysis, for identifying break spells, is rather stringent. We have excluded those cases that had even a small degree of ambiguity. Therefore, it is quite possible, that a few breaks might have been left out. However the 25 cases (191 days) that have been selected seem to be quite reliable. Moreover, these 25 break cases (191 break days) represent a sizeable sample for making statistically significant inferences.

### Table 1. List of break monsoon days identified.

<table>
<thead>
<tr>
<th>Case</th>
<th>Year</th>
<th>Break period</th>
<th>No. of days</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1979</td>
<td>1 Jul–6 Jul</td>
<td>6</td>
</tr>
<tr>
<td>2</td>
<td>1979</td>
<td>18 Jul–22 Jul</td>
<td>5</td>
</tr>
<tr>
<td>3</td>
<td>1979</td>
<td>12 Aug–27 Aug</td>
<td>16</td>
</tr>
<tr>
<td>4</td>
<td>1979</td>
<td>2 Sep–7 Sep</td>
<td>6</td>
</tr>
<tr>
<td>5</td>
<td>1980</td>
<td>7 Sep–10 Sep</td>
<td>4</td>
</tr>
<tr>
<td>6</td>
<td>1981</td>
<td>23 Aug–31 Aug</td>
<td>9</td>
</tr>
<tr>
<td>7</td>
<td>1982</td>
<td>24 Jun–7 Jul</td>
<td>14</td>
</tr>
<tr>
<td>8</td>
<td>1983</td>
<td>4 Jul–8 Jul</td>
<td>5</td>
</tr>
<tr>
<td>9</td>
<td>1984</td>
<td>19 Jul–22 Jul</td>
<td>4</td>
</tr>
<tr>
<td>10</td>
<td>1985</td>
<td>29 Jun–3 Jul</td>
<td>5</td>
</tr>
<tr>
<td>11</td>
<td>1985</td>
<td>23 Aug–26 Aug</td>
<td>4</td>
</tr>
<tr>
<td>12</td>
<td>1985</td>
<td>5 Sep–8 Sep</td>
<td>4</td>
</tr>
<tr>
<td>13</td>
<td>1986</td>
<td>20 Aug–8 Sep</td>
<td>20</td>
</tr>
<tr>
<td>14</td>
<td>1987</td>
<td>14 Jul–3 Aug</td>
<td>21</td>
</tr>
<tr>
<td>15</td>
<td>1987</td>
<td>30 Aug–2 Sep</td>
<td>4</td>
</tr>
<tr>
<td>16</td>
<td>1988</td>
<td>Not identified</td>
<td></td>
</tr>
<tr>
<td>17</td>
<td>1989</td>
<td>2 Sep–8 Sep</td>
<td>7</td>
</tr>
<tr>
<td>18</td>
<td>1990</td>
<td>26 Jul–29 Jul</td>
<td>4</td>
</tr>
<tr>
<td>19</td>
<td>1991</td>
<td>14 Jun–21 Jun</td>
<td>8</td>
</tr>
<tr>
<td>20</td>
<td>1991</td>
<td>7 Sep–11 Sep</td>
<td>5</td>
</tr>
<tr>
<td>21</td>
<td>1992</td>
<td>23 Jun–9 Jul</td>
<td>15</td>
</tr>
<tr>
<td>22</td>
<td>1993</td>
<td>7 Aug–13 Aug</td>
<td>7</td>
</tr>
<tr>
<td>24</td>
<td>1994</td>
<td>7 Sep–10 Sep</td>
<td>4</td>
</tr>
<tr>
<td>25</td>
<td>1995</td>
<td>2 Jul–5 Jul</td>
<td>4</td>
</tr>
<tr>
<td>26</td>
<td>1995</td>
<td>14 Aug–17 Aug</td>
<td>4</td>
</tr>
</tbody>
</table>

Total = 191

### b. Spatial pattern of break anomalies

An examination of the spatial distribution of anomalous features can provide useful insight about possible linkages between monsoon breaks and anomalies located over other regions. For this purpose, we shall pre-
pare composites of OLR and wind anomalies from the 191 break days presented in Table 1.

1) Convection Anomalies

A composite plot of OLR anomalies during monsoon breaks is shown in Fig. 2a. Regions of suppressed (enhanced) convective activity can be inferred from the positive (negative) anomalies that correspond to high (low) cloud-top temperatures associated with a scarcity (abundance) of convective clouds. The four dominant regions of anomalous convection in Fig. 2a are described below.

- The large positive OLR anomaly over India, which has a maximum value of about 40 W m$^{-2}$, is centered around (24$^\circ$N, 76$^\circ$E). This anomalous feature strikingly represents the strong suppression of monsoon activity over India during breaks. This positive OLR anomaly extends westward into a broad region covering Pakistan, the Arabian Sea, parts of Arabia, and equatorial Africa.

- On the eastern side the positive OLR anomaly extends into the Bay of Bengal, southeast Asia, the South China Sea, and the equatorial western Pacific. One can notice widespread suppression of convection over Burma, Vietnam, Thailand, the Philippines, and Indonesia indicating that the convective anomaly over Southeast Asia is in phase with that over the Indian subcontinent.

- Another important feature is the negative OLR anomaly over the equatorial Indian Ocean centered around (3$^\circ$S, 85$^\circ$E) with a maximum amplitude of $-10$ W m$^{-2}$. The fact that convection over the equatorial Indian Ocean starts dominating the continental convection during monsoon breaks is consistent with the findings of Yasunari (1979) and Sikka and Gadgil (1980). Also important is a weak negative anomaly over the southern tip of Africa close to the Mascarene high. The negative anomaly over this region indicates a weakening of the Mascarene high resulting from a weaker monsoon Hadley circulation during the break phase (see Krishnamurti and Bhalme 1976).

- The last region of strong convection anomaly is located over eastern China, the western North Pacific Ocean, and southern Japan. The opposite polarity of the anomaly indicates that the convective activity in this region is out of phase with that over India. This suggests that the summertime quasi-stationary fronts over China and Japan (e.g., Baiu fronts during June and July) can intensify during breaks in the Indian monsoon.

Additionally one can note a slight enhancement of convection over the highlands of central Asia (40$^\circ$N, 75$^\circ$E). The tiny region of enhanced convection anomaly over northeast India in Fig. 2a, which is apparently related to the northward shift of the monsoon trough, can be seen prominently during the initial few days of a break spell. Because of the uneven time duration of each of the individual 25 break epochs, the compositcd anomalous feature over northeast India appears to be weaker in Fig. 2a. It is verified that this anomaly is more pronounced if the composite is obtained by uniformly averaging only the first 4-5 days of each individual break case. The coarse resolution of the OLR data may be a reason for the enhanced convection not being seen over eastern peninsular India in Fig. 2a. Usually a dense network of station rainfall data is used in the construction of rainfall anomalies such as in Fig. 1. Since the 2.5° × 2.5° gridded OLR data are rather smooth, it will not be possible to resolve finer break monsoon features specifically such as those over eastern peninsular India. Far from the monsoon region, we noted weak negative OLR anomalies over the eastern equatorial Pacific Ocean, Mexico, and Central America (not shown in Fig. 2a). These remote anomalies are due to the shift of the tropical convection anomalies to the central and eastern Pacific Ocean during the well-known El Niño years (e.g., 1982, 1987, 1992) that are included in the break monsoon composite. Further we have examined the short (4-9 days) and long (≥10 days) duration breaks separately. It is seen that both the short and long breaks exhibit the four major regions of convective anomalies shown in Fig. 2a. A composite of all short breaks shows that suppressed convective anomalies prevail over northwest India for over a week following the break onset. However for the long breaks, the condition of suppressed convection over northwest India continues for more than 2 weeks after day 0. Also, the convective anomalies associated with long breaks appear to be relatively more intense than the short duration breaks. An understanding as to why some of the breaks tend to extend for a significantly longer time needs a separate study. It is interesting to note from Table 1 that some of the long breaks have occurred during El Niño years (e.g., 1982, 1987, 1992). However it is not clear to us at the moment how warm ENSO conditions in the Pacific can lead to prolonged break situations over northwest and central India. A full-fledged study of the comparative evolution of short and long breaks involves separate discussions, which are beyond the scope of the present paper.

On the whole, it should be pointed out that the above spatial structure of convection anomalies during monsoon breaks is quite consistent with the spatial pattern of correlation coefficients of cloudiness with reference to a point over central India obtained by Yasunari (1979) and also with the second empirical orthogonal function obtained by Lau (1992). The correlation pattern of Yasunari (1979) was based on daily cloudiness data during 1973 for the period (1 June–30 September). However, it is important to note that the method of compositing used in our study not only helps in inferring about the spatial distribution of the anomalies but also provides quantitative estimates of the anomalies. Such an objective determination of both the structure and amplitude
Fig. 2. (a) Break composite of OLR anomalies; contour interval is 7 W m$^{-2}$, negative values are shaded, and zero contour is suppressed. (b) Statistical levels of significance; only 1% and 5% significant levels are shown. (c) Break composite of wind anomalies at 850 hPa; scale of the vector arrow is 2 m s$^{-1}$. (d) Break composite of wind anomalies at 200 hPa; scale of the vector arrow is 4 m s$^{-1}$. 
of convection anomalies during monsoon breaks was possible because of the availability and use of 17 yr of daily OLR data in this study.

2) Statistical test

The statistical significance of the composited break anomaly obtained from the 191 days is evaluated using the Student’s t-test for unequal variances (see Press et al. 1992). The null hypothesis assumes that the composited break anomaly is not statistically significant. In order to test the null hypothesis we first compute, at each grid point, the test statistic given below:

\[
t = \frac{1}{n_1} \sum_{i=1}^{n_1} a_i - \frac{1}{n_2} \sum_{i=1}^{n_2} c_i \times \left(\frac{\sigma_1^2}{n_1} + \frac{\sigma_2^2}{n_2}\right)^{1/2},
\]

where \(a_i\) are the OLR anomalies for the 191 break days and \(c_i\) are daily OLR anomalies of 94 days (14 June–15 September) for each of the 17 yr. Therefore, \(n_1 = 191\) and \(n_2 = 1598\). The value of the term \((1/n_2) \Sigma_{i=1}^{n_2} c_i\) is zero, since it represents the mean of daily anomalies for the entire time period. Here \(\sigma_1\) and \(\sigma_2\) correspond to standard deviations of \(a_i\) and \(c_i\), respectively. The number of degrees of freedom for the statistic is given by

\[
\nu = \frac{\left[\frac{\sigma_1^2}{n_1} + \frac{\sigma_2^2}{n_2}\right]^2}{\frac{1}{n_1 - 1}\left[\frac{\sigma_1^2}{n_1}\right]^2 + \frac{1}{n_2 - 1}\left[\frac{\sigma_2^2}{n_2}\right]^2}.
\]

To reject the null hypothesis, that the break anomaly in Fig. 2a does not represent a significant change, the computed test statistic should exceed the appropriate quantity for any specified level of significance, which (quantity) can be found from standard tables.

Contour maps of significance levels (only 1% and 5% levels) computed from the t-test for the break OLR anomalies are shown in Fig. 2b. Areas enclosed by the inner and outer contours correspond to 1% and 5% significance levels, respectively. It is clearly evident from Fig. 2b that the break anomalies located over the four primary regions are indeed statistically significant. As noted by Livezey and Chen (1983) and Wilks (1995), we do realize the limitations of the above method for testing the statistical significance of multidimensional fields. In particular, it is realized that lack of independence among individual observations can sometimes lead to large variance in the data. However, the statistical significance shown in Fig. 2b may not be affected by these problems in view of the large sample size used in our study. Moreover, we shall use the above significance map only for reference purpose and while interpreting it we try to recognize the overall pattern and keep in mind the dynamical processes involved.

3) Wind anomalies

The composited wind anomalies at 850 and 200 hPa for the 191 break days are shown in Figs. 2c and 2d, respectively. The wind anomalies are computed using 17 yr (1979–95) of daily NCEP–NCAR reanalysis data. Just like the OLR anomalies, the wind anomalies are constructed by subtracting the daily climatological values from the selected 191 break days. From a first glance, one can notice that the wind anomalies are highly consistent with the OLR anomalies. This point is very important because the satellite estimates of OLR and the wind data from NCEP Reanalysis are based on two completely independent data sources. It may be recalled that the break days were identified from the OLR data alone. The fact that there is a high degree of consistency between the OLR and wind composites is an indication about the robustness and high reliability of the break days selected for this study.

The salient features in Fig. 2c are the prominent low-level anticyclonic wind anomaly over India; anomalous low-level easterlies over the Arabian Sea; a weak trough over the foothills of the Himalaya; a cyclonic shear in the equatorial Indian Ocean; a low-level cyclonic anomaly over eastern China, the northwest Pacific Ocean, and southern Japan; an anticyclonic anomaly over Southeast Asia; and a weak cyclonic circulation over southern Africa. The break monsoon wind anomalies over India match well with the results of Alexander et al. (1978). The low-level clockwise wind anomaly over India and the anomalous easterlies in the Arabian Sea are suggestive of a major reduction in the moisture transport into the monsoon region.

In the upper troposphere (Fig. 2d), the strong southerly wind anomalies over India indicate a weakening of the tropical easterly jet. The Tibetan high is displaced eastward from its normal position and is located over eastern China. This shift of the Tibetan anticyclone during inactive periods of monsoon has been noted in observations (see Krishnamurti et. al 1989; Raman and Rao 1981). Due to this eastward shift from its normal position, one can see a continued presence of westerly wind anomalies over the north of India. Such southward penetrating westerly troughs bring dry extratropical air into the tropical troposphere, which is adverse for the monsoon activity (see Ramaswamy 1962). The southward intrusion of extratropical troughs can also be observed in the lower levels (Fig. 2c) over Indo–Pakistan. Over the central and eastern Pacific basin (not shown in Fig. 2) anomalous equatorial westerlies (easterlies) were seen in the lower (upper) troposphere, which indicated a reduction in the intensity of the east–west divergent circulations and a weakening of the trade wind system that is typical of weak monsoons (Kanamitsu and Krishnamurti 1978). An examination of the statis-
tical significance of the zonal wind anomalies for the 191 break days revealed that the circulation changes over the four major anomalous regions were statistically significant (figures not shown) and consistent with the results described in Fig. 2b.

c. Space±time evolution of anomalies

To gain deeper insight about monsoon breaks, it is necessary to understand the evolution of convection and circulation features preceding, during, and succeeding the break epochs. For this purpose, we shall construct a time sequence of OLR and wind composites by fixing the day of commencement of a break spell (i.e., day 0) as the reference time. The day 0 composite for any variable is prepared by averaging the day 0 values from all the 25 break cases. Likewise, one can construct composites corresponding to 1, 2, 3, . . . , n days before (after) day 0—through backward (forward) shifting of data by 1, 2, 3, . . . , n days with respect to (w.r.t.) day 0 for each of the 25 break cases.

In the previous section, we had identified core regions of convection and circulation anomalies during monsoon breaks. We will now examine the time evolution of spatially averaged OLR anomalies over each of these primary zones. The time sequence of convection changes over the Indian subcontinent is shown in Fig. 3a. The continental convection over northwest and central India is fairly active prior to day 0, as shown by the negative OLR anomalies in Fig. 3a. The area-averaged OLR anomaly reaches a lowest value around $-9 \text{ W m}^{-2}$ about two days before the commencement of a break. Following this stage, the onset of break is heralded by a sharp decrease in convective activity. In response to such a dramatic change, the area-averaged OLR anomaly reaches a maximum value of about $+27 \text{ W m}^{-2}$ on day 1. The break anomaly is well maintained for the next few days and later it gradually weakens in intensity. Although the duration of breaks varies from one case to another, the condition of suppressed convective activity generally persists as long as one week and sometimes even beyond.

The time evolution of convection over the equatorial Indian Ocean is shown in Fig. 3b. It can be seen that convection over this region remains suppressed from day $-6$ through day $-1$. However in response to the sudden weakening of the monsoon, the anomalous convection over the equatorial Indian Ocean begins to show an abrupt increase from day 0 onward and a maximum value of $-11 \text{ W m}^{-2}$ is attained around day $+6$. Although the convective anomaly keeps fluctuating, a condition of above normal convection persists over the equatorial Indian Ocean during the entire postbreak period. To a first approximation, the time sequence of convection change over the equatorial Indian Ocean behaves in a manner that is the opposite of that over the Indian subcontinent. Apparently, one can notice that the oceanic convection tends to dominate whenever the continental convection remains subdued and vice versa. Such a competitive influence between the two convection zones has since long been recognized as the basis for low-frequency oscillations in the monsoon system (see Yasunari 1979; Sikka and Gadgil 1980).

The sequence of convection changes over eastern China and the northwest Pacific Ocean is shown in Fig. 3c. The positive OLR anomalies from day $-8$ through day $-3$ suggest that convective activity over this region is subdued till day $-3$. The convective activity begins to show an increase from day $-2$ onward and reaches a maximum on day 0. The state of enhanced convection over this region is maintained for more than a week following the onset of break. It will be seen later that changes in this region are induced due to a strengthening of the subtropical westerlies following a weakening of monsoon activity over India.

The convective activity over Southeast Asia (Fig. 3d) shows a slow decreasing trend that can be traced back to nearly 7–8 days before day 0. Above normal convection prevails in this region about two weeks prior to commencement of a break spell and the area-averaged OLR anomaly is less than $-6 \text{ W m}^{-2}$. With the passage of time, the convective activity starts decreasing progressively. The area-averaged OLR anomaly attains a maximum value of about $+6 \text{ W m}^{-2}$ around day $+6$. The positive OLR anomalies continue to persist, over this region, for more than 10 days beyond day 0. The dynamical cause for this steady weakening of convection over Southeast Asia will be discussed subsequently.

Apart from the four primary anomalous regions, we shall also examine the sequence of convection changes over the Bay of Bengal (Fig. 3e). In the subsequent discussions it will become evident that this region is characterized by fast moving transient waves. It can be seen from Fig. 3e that strong convective activity prevails over the Bay of Bengal from day $-15$ to day $-6$. A very prominent decreasing trend in convective activity can be seen from day $-6$ to day $-1$ and the corresponding rate of increase of OLR anomaly is approximately $4.0 \text{ W m}^{-2}$ per day. This abrupt decrease in convective activity over the Bay of Bengal from day $-6$ to day $-1$ could very well turn out to be an useful precursor of monsoon breaks over India. After reaching a maximum value of about $+21 \text{ W m}^{-2}$, the OLR anomaly begins to sharply fall from day 0 onward. Thereafter, one sees small positive OLR anomalies that continue to persist during the postbreak period.

**Propagation characteristics**

Using the sequence of anomaly composites, we shall now examine the propagation characteristics of monsoon breaks. The process of compositing serves as an efficient method for filtering out possible noise that can contaminate the regional-scale features. Instead of plotting the daily maps, we have shown the time sequence
Fig. 3. Composited time evolution of area-averaged OLR anomalies (W m$^{-2}$) before and after breaks: (a) India (16°–28°N, 70°–85°E), (b) the equatorial Indian Ocean (10°S–0°N, 75°–95°E), (c) eastern China and northwest Pacific (20°–30°N, 105°–135°E), (d) Southeast Asia (5°–10°N, 95°–130°E), and the (e) Bay of Bengal (5°–15°N, 80°–95°E).
of 3-day averages (triads) of composited anomalies in Fig. 4. In general, triad $n$ refers to the composited anomaly averaged over (day $3n - 1$, day $3n$, day $3n + 1$). Thus triad 0, which corresponds to the onset of break, refers to the composited anomaly averaged over (day $-1$, day 0, day +1). The triads preceding the break are given by (tria. -1, ..., triad -4) and those following the onset of break are given by (tria. +1, ..., triad +4).

An interesting feature prior to the onset of break is the occurrence of enhanced convection over India and the Bay of Bengal (Figs. 4a–c). It can be seen that this mass of enhanced convection spreads from the Bay of Bengal toward central and northwest India and the increased convection over India continues up to triad $-1$. Thus there are no symptoms of the withdrawal of continental convection even up to triad $-1$. In contrast, one can notice a steady buildup of convectively stable (positive OLR) anomalies over the central Bay of Bengal, Southeast Asia, and the equatorial Indian Ocean from triad $-2$ to triad 0. It appears that the generation of large suppressed convective anomaly over the equatorial eastern Indian Ocean in triad $-2$ may be partly induced (via anomalous subsidence) by the mass of enhanced convection to the north of $10^\circ$N. Another possibility for the development of suppressed convection over the eastern Indian Ocean could be due to an eastward spreading of positive OLR anomalies in the equatorial region. For instance, weak positive OLR anomalies can be seen around $5^\circ$S, $85^\circ$E in triad $-4$, which subsequently spread zonally and intensify over the eastern Indian Ocean in triad $-2$ and triad $-1$. Similarly we have noted that the positive OLR anomaly around $60^\circ$E over the western Indian Ocean (see Figs. 4a,b) spreads slowly eastward. In short, one can notice the manifestation of suppressed convective activity, over the eastern Indian Ocean, the Bay of Bengal, and Southeast Asia, almost one week prior to the onset of a break event. This is consistent with the histograms of convective activity shown in Figs. 3d and 3e.

It can be seen from Figs. 4d and 4e that the onset of break is characterized by an abrupt northwest movement of the convectively stable anomalies from the central Bay of Bengal to northwest India. It is evident that much of the rapidity is associated with the northwest transition from the central Bay of Bengal, while the positive OLR anomaly over the Arabian Sea does not show significant meridional displacement from triad $-1$ to triad 0. The rapid shift of the convectively stable anomalies from the central Bay of Bengal into northwest India takes place in a span of about 2–3 days. One can also observe from Figs. 4e,f a tendency for the eastward spreading convectively stable anomalies to get decoupled from the rapid westward propagating anomalies. Associated with this decoupling effect is the emergence of a convectively unstable (negative OLR) anomaly over the equatorial Indian Ocean in triad $+1$. Subsequently, the negative OLR anomaly over the equatorial Indian Ocean starts intensifying. Following the eastward spreading of positive OLR anomalies in triad $+1$ and triad $+2$, a condition of suppressed convective activity continues to prevail over Southeast Asia. With the establishment of a monsoon break, a weak positive OLR anomaly over the Bay of Bengal connects India and Southeast Asia. Over eastern China and the northwest Pacific Ocean, there is an enhancement of convective activity in response to the development of monsoon break. The intensity of the break monsoon spell over northwest India decreases about a week after day 0. It is seen that the evolution of both the short and long duration breaks share several common characteristics—like the development of suppressed convection over the equatorial Indian Ocean and the Bay of Bengal and a subsequent northwest movement into the Indian subcontinent.

From the above discussions it is clear that a progressive weakening of convective activity over the Bay of Bengal and Southeast Asia, seen almost one week prior to the commencement of a break monsoon spell, is partly related to an eastward spreading of dry convectively stable anomalies over the Indian Ocean. The dynamics of such slow eastward propagating transients can be understood in terms of equatorial Kelvin waves. The results of Chang (1977) suggest that interaction between low-latitude wave dynamics and tropical convective heating can excite equatorial Kelvin waves that explain eastward propagating transients such as the well-known 40–50-day oscillations discovered by Madden and Julian (1971, 1972). However, since the meridional component of the wind anomalies near $60^\circ$–$100^\circ$E (Figs. 5a–c) is not small, we feel that the evolution of the convective anomalies may not be entirely governed by Kelvin wave dynamics but also partly controlled by the cross-equatorial flow associated with the summer monsoon circulation. The fact that the convectively stable anomaly continues to spread eastward even after the break onset accounts for the OLR maximum seen on day $+6$ over Southeast Asia in Fig. 3d. It is possible that the increasing trend in OLR over Southeast Asia could just be part of a longer period (e.g., 40–50 days) oscillation. Likewise, the OLR maximum on day $-1$ over the Bay of Bengal (Fig. 3e) is partly associated with an eastward spreading of the suppressed anomaly. However, the subsequent rapid northwest propagation in triad 0 and the decoupling of the eastward spreading component in triad $+1$ produce a sharp decrease in the OLR value from day 0 to day $+4$ in Fig. 3e.

Consistent features can be noted in the evolution of the composited wind anomalies at 850 hPa (Fig. 5). Triad $-4$ shows moderately strong westerly wind anomalies over central and peninsular India and a cyclonic circulation over the Bay of Bengal. In triad $-3$, the anomalous cyclone over the Bay of Bengal organizes and spreads into central India. In triad $-2$, the low-level cyclonic anomaly is well developed and extends farther westward covering the entire subcontinent. It is interesting to note the initiation of an anomalous ridge to
Fig. 4. Sequence of composited OLR anomalies showing the evolution of monsoon breaks: (a) triad -4, (b) triad -3, (c) triad -2, (d) triad -1, (e) triad 0, (f) triad +1, (g) triad +2, (h) triad +3, and (i) triad +4. Contour interval is 3 W m$^{-2}$ and zero contour is suppressed.
Fig. 5. Same as Fig. 4 except for 850-hPa winds. The scale of the vector arrow is 2 m s$^{-1}$. 
the south of 10°N, in triad −2 over the eastern Indian Ocean is consistent with the suppressed OLR anomaly in that region shown in Fig. 4c. As discussed earlier, it seems that the mass of enhanced convection over the Bay of Bengal and India (Figs. 4a–c) is helpful in triggering this anomalous ridge over the eastern Indian Ocean. Although there is a weakening of the cyclonic anomaly in triad −1, there is still no indication of the onset of the break over India. However a striking aspect in triad −1 is the development of a pair of anomalous anticyclones over the central Bay of Bengal and on the southern side of the equatorial Indian Ocean. One can notice that the anticyclonic pair has a nice symmetric structure w.r.t the equator. The Northern Hemispheric anticyclone is somewhat more intense and well organized than its southern counterpart. The structure of the pair of anomalous anticyclones resembles that of a symmetric Rossby wave (see Matsuno 1966; Gill 1980). The pair of anomalous anticyclones in Fig. 5e can be interpreted as Rossby wave response induced by the large convectively stable anomaly over the Bay of Bengal and Southeast Asia seen in triad −1 of Fig. 4. One can observe a sudden westward movement of the Northern Hemispheric (NH) anticyclone from triad −1 to triad 0, which results in the onset of break over India. The break onset is also characterized by a major intensification and intrusion of subtropical westerlies over north India and east Asia in triad 0. The anomalous anticyclonic circulation over India intensifies in triad +1. This anticyclonic anomaly can be seen in the next three triads, but with weaker intensity. Following the establishment of the anticyclonic anomaly over India, the flow on the southern side of the equatorial Indian Ocean gradually organizes into a cyclonic anomaly, which can be clearly seen in triad +3. One can also notice the development of a cyclonic circulation anomaly over eastern China and the northwest Pacific Ocean following the establishment of the monsoon break over India. Thus the evolution of the composites wind anomalies from NCEP reanalysis data is dynamically consistent with that of the observed OLR anomalies.

3. Rossby wave propagation in a simple model

An intriguing point that is of major concern to us is the initiation of break monsoon spells due to the propagation of suppressed convective anomalies from the Bay of Bengal to northwest India. Krishnamurti and Ardanuy (1980) have noted westward propagating quasi-biweekly modes in association with monsoon breaks. Further, it was seen in Fig. 5d that the spatial structure of the anticyclonic wind anomalies over the Bay of Bengal resembled that of a symmetric Rossby wave. These points suggest a possible role for Rossby wave dynamics in contributing to the rapid northwest propagation of high pressure anomalies during monsoon breaks. We shall now perform experiments using a simple shallow-water model, in order to further illustrate our hypothesis that fast westward propagation of Rossby waves from the Bay of Bengal into the Indian subcontinent can give rise to a break monsoon situation over India.

The model used in this study is based on a fully nonlinear system of shallow-water equations on a sphere. A description of a linear version of shallow-water equations can be found in Davey (1989). Physical processes such as moist convection, radiation, boundary layer physics, etc. are not included in the model. It is essentially a dry shallow-water system of equations. The only forcing in the model is the nonadiabatic heating term in the thermodynamic energy equation. The prognostic and diagnostic variables are expanded in terms of spherical harmonics truncated at rhomboidal wavenumber 40 (R40). The nonlinear terms are first calculated at grid points and later transformed to spectral domain using the transform technique (see Bourke 1974). A semi-implicit scheme with a time step of 900 s is used to integrate the model equations. Details of the model numerics are presented in the appendix. Two sets of experiments, a) control and b) free (no forcing), are carried out in this study. The purpose of these experiments is to understand the influence of the background monsoon flow on the westward propagation of the high pressure anomalies during monsoon breaks.

a. Control experiment

In the control experiment, we examine the westward propagation of high pressure anomalies in the presence of a steady-state monsoon low-level flow. The first step in the control experiment is to obtain a reasonably realistic background monsoon flow. For this purpose, the model was forced by the observed summertime diabatic heating (Fig. 6a) and integrated for 100 days, starting from a state of rest, with the forcing kept fixed throughout the time of integration. The observed diabatic heating, based on the European Centre for Medium-Range Weather Forecasts analysis compiled by Hoskins et al. (1989), shows the monsoon convective heat source over India and East Asia and the long-wave cooling in the Southern Hemisphere. It is seen that the flow in the model attains a near-equilibrium condition around day 70. The flow at the end of 100th day is taken as the steady-state response, which is shown in Fig. 6b. The salient features of the low-level circulation such as the monsoon southwesterlies; the cyclonic circulation extending across northern India and East Asia, the cross-equatorial flow; the easterlies in the Southern Hemisphere, etc. can be clearly seen in Fig. 6b. The major features of the monsoon large-scale flow are seen in the simulated steady-state response, although there are some differences between the observed and modeled low-level flow patterns. It must be mentioned that the experiments are not aimed at accurately reproducing all the features of the summer monsoon low-level circulation. The main objective of the model experiments is to un-
Fig. 6. (a) The summer (Jun–Aug) climatological diabatic heating (K day$^{-1}$) vertically averaged from surface to 100 mb; contour interval is 0.5 units; negative values are shaded and zero contour is suppressed. (b) Equilibrium response of low-level winds; scale of vector arrow is 15 m s$^{-1}$. (c) Prescribed heat sink (K day$^{-1}$); contour interval is 0.2 units. (d) Initial pressure anomaly; contour interval is 1 hPa.
understand the impact of the background monsoon westerlies on the evolution of the break anomalies.

The second step is to generate initial perturbations in the model. It was clear from the earlier discussions that a major suppression of convection occurs, prior to day 0, over the Bay of Bengal and the equatorial eastern Indian Ocean. It was also seen that anomalous anticyclones were generated in response to such an intense suppression of convective activity over this region. Our aim is to generate such anticyclonic anomalies in the model and examine how they evolve in the presence of a monsoon westerly flow. For generating the anticyclonic anomalies, we force the model with the heat sink (Fig. 6c) and perform an independent short two-day run starting from rest. The prescribed heat sink has been derived from the positive OLR anomaly in triad 1 (see Fig. 4d). The heating rate values were linearly scaled with the values of the OLR anomaly by assuming that the maximum value of the OLR anomaly over the Bay of Bengal corresponds to a cooling rate of $-10^8 \text{K day}^{-1}$. The heat sink signifies the enhanced long-wave cooling associated with the intensification of the dry convectively stable anomaly. It can be seen that the cooling to the north of the equator is larger in magnitude and covers a wider area than to the south of the equator. The high pressure anomaly, in response to the prescribed heat sink, simulated from the short two-day run is shown in Fig. 6d. It can be noticed that the pattern of the initial high pressure anomaly resembles that of a distorted Rossby wave with the Northern Hemisphere (NH) high somewhat stronger in comparison to the Southern Hemisphere (SH) high. To the east of $90^\circ\text{E}$, the pattern of low-latitude response closely resembles that of a Kelvin wave (see Matsuno 1966; Gill 1980). The wind field associated with the initial pressure anomaly shows low-latitude (easterlies) westerlies to the west (east) of $90^\circ\text{E}$ (figure not shown).

In the final step, we shall examine the evolution of the anomalous anticyclonic couplet in presence of the background monsoon flow. For this purpose we superposed the initial perturbations, obtained from the short 2-day run, on the equilibrium response of the control experiment and integrated the model for 20 days. During the course of this transient integration, only the forcing due to the monsoon differential heating (Fig. 6a) was retained in the model. However, the heat sink (Fig. 6c) was completely absent during the 20-day transient integration. We monitored the evolution of the perturbations by subtracting the steady-state response from the instantaneous response. The evolution of the pressure anomalies during the next 4 days is shown in Figs. 7a–d.

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Fig. 7. Evolution of the pressure anomaly in the control experiment after (a) 24 h (b) 48 h (c) 72 h and (d) 96 h. Contour interval is 1 hPa; negative anomalies are shaded and zero contour is suppressed.
After 24 h, it can be seen that the anomalous highs in both the hemispheres have moved considerably westward and weakened in intensity. At this time, the NH high is located over south-central India. At the end of 48 h, the NH high is seen over western and central India. Following farther northwest movement, the NH high is located around (23°N, 65°E) at the end of 72 hr. Following further westward movement, the high pressure anomaly is located around 60°E at the end of 96 h. A comparison of the evolution of the NH and the SH highs suggests that the latter has a faster westward movement and also a faster decay rate than the former. For example, in Fig. 7c, the NH high is not only stronger in intensity but is also located considerably eastward relative to its SH counterpart. This anomaly pattern with the SH high leading the NH high is consistent with observations that the SH anticyclonic circulation at 850 hPa is to the west of the NH one (see Fig. 5h). It must be mentioned that we have carried out separate experiments where we examined the evolution of an initial pressure anomaly that was perfectly symmetric w.r.t. the equator (figures not shown). These experiments also showed that the movement and decay rate of the NH high were slower as compared to the SH high. Thus the faster decay rate of the SH high seen in the transient run is not merely due to the fact that the initial pressure anomaly (Fig. 6d) in the SH was weaker than in the NH. The slow decay rate of the NH high suggests that the monsoon heating and the background flow may be supporting the maintenance of the high pressure anomalies during the course of their northwest traverse. The trailing low pressure anomalies seen over China and the Far East are generated due to the westward movement of the high pressure Rossby wave anomalies over the Bay of Bengal and India. We feel that the model simulations are too simple in order to infer about the circulation changes over the subtropics and midlatitude regions of China and the Far East. In reality there are large changes in the intensity and position of the subtropical westerlies during monsoon breaks (Ramaswamy 1962). Also the heating and circulation continuously change and mutually interact with each other in the real world. However the background monsoon differential heating is kept fixed in the model experiments. In view of these model simplifications, we have mostly restricted our discussion to the evolution of the high pressure anomalies over the Bay of Bengal and the Indian subcontinent. The above modeling experiment is quite simple minded. Therefore it is possible that there exist differences between the observed and modeled wave propagation characteristics. However, from this simple modeling study it is clear that the dynamical evolution of Rossby waves is the underlying crux for tracking the propagation of anomalous features associated with monsoon breaks.

b. Free experiment

In the free experiment, the model is integrated for 20 days without the background monsoon heating. The initial condition for the model corresponds to the initial perturbation generated from the short two-day run with the prescribed heat sink (Fig. 6c). It is important to mention that the initial perturbation in this experiment is exactly identical to that used in the control experiment. For example, the initial pressure anomaly for both the free and control experiments refers to the same pattern shown in Fig. 6d. By comparing the transients in the control and free experiments, one can infer the influence of the monsoon background flow on the evolution of the anomalous Rossby waves. The pressure anomalies at the end of 48 and 72 h are shown in Figs. 8a and 8b, respectively. In order to avoid repetitive discussions, the anomalies at the end of 24 and 96 h are not shown. As expected, one can notice a westward movement of high pressure anomalies in the free experiment. The pressure anomalies at the end of 48 and 72 h are shown in Figs. 8a and 8b, respectively. In order to avoid repetitive discussions, the anomalies at the end of 24 and 96 h are not shown. As expected, one can notice a westward movement of high pressure anomalies in the free experiment. The pressure anomalies at the end of 48 and 72 h are shown in Figs. 8a and 8b, respectively. In order to avoid repetitive discussions, the anomalies at the end of 24 and 96 h are not shown. As expected, one can notice a westward movement of high pressure anomalies in the free experiment. The pressure anomalies at the end of 48 and 72 h are shown in Figs. 8a and 8b, respectively. In order to avoid repetitive discussions, the anomalies at the end of 24 and 96 h are not shown. As expected, one can notice a westward movement of high pressure anomalies in the free experiment. The pressure anomalies at the end of 48 and 72 h are shown in Figs. 8a and 8b, respectively. In order to avoid repetitive discussions, the anomalies at the end of 24 and 96 h are not shown. As expected, one can notice a westward movement of high pressure anomalies in the free experiment. The pressure anomalies at the end of 48 and 72 h are shown in Figs. 8a and 8b, respectively. In order to avoid repetitive discussions, the anomalies at the end of 24 and 96 h are not shown. As expected, one can notice a westward movement of high pressure anomalies in the free experiment. The pressure anomalies at the end of 48 and 72 h are shown in Figs. 8a and 8b, respectively. In order to avoid repetitive discussions, the anomalies at the end of 24 and 96 h are not shown. As expected, one can notice a westward movement of high pressure anomalies in the free experiment. The pressure anomalies at the end of 48 and 72 h are shown in Figs. 8a and 8b, respectively. In order to avoid repetitive discussions, the anomalies at the end of 24 and 96 h are not shown. As expected, one can notice a westward movement of high pressure anomalies in the free experiment.
are comparable in Figs. 8a,b. Due to the above similarity in the propagation characteristics of the NH and SH highs, the pair of anticyclones in the free experiment tends to acquire symmetry w.r.t. the equator during the course of its westward traverse.

We shall now compare the transients in the control and free experiments. It can be seen from Figs. 7 and 8, that the NH high shows a much slower propagation in the control experiment as compared to the free experiment. Also the NH high is stronger in the control experiment relative to the free experiment. This implies that the rate of decay of the NH high is slower in the control experiment. On the other hand, the SH decays somewhat faster in the control experiment relative to the free experiment. The slower (faster) westward propagation and decay rate of the high pressure Rossby wave anomalies in the NH (SH) leads to a pronounced north–south asymmetry in the control run (Figs. 7b–d). Such a latitudinal asymmetry of the anomalous highs in the two hemispheres is also consistent with observations (Fig. 5h). In short, it is seen that the monsoon westerly background flow contributes to the maintenance of the northwest moving high pressure anomalies. In contrast, the high pressure anomalies tend to decay faster in the presence of a easterly flow in the SH. The result that monsoon westerlies can slow down and sustain the Rossby waves for a longer time implies that energy supply from the background flow plays a crucial role in assisting the maintenance of the high pressure anomaly over India during monsoon breaks.

4. Discussions and concluding remarks

Although earlier investigations of intraseasonal variability of the ISM had reported the occurrence of a mutual competition between convection over the Indian subcontinent and that over the equatorial Indian Ocean, the dynamical details of such a competitive interaction were not adequately clear. By combining diagnostic analysis of observations and simple modeling experiments, we have made an attempt to elucidate the basic dynamics of this competitive interaction in the context of monsoon breaks. An objective determination of the spatial structure, amplitude, and phase propagation associated with the evolution of monsoon breaks was possible through a comprehensive diagnostic analysis of daily OLR and wind data for 17 yr. Supporting experiments with a simple numerical model have enabled us to provide a new dynamical perspective of the phenomenon of monsoon breaks. Here we shall present a concise summary of the main findings of this study and explicitly identify their implications.

Based on the results of this study, the chronological sequence of events starting from an above normal phase of the monsoon is described below so as to illustrate the life cycle of monsoon breaks. During an above normal phase of the ISM, subdued convective activity and high pressures prevail over the equatorial Indian Ocean. At this time, it is noted that an excessive weakening of convective activity over the equatorial Indian Ocean induces development of a dry convectively stable (nonconvective) anomaly over this region. It is found that this nonconvective anomaly steadily spreads eastward producing a gradual weakening of convection over the Bay of Bengal, the eastern Indian Ocean, Indonesia, Southeast Asia, and the equatorial western Pacific. Indications of a progressive decrease in convective activity over these regions can be noted almost 1 week prior to the onset of the break over India. This anomalous eastward propagation is probably an outcome of interplay between low-latitude wave dynamics and moist convection in the Tropics.

It is seen that a major intensification of the convectively stable anomaly occurs over the Bay of Bengal about 2–3 days prior to the commencement of a monsoon break over India. One of the primary findings from this study is the excitation of rapid northwest propagating Rossby waves in response to a large strengthening of the nonconvective anomaly over the Bay of Bengal. This feature is noted in observations and confirmed through modeling experiments. These Rossby modes traverse along the axis extending from the central Bay of Bengal to northwest India in about 2–3 days. It is shown that the arrival of the high pressure Rossby wave anomaly over northwest India marks the initiation of a break monsoon spell. With the establishment of a break spell, it is found that the eastward spreading nonconvective anomaly decouples from the fast northwest propagating anomaly. This decoupling effect is shown to favor generation of a convectively unstable anomaly over the equatorial Indian Ocean. Such a phase reversal of convection over the equatorial Indian Ocean restores the mutual competition between the two convection zones. From the above chain of events, it is clear why monsoon break periods are characterized by enhanced convective activity over the equatorial Indian Ocean and suppressed convective activity over Southeast Asia and the equatorial western Pacific (Fig. 2a). Modeling experiments indicate that the low-level monsoon westerly flow is important in determining the Rossby wave propagation characteristics. Results from the control experiment suggest a slowing down of the movement and decay rate of the northwest propagating high pressure anomaly in the NH relative to that in the SH. Therefore, the anomalous pressure pattern in the control experiment exhibits a prominent north–south asymmetry, with the SH high leading the NH high, as is also consistently seen in observations. However in the absence of the monsoon background flow, the anomalous highs in both the hemispheres show a faster westward movement, a quicker decay rate, and a tendency to acquire symmetry w.r.t. the equator. Based on these results, it is conjectured that the monsoon westerlies in the NH play an important role in sustaining the anomalous Rossby waves over India through supply of energy.

In short a viable mechanism, based on Rossby wave
dynamics, has been proposed in order to elucidate some of the important characteristics associated with the evolution of monsoon breaks. It is convincingly demonstrated that the initiation of rapid northwest propagating Rossby waves from the Bay of Bengal to northwest India and the decoupling of the eastward and northwest propagating anomalies are two vital dynamical factors that determine the transition to a break monsoon condition and also help restore the mutual competition between the two convection zones. The use of a dry shallow-water model in the present study has limited our understanding of the interactive effects between dynamical and physical processes that can modulate the convective activity over the subcontinent, the equatorial Indian Ocean, and the subtropical regions. We would like to address these issues in our future studies.

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APPENDIX

Nonlinear Global Shallow-Water Model

The equations of the nonlinear global shallow-water model are given below. The basic prognostic variables are the low-level zonal wind velocity $u$; the meridional wind velocity $v$; and the potential temperature $\theta$. The equation of hydrostatic balance that relates $\theta$ and pressure perturbations $p$ is given by $p/\rho_a = -gH_0/\theta_0$ (see Davey 1989). In the above equation, $\rho_a (=1.2$ kg m$^{-3}$) is the surface air density; $H_0$ is the mean depth of the lower troposphere (=6000 m), and $\theta_0 (=320$ K) is the mean temperature of middle troposphere. For the purpose of convenience, the zonal and meridional momentum equations in the model are cast in the form of vorticity and divergence equations. The thermodynamic equation is rewritten in terms of a new variable $\theta^* = u \cos \phi$, $V = v \cos \phi$, while $\zeta$ is the vertical component of relative vorticity, $D$ is the horizontal divergence, $A = \zeta U$, $B = \zeta V$, and $E = (U^2 + V^2)/(2 \cos^2 \phi)$:

$$\frac{\partial \zeta}{\partial t} = -\frac{1}{a \cos^2 \phi} \left( \frac{\partial A}{\partial \lambda} + \cos \phi \frac{\partial B}{\partial \phi} \right) - 2\Omega \left( \sin \phi D + \frac{V}{a} \right)$$

$$+ K_u \nabla^2 \zeta - \epsilon \zeta \quad (A1)$$

$$\frac{\partial D}{\partial t} = \frac{1}{a \cos^2 \phi} \left( \frac{\partial B}{\partial \lambda} - \cos \phi \frac{\partial A}{\partial \phi} \right) + 2\Omega \left( \sin \phi \zeta - \frac{U}{a} \right)$$

$$+ K_n \nabla^2 D - \epsilon D - \nabla^2 (E - \theta^*) \quad (A2)$$

$$\frac{\partial \theta^*}{\partial t} = -\frac{1}{a \cos^2 \phi} \left( \frac{\partial (U\theta^*)}{\partial \lambda} + \cos \phi \frac{\partial (V\theta^*)}{\partial \phi} \right) + \theta^* D$$

$$+ K_n \nabla^2 \theta^* - \epsilon \theta^* + N^2 H_0^2 D + \left( \frac{gH_0}{2\theta_0} \right) \frac{Q}{c_p} \quad (A3)$$

The equivalent depth of the first baroclinic mode is given by $(H_{eq} = N^2 H_0^2/g = 367$ m). In the model equations, $a$ is the radius of the earth; $\phi$ is latitude; $\Omega$ is the earth’s rotation rate; $c_p$ is the specific heat at constant pressure, $N$ is the buoyancy frequency (=0.01 s$^{-1}$); $g$ is the the acceleration due to gravity (=9.8 m s$^{-2}$); $Q/c_p$ is the diabatic heating rate; $\epsilon$ (=1.5 $\times$ 10$^{-6}$ s$^{-1}$) is the Rayleigh friction and Newtonian cooling term, which corresponds to a decay timescale of about 8 days, and $K_n (=2.5 \times 10^5$ m$^2$ s$^{-1}$) is the horizontal diffusion coefficient. The factor 2 in the denominator of the diabatic heating term arises because the heating is entirely projected on the first baroclinic mode.

REFERENCES


Kanamitsu, M., and T. N. Krishnamurti, 1978: Northern summer trop-


