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**Understanding the Revival of the Indian Summer Monsoon after Breaks**

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## ABSTRACT

In this paper, we suggest a dynamical mechanism involved in the revival of summer monsoon after breaks. In this context, we carry out a diagnostic analysis using the datasets from National Centres for Environmental Prediction reanalysis-II for the period 1979-2007 to identify a robust mechanism that typifies breaks and subsequent revival of monsoon. We find that during the peak of significant breaks, an anomalous southward shift of subtropical westerly jet stream, which is invariably accompanied by anomalous northward shift of a stronger-than-normal easterly jet. These major changes during a break facilitate an instability mechanism, which apparently leads to formation of a synoptic disturbance. Formation of such a disturbance is critical to the subsequent revival of summer monsoon in 61% of the observed break to active revivals.

Computations of energetics and correlation analysis carried out suggest an increase in the eddy kinetic energy at the expense of the mean kinetic energy during the breaks, in agreement with the formation of the synoptic disturbance. This demonstrates that barotropic instability in the presence of a monsoon basic flow is the primary physical mechanism that controls the revival of the summer monsoon subsequent to the break events.


Key words: Barotropic instability; Indian summer monsoon; monsoon breaks

## 1. INTRODUCTION

The spatial and temporal variability of during the Indian summer monsoon (ISM) is very important for a country like India, which is mainly based on agriculture. The ISM experiences, in addition to the dominant interannual variability, intraseasonal variability in the form of active and break spells of rainfall. Blanford (1886), in a pioneering work suggested the "intervals of drought" as the break periods during the peak monsoon months of July-August. Also, recent studies suggest that droughts are associated with longer breaks (Joseph et al. 2009, Raman and Rao 1981). Typically, during the monsoon breaks, the monsoon trough in the sea level pressure, normally extending from the Head Bay of Bengal northwestward into Gujarat and adjoining Pakistan, is seen to propagate further north into the foothills of Himalayas. This results in anomalously surplus rainfall in the Himalayan regions, and below normal rainfall to the south (Ramamurthy 1969, Krishnamurti and Ardanuy 1980, Krishnan et al. 2000, 2009, Rajeevan et al. 2008, 2010). The active condition of the ISM, on the other hand, is when the sea level pressure trough moves south of its normal position, resulting in above normal rainfall along the climatological monsoon trough regions and in many places of the peninsular (Sikka and Narsimha 1995, Rao 1976, Alexander et al. 1978, Das 2002, Rajeevan et al. 2010, Choudhury and Krishnan 2011). Compared with other scales intraseaonal variability of the ISM represents higher amplitude of the seasonal mean (Goswami 2011, Waliser 2006). Goswami (2003) suggest that emphasis of meridional shear of zonal winds and cyclonic vorticity along the monsoon trough results in increased (decreased) frequency of occurrence of low pressure systems during active (break) phase by the intraseasonal oscillations. The intraseasonal variability of ISM manifests as two broad peaks of variability, namely a 10-20 day and a 30-60 day variability, with active and break phases which are linked to the northward migration of monsoon trough/ridge (Pai et al. 2009, Krishnamurti and Subrahmanayam 1982, Joseph and Sijikumar 2004, Krishnamurti and Shukla 2007).

The revival of active conditions during the ISM is facilitated by the formation of synoptic disturbances in the Bay of Bengal, monsoon depressions and low pressure systems that travel towards the northwest from Bay of Bengal into the Indian region (Chen et al. 2005, Sikka & Dixit 1972, Boos et al. 2015, Sikka & Gadgil 1980), many a times along the monsoon trough, and cause copious rainfall. Krishnamurthy and Ajayamohan (2010) have shown that the absence of low pressure systems such as lows, depressions, cyclonic storms etc., represents the break phase and its presence concluded as active phase of ISM.

From a dynamical perspective, some pioneering papers by Ramaswamy (1956, 1962) highlight the importance of anomalous southward shift of large-amplitude westerly troughs from the mid-latitudes into the Indo-Pakistan region during breaks in the ISM. Importantly, further analysing a case study studied by Ramaswamy (1962), Rao (1971) documents a manifestation of barotropic instability associated with increased horizontal shear due to the southward shift of the westerly troughs in the subtropical westerly jet at mid-tropospheric level in the aforementioned break event, and a subsequent revival associated with the formation of a synoptic disturbance. Rao (1971) hypothesized that manifestation of the barotropic instability during break leads to the formation of disturbances, which in turn invigorate the ISM an active phase. Satyan et al. (1980) addressed the problem by using a two-layer quasigeostrophic model and carried out a stability analysis of the simulated monsoon zonal flow corresponding to break conditions, and in this work, Satyan et al. also document the revival of the post-break monsoon through formation of a synoptic disturbance. Further, while the upper level flow in the simulations of Satyan et al. (1980) is found to be stable during the break monsoon conditions, it was found to be unstable a day before the formation of depression, supporting the argument of Rao (1971).

In the next few sentences, we briefly discuss some of the possible mechanisms such as the barotropic, baroclinic instabilities and other combined mechanisms, which have been suggested to explain the growth of the synoptic disturbances. The combined barotropic-baroclinic wind field study of the monsoon by Shukla (1977); using a ten layered quasi-geostrophic model, found that the barotropic mode is the only source for the upper tropospheric growing mode at 150 hPa. Shukla (1978) numerically integrated the linearized perturbation equations for a three-layer quasigeostrophic model and performed a combined CISK-barotropic-baroclinic instability analysis, which shows the maximum growth rate occurs for the smallest scales. On the other hand, Goswami (1980), while suggesting that a large meridional shear of the eastward component of winds at 200 hPa level and a high cyclonic vorticity at low levels over the monsoon trough region during break periods favour growth of barotropic and baroclinic instabilities, adds that these instabilities cannot explain the initial growth for monsoon depressions. Therefore, the question remains whether instabilities generated by large scale processes lead to subsequent revival of the monsoon through a barotropic instability mechanism and formation of a synoptic disturbance. In this paper, we attempt to answer this question. The availability of reanalysis datasets in the recent decades is a great opportunity in this sense. Analysis of multiple cases will also help us to refine any theoretically-based thresholds and indices that represent a phenomenon. For example, theory (Kuo 1953, Starr & White, 1954, Aihara 1959) suggests that barotropic instability occurs only in disturbances of very long wavelengths. The case study of a break monsoon Rao (1971) suggests that synoptic waves in subtropical westerly jet in the Indian region with a wavelength of above (below) 3,000 km are unstable (stable). We revisit this aspect in this study. The details of the datasets used and methods of analysis are described in the next section. We present our results in section3, followed by a section on the conclusions and discussion.

## 2. DATA AND METHODS

### *2. a. Data*

For the present study, we have used break monsoon periods based on Rajeevan et al. (2008) of ISM for the period 1979-2007. Following Rajeevan et al. (2010), we choose the region bounded by from

18.0° N to 28.0° N and 65.0° E to 88.0° E as the core monsoon region; indeed, on inter-annual scale, the area-averaged rainfall in this region is highly correlated at 0.91 with that of inter-annual variation of the Indian summer monsoon, (Rajeevan et al. 2010). The daily data employed in the study are zonal (U) wind at 200 hPa, Meridional (V) wind, Geopotential Height and Sea level pressure (SLP). All of these products were obtained from National Centres for Environmental Prediction (NCEP) reanalysis 2 (Kanamitsu et al. 2002). These Datasets are of spatial resolution 2.5° lat x 2.5° long global grid and temporal coverage are 4-times daily values for 1979/01/01 to 2007/12/31. In addition, the dates of the synoptic disturbances and locations were collected from the IMD (Cyclone eAtlas) data. In addition, the sea level pressure (SLP) data from the NCEP reanalysis 2 data were used to reconfirm the dates of formation of the synoptic disturbances. We adopt the breaks and active event dates following Rajeevan et al. (2010).

*2. b. Method*

Following Kuo (1953), Syono & Aihara, (1957) and Rao (1971), an index for barotropic instability is defined as the meridional shear in the daily 200 hPa zonal wind. Further, the critical wavelength (neutral wavelength) of a zonal wave at this level is computed as

$$L_c = \frac{2D}{\sqrt{3}} \qquad (1)$$

Where D/2 is the zonal width between subtropical westerly jet and tropical easterly jet. Indeed, Waves longer than $L_c$ (wavelength) become unstable and below $L_c$ are stable (Starr & White, 1954; Aihara, 1959). The rate of Conversion of Mean Kinetic Energy (CMKE) is obtained by

$$C(\overline{K}, K') = \int U \frac{\partial}{\partial y} \overline{v'u'}\, dm \qquad (2)$$

Where m is the mass, $\overline{K}$ is the Mean kinetic energy (Joule/sec), $K'$ is the Eddy kinetic energy, U is the zonal wind (m/s) and V is the meridional wind (m/s). A complete list of the symbols/notations representing various variables/parameters in this study is provided in Table 1. The u' and v' have been obtained as the daily anomalies from the zonal mean of the respective circulation component avaeraged over 20° E and 120° E. Equation 2 means that if there is divergence (convergence) of eddy momentum transport in region of westerlies, $\overline{K}$gets converted into $K'$($K'$ gets converted into $\overline{K}$), that is, the disturbance is barotropically unstable (stable). In our analysis, we use the criterion by Kuo (1951), which states that, for barotropic instability to happen at a location, the meridional gradient of the absolute vorticity has to be either maximum or minimum. The corresponding mathematical expression is shown in the equation 3.

$$\frac{d\zeta}{dy} = 0$$

As per Kuo (1951) the above expression for the largely zonal flow can be approximated as,

$$\frac{d}{dy}\left(\frac{-d\overline{U}}{dy} + f\right) = 0 \tag{3}$$

Where $\overline{U}$ is the mean zonal wind, f is the Coriolis force and ζ is the absolute vorticity. We use the criterion shown in equation (3) to explain the mechanism behind the formation of the post-break synoptic disturbances over the Indian region and the Bay of Bengal, which reactivate the Indian summer monsoon.

## 3. RESULTS AND DISCUSSION

### *3. a. Barotropic instability in the aftermath of breaks*

From the works of Starr and White (1954) and Rao (1971), we can suppose that such a break condition will result in barotropic instability, which may *in turn* manifest as a synoptic disturbance for the revival of ISM. In this context, from the Table 1, following Rajeevan et al. (2008), we list the dates of various post-break revival events of ISM. Of the 41 total events (Table 2), 18 revivals occurred with formation of a low pressure in the Bay of Bengal (Fig. 1, shown as an example), and

7 others with formation of a low pressure on land (Figures not shown). This result suggests that about 61% of the post break revivals are associated with formation of a low pressure in the Bay of Bengal or land regions, providing a general support to the hypothesis of Rao (1971) and Raghavan (1973).

Now, eddy formation due to barotropic instability would necessitate a conversion of the $\overline{K}$ into $K'$ , as shown by the equation 2. Indeed, this is true in many 30 out of the 41 cases i.e. 73% of post-break revival events, as evidenced by the positive values of CMKE (Table 3) (Fig. 2). This indicates that the barotropic instability is the primary possible large scale dynamical instability mechanism during the ISM breaks, and many times leading to formation of synoptic eddies. Another way to ascertain this further is by checking that there exists a significant negative correlation between the CMKE and wavelength, an indication of barotropic instability (e.g. Rao, 1971). We find a strong correlation of -0.285 (Table 2), which is significant at 95% confidence level from a Student's two tailed test. This significant correlation confirms that barotropic instability is indeed manifested after the break monsoon events, and is a necessary condition for the revival of Indian Summer Monsoon after break conditions.

What is the potential mechanism for such manifestation of barotropic instability in these sub-seasonal events? As known, barotropic disturbances derive energy from the mean kinetic energy. Energy considerations (e.g. Kuo 1951) show that for a disturbance to grow, it must tilt in a direction opposite to that of the meridional gradient of zonal wind. To be specific, a tilt from Southwest to Northeast (SW-NE) in a westerly zonal flow will meet this criterion. That is, waves with a tilt from the SW-NE will result in a maximum vorticity to the south (see equation 3, which is from Kuo 1949). From supplementary figures S1 & S2, it is seen most of the break days are also indeed associated with such a SW-NE tilt in the 200 hPa zonal flow. Such a tilt in the mean 200 hPa subtropical westerly jet over the Indian region on a typical break day, shown in Fig. 3a as well as example, along with the corresponding geopotential field (Fig. 3b), is associated with a northward transfer of westerly momentum (Kuo, 1949). In such a case, the zonally averaged eddy momentum transport ($\overline{u'v'}$) will be positive, and is, importantly, conducive to the formation of an eddy disturbance (Fig. 4a) associated with maximum vorticity to its south (Kuo, 1949). Truly, the corresponding zonal wind structure at 200 hPa shows a southward shift of the westerly jet during the break period and a northward shift of the tropical easterly jet (Fig. 4b).

From the point of Rao (1971), it will be instructive to verify that the barotropic instability is a mechanism that would help the aforementioned eddies grow in such situations. To that end, the meridional vorticity distribution of the absolute vorticity $\zeta$ in the Indian region during the break events are presented in Fig. 5a, along with the corresponding composite in Fig. 5b. Importantly, we see maximum or minimum in absolute vorticity $\zeta$ around 29° N in the composite, with the individual values varying between 25° to 30° N. Manifestation of such maximum or minimum values is a necessary condition for the barotropic instability (Kuo, 1951)from the individual case also indicates such manifestation (Fig. 5b). All this highlights the importance of the mean seasonal zonal wind structure, with westerlies to the north and easterlies to the south of the Indian sub-continent, in facilitating such a dynamical instability manifested by the breaks.

### *3. b. Wavelength Threshold for manifestation of a post-break synoptic disturbance*

Ramaswamy (1962) & Rao (1971), claim from their individual case studies, a decrease in channel width (D/2) between subtropical westerly and tropical easterly jets that manifest as a dynamical instability. We revisit this aspect by computing the D/2 during the break events in the study period. Our results, shown in Tables 4, Fig. 6, show that 32 out of 41 break events (78 %) indeed show a decrease in channel width. From this, we can deduce that a dynamical instability during the breaks is facilitated either due to a transient southward shift of the westerlies over the northern portions of the subcontinent and/or a transient northward shift of the tropical easterly jet stream over the peninsular region. Such a decrease in the channel width in the zonal width can also manifest with a weakening (strengthening) of the upper level westerlies (easterlies) in the Indian region.

Theory (Kuo 1953, Syono & Aihara 1957) shows that barotropic instability occurs only in zonal waves of wavelength shorter than a critical wavelength $L_c$ (see Equation 1). Rao (1971), from his sole case study, estimates $L_c$ of the upper level westerly Jetstream in the Indian region to be the ~3000 km. However, given that it was only a single case, and the relatively poor quality of the upper air data during that period, we use the reanalysed gridded datasets for multiple break monsoon cases to revisit this important finding by Rao (1971). Our analysis using equation 1(Table 4) shows that (i) wavelengths in the upper level westerlies north of Indian region during the summer monsoon season reach a minimum value during breaks as compared to a few days prior and after the event, and (ii) The critical mean critical value of the aforementioned wavelength, obtained by averaging it over all break events, comes to 7411 km. The minimum $L_c$ we find is just 5127 km (Fig. 7; also see Table 5).

**4. CONCLUSIONS**

Ramaswamy (1962) and Rao (1971) show, through individual case studies that transition from break to active conditions occurs during the Indian summer monsoons (ISM) owing to the manifestation of barotropic instability, which leads to formation of a synoptic disturbance. Given the critical importance of break-active cycles in defining the seasonal rainfall envelope (Goswami 2003 or Goswami and Ajayamohan 2001) during the ISM, it is very important to revisit the conclusions of these case studies. With this goal in mind, using the atmospheric circulation datasets from the NCEP-NCAR reanalysis II (Kanamitsu et al. 2002) for the period 1979-2007, we explore the potential role of break-monsoon conditions in subsequent revival of the monsoons through formation of a synoptic disturbance in the Indian region. We adopt the active and break monsoon calendar documented by Rajeevan et al. (2008). We find that barotropic instability manifests in the Indian region during break monsoons in 61% of the cases. Such a revival is found to be associated with a reduction of the ‘zonal width’ between the upper level subtropical westerlies and tropical easterlies. Our correlation analysis between the wave length of zonal winds in the Indian region and rate of conversion of mean kinetic energy values for the study period is -0.285, statistically significant at 95% confidence level, which confirms the role of barotropic instability for formation of the post-break synoptic disturbance (e.g. Aihara 1959). During the break monsoon period over most of the country there is no rainfall, and therefore the succeeding disturbances are not generated by the condensation heating. Thus, the argument that generation of monsoon depressions and synoptic disturbances due to the break-induced barotropic instability is reasonable. We also find that the mean wavelength of westerlies during boreal monsoon events north of the Indian region, which leads to the revival of the monsoons, is about 7400 km. While Rao (1971) suggests a threshold wavelength of 3,000 km from his study, our analysis of the 41 cases suggests an apparent threshold from our sample to be above 5,000 km.

As this study has been mainly carried out using the NCEP-NCAR reanalysis II (Kanamitsu et al., 2002), in future, we plan to explore these issues in other reanalysis data and various available medium range hindcast runs (e.g. Mitra 2003, 2009, 2013), and by conducting a few numerical sensitivity experiments.

We also need to remember that the Indian summer monsoon variability is controlled by several factors and drivers. In addition, formation of a disturbance depends on various other factors such as the SST, moisture availability, etc. The monsoon can also revive due to large scale circulation changes, in which case the manifested instability may be different. From this context, the relevance of the other mechanisms, such as the baroclinic instability, in the remaining cases of the break-active transitions that happen without the formation of a synoptic disturbance needs further examination.

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## Appendix-I

These are two ways of studying the development of disturbances, namely,

1. Eigen value problem (Dynamic Meteorology by Holton)
2. The initial value problem (Kuo 1953; also see Chapter 6 of Krishnamurti, 2013)

Here we have adopted the initial value problem. The symbols/notations representing various variables/parameters in the appendix are listed below (Table A1).

In order to estimate the energy exchange between the basic zonal current and a superimposed disturbance in a barotropic, non-divergent and frictionless atmosphere, we use the barotropic vorticity equation in the form.

$$\frac{d}{dt}(f + v_E) = 0 \qquad (1)$$

where

$$\frac{d}{dt} = \frac{\partial}{\partial t} + u\frac{\partial}{\partial x} + v\frac{\partial}{\partial y}$$

$f = 2\Omega sin(\phi)$ ; Coriolis force term

$\phi$ – Latitude

$v_E = \nabla^2\psi$- Relative vorticity

$\psi$- Stream function

$u$ and $v$ are the zonal and meridional components of the horizontal velocity vector, and can be expressed as

$$u = \frac{-\partial\psi}{\partial y}\,;\, v = \frac{\partial\psi}{\partial x}$$

As can be understood, x & y are the co-ordinate axes taken positive towards east and north respectively.

Linearization of equation (1) yields

$$\frac{\partial}{\partial t}\nabla^2\psi + U\frac{\partial}{\partial x}\nabla^2\psi + \frac{\partial\psi}{\partial x}\left(\beta - \frac{\partial^2 U}{\partial y^2}\right) = 0 \qquad (2)$$

U is the mean zonal current and $\psi$ is the stream function for the perturbation flow.

$\beta = \frac{df}{dy}$ --- Rossby factor

A typical solution for equation (2) will be

$$\psi = A(y,t)sin(kx) + B(y,t)cos(kx) \qquad (3)$$

Where $k = \frac{2\pi}{L}$ is the wave number, and L the Wavelength

Substituting solution (3) in the equation (2) and equating the coefficients of Sin (kx) and Cos (kx) terms, we get the following equations:

$$\frac{\partial^2}{\partial y^2}\left(\frac{\partial A}{\partial t}\right) - k^2\frac{\partial A}{\partial t} = -U\left(k^3 B - K\frac{\partial^2 B}{\partial y^2}\right) + kB\left(\beta - \frac{\partial^2 U}{\partial y^2}\right) \quad (4)$$

$$\frac{\partial^2}{\partial y^2}\left(\frac{\partial B}{\partial t}\right) - k^2\frac{\partial B}{\partial t} = U\left(k^3 A - K\frac{\partial^2 A}{\partial y^2}\right) - kA\left(\beta - \frac{\partial^2 U}{\partial y^2}\right) \quad (5)$$

(4) and (5) are two unknown equations in two unknowns, $\frac{\partial A}{\partial t}$ and $\frac{\partial B}{\partial t}$ and so form a closed system of equations.

From the prescribed initial values of u, A, B, and $\frac{\partial^2 U}{\partial y^2}$, and with proper boundary conditions, we can find solutions for $\frac{\partial A}{\partial t}$ and $\frac{\partial B}{\partial t}$.

**Initial conditions**

$A_o=0$ and $B_o$=a sinly , $l = \frac{\pi}{D}$ (6)

where D is the channel width, and suffix ‘o’ represents the initial value . As pointed by Platzman (1952), it is desirable to take initial conditions in such a way as to make the first derivative of perturbations kinetic energy zero. As would be shown later specifically in equation (10), the above condition (6) will fulfil the requirement.

**Boundary conditions-- Meridional direction**

A=0 at y=0 and y=D; $\frac{\partial A}{\partial t}$=0 and $\frac{\partial B}{\partial t}$=0 at y=0 and y=D (7)

In the X-direction we assume that the disturbance quantities have cyclic periodicity at intervals of one wavelength L. If Q is any disturbance quantity, then Q(x.y)=Q(x±L,y). Thus it is sufficient to consider the domain of integration as the area bounded by one wavelength 'L' in the X- direction and distance D in the y- direction to evaluate various kinds of energies.

**Time tendency of Amplitudes**

Amplitudes A and B after a time Δt are given by the Taylor's series

$$A(\Delta t) = A_o + \left(\frac{\partial A}{\partial t}\right)_o \Delta t + \frac{1}{2}\left(\frac{\partial^2 A}{\partial t^2}\right)_0 \Delta t^2 \pm - - \qquad (8)$$

$$B(\Delta t) = B_o + \left(\frac{\partial B}{\partial t}\right)_o \Delta t + \frac{1}{2}\left(\frac{\partial^2 B}{\partial t^2}\right)_0 \Delta t^2 \pm - - \qquad (9)$$

If Δt is sufficiently small, the above series can be truncated after the second derivative. This will no doubt introduce some error in the forecasted amplitudes. Nevertheless, it is not an essential shortcoming as shown by the results.

With initial conditions (6), (5) becomes

$$\frac{\partial^2}{\partial y^2}\left(\frac{\partial B}{\partial t}\right) - k^2\left(\frac{\partial B}{\partial t}\right) = 0 \qquad (10)$$

It can easily be shown from (10) and (7) that $\left(\frac{\partial B}{\partial t}\right)_o = 0$, everywhere,

Equations for $\left(\frac{\partial^2 A}{\partial t^2}\right)_0$ and $\left(\frac{\partial^2 B}{\partial t^2}\right)_0$ can be obtained by differentiating (4) and (5) with respect to time. They take the form

$$\frac{\partial^2}{\partial y^2}\left(\frac{\partial^2 A}{\partial t^2}\right) - k^2\frac{\partial^2 A}{\partial t^2} = -U\left(k^3\frac{\partial B}{\partial t} - K\frac{\partial^2}{\partial y^2}\left(\frac{\partial B}{\partial t}\right)\right) + k\frac{\partial B}{\partial t}\left(\beta - \frac{\partial^2 U}{\partial y^2}\right) \qquad (11)$$

$$\frac{\partial^2}{\partial y^2}\left(\frac{\partial^2 B}{\partial t^2}\right) - k^2\frac{\partial^2 B}{\partial t^2} = U\left(k^3\frac{\partial A}{\partial t} - K\frac{\partial^2}{\partial y^2}\left(\frac{\partial A}{\partial t}\right)\right) - k\frac{\partial A}{\partial t}\left(\beta - \frac{\partial^2 U}{\partial y^2}\right) \qquad (12)$$

Initial conditions (6) are used to obtain (11) and (12) since $\left(\frac{\partial B}{\partial t}\right)_o = 0$ from equations (11) and (7) it can easily be shown that $\left(\frac{\partial^2 A}{\partial t^2}\right)_0 = 0$ everywhere, so (8) and (9) reduce to

$$A(\Delta t) = \left(\frac{\partial A}{\partial t}\right)_o \Delta t \qquad (13)$$

$$B(\Delta t) = B_o + \frac{1}{2}\left(\frac{\partial^2 B}{\partial t^2}\right)_o \Delta t^2 \qquad (14)$$

so after time Δt, $\psi$ is given by

$$\psi(\Delta t) = A(\Delta t)sin(kx) + B(\Delta t)cos(kx)$$

$$\psi(\Delta t) = R_\psi cos(kx - \delta\psi) \qquad \text{where } R_\psi = [A^2(\Delta t) + B^2(\Delta t)]^{1/2}$$

and $tan(\delta\psi) = \frac{A(\Delta t)}{B(\Delta t)}$ (15)

Thus the amplitude and phase of $\psi$ wave can be found after time Δt from (15)

**Initial change of kinetic energy**

The rate of change of kinetic energy may be regarded as the rate of amplification of the disturbances. If it is positive, kinetic energy tends to increase with time, and disturbance is said to be unstable. If it is negative, kinetic energy tends to decrease, and the disturbance is said to be stable or damping. If the rate of change of kinetic energy is zero, the kinetic energy remains constant, and the disturbance is said to be neutral.

The kinetic energy of the disturbance is given by

$$K_r = \int_0^D \int_0^L \frac{u^2+v^2}{2} dxdy \tag{16}$$

But

$$u = \frac{-\partial\psi}{\partial y} = -\left[\frac{\partial A}{\partial y} sin(kx) + \frac{\partial B}{\partial y} cos(kx)\right] \tag{17}$$

$$v = \frac{\partial\psi}{\partial x} = -k[Acos(kx) - \mathrm{B}sin(kx)] \tag{18}$$

Inserting (17) and (18) into (16) we get

$$K_r = \frac{\pi}{2k}\int_0^D \left[\left(\frac{\partial A}{\partial y}\right)^2 + \left(\frac{\partial B}{\partial y}\right)^2 + k^2(A^2 + B^2)\right] dy \tag{19}$$

Differentiating (19) with respect to time and using (4), (5) and (7) we get

$$\frac{\partial K_r}{\partial t} = -\pi \int_0^D U \left[A\frac{\partial^2 B}{\partial y^2} - B\frac{\partial^2 A}{\partial y^2}\right] dy \tag{20}$$

The equation for the time change of the zonal wind is

$$\frac{\partial U}{\partial t} = \frac{-\partial}{\partial y}\overline{uv} \qquad (21)$$

where the overbar denotes a zonal average.

Multiplying (21) by U and integrating over the region we get the equation for the time change of zonal kinetic energy as

$$\frac{\partial}{\partial t}K_{r_z} = -\int_0^D \int_0^L U \frac{\partial}{\partial y}\overline{uv}dxdy \qquad (22)$$

Where $K_{r_z}$ is the zonal kinetic energy given by

$$K_{r_z} = -\int_0^D \int_0^L \left(\frac{U^2}{2}\right) dxdy$$

Using (17) and (18)

$$\overline{uv} = \frac{k}{2}\left[B\frac{\partial A}{\partial y} - A\frac{\partial B}{\partial y}\right] \qquad (23)$$

Using (23) and (22) becomes

$$\frac{\partial}{\partial t}K_{r_z} = \pi \int_0^D U\left[A\frac{\partial^2 B}{\partial y^2} - B\frac{\partial^2 A}{\partial y^2}\right] dy \qquad (24)$$

It is seen from (20) and (24) that the right hand side of (24) is the same as the right hand side of (20) but with opposite sign. Thus this term represents the interaction between the zonal and perturbation kinetic energies.

In view of our initial conditions,

$$\left(\frac{\partial K_r}{\partial t}\right)_0 = \left(\frac{\partial K_{r_z}}{\partial t}\right)_0 = 0 \tag{25}$$

Thus, as pointed out, earlier our initial conditions are such that the first derivative of perturbation kinetic energy is made equal to zero. So we have to consider the second derivative of perturbation kinetic energy $K_r$, in order to find out the initial change of kinetic energy, then

$$\frac{\partial^2}{\partial t^2} K_{r_z} = -\pi \int_0^D U \left[\left(\frac{\partial A}{\partial t}\right)_o \frac{\partial^2 B_o}{\partial y^2} - B_o \frac{\partial^2}{\partial y^2}\left(\frac{\partial A}{\partial t}\right)_o\right] dy \tag{26}$$

Initial conditions are used to get (26)

Now, we will study the stability properties of different zonal currents with initial disturbance

$$\psi = asin(ly)cos(kx), \text{ i.e.., } B_o = asin(ly) \text{ and } A_o = 0 \tag{27}$$

The actual forms of the zonal current will be selected in such a way as to study different aspects of the problem.

(i) the zonal current U is given by

$$U = ccos(2ly) \quad \text{Where } l = \frac{\pi}{D}$$

We shall discuss this symmetric mean zonal current. This profile has two inflections points (where $\frac{\partial^2 u}{\partial y^2} = 0$ ) midway between the axis of the flow and the walls. Kuo (1949) found that the presence of flex points plays an important role in the barotropic stability problem.

We now need to solve equation (4) for the above prescribed zonal wind profile and $B_o$ (given by equation 27). $\beta$ is given as

$$\beta = \frac{2\Omega cos(\phi)}{R} = \frac{\Omega}{R}\left(1 + cos(2\phi)\right)^{1/2} = \frac{\Omega}{R}\left(1 + \alpha cos(ly)\right) \qquad (28)$$

where R is the radius of the earth and $\alpha < 1$,

With the prescribed expressions for U, B and β, equation (4) is solved with the boundary conditions (7) to give

$$\left(\frac{\partial A}{\partial t}\right)_o = \frac{-E}{k^2+l^2} sin(ly) - \frac{F}{k^2+4l^2} sin(2ly) - \frac{G}{k^2+9l^2} sin(3ly) - - - - \qquad (29)$$

Where

$$E = \frac{ka\Omega}{R} + \frac{k^3 ac}{2} - \frac{3\mathrm{ka}cl^2}{2}$$

$$F = \frac{\mathrm{ka}\Omega\alpha}{2R}$$

$$G = \frac{3}{2} l^2 \mathrm{ka}c - \frac{k^3 ac}{2}$$

with the expressions for $\left(\frac{\partial A}{\partial t}\right)_o$, $B_o$ and U, the integral in (26) is evaluated to give

$$\left(\frac{\partial^2 K_r}{\partial t^2}\right)_o = \frac{\mathrm{ka}^2 c^2 l^2 \pi D}{k^2+9l^2}(3l^2 - k^2) \qquad (30)$$

$$\left(\frac{\partial^2 K_r}{\partial t^2}\right)_o = 0 \quad \text{when } L = \frac{2D}{\sqrt{3}} \qquad (31)$$

$$> 0 \quad \text{when } L > \frac{2D}{\sqrt{3}}$$

$$< 0 \quad \text{when } L < \frac{2D}{\sqrt{3}}$$

*Thus, the neutral wavelength* $L < \frac{2D}{\sqrt{3}}$ *separates the stable shorter waves and unstable longer waves.* It is to be noted that the terms due to earth's rotation will not appear in (30). So earth's rotation will not contribute to $\left(\frac{\partial^2 K_r}{\partial t^2}\right)_o$ with the symmetric profile for U considered. $\left(\frac{\partial^2 K_r}{\partial t^2}\right)_o$ is maximum at a wavelength 2.1 D and so is the most unstable disturbance.

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## Tables

**Table 1:** A complete list of the symbols/notations representing various variables/parameters in the study.

| **Symbol** | **Definition** |
|---|---|
| U | Zonal wind |
| V | Meridional wind |
| D/2 | Latitudinal distance |
| $L_c$ | Critical wavelength |
| $\overline{K}$ | Mean Kinetic energy |
| $K'$ | Eddy Kinetic energy |
| $\zeta$ | Absolute vorticity |
| f | Coriolis force |

| | |
|---|---|
| m | mass |
| u' | eddy zonal wind |
| v' | eddy meridional wind |

**Table 2:** Gives the formation of a synoptic disturbance on the particular day after every break event during the 1979-2007 period. ' *** ' represents the revival of ISM without low formation.

| Year | Low formation day after every break event | Condition |
|---|---|---|
| 1979 | | *** |
| 1979 | | *** |
| 1980 | 23 July | Bay of Bengal |
| 1980 | 26 August | Bay of Bengal |
| 1981 | | *** |
| 1982 | 19 July | Bay of Bengal |
| 1983 | 4 August | Bay of Bengal |
| 1984 | 30 July | Bay of Bengal |
| 1985 | 28 August | Bay of Bengal |
| 1986 | 9 September | Bay of Bengal |
| 1987 | 11 August | Land Region |
| 1987 | 19 August | Bay of Bengal |
| 1988 | | *** |
| 1989 | 2 August | Bay of Bengal |

| | | |
|---|---|---|
| 1989 | 16 August | Bay of Bengal |
| 1992 | 18 July | Land Region |
| 1993 | | *** |
| 1993 | | *** |
| 1993 | 6 September | Bay of Bengal |
| 1995 | | *** |
| 1995 | | *** |
| 1996 | | *** |
| 1997 | 22 July | Bay of Bengal |
| 1997 | 16 August | Bay of Bengal |
| 1998 | 1 September | Land Region |
| 1998 | 29 July | Land Region |
| 1999 | | *** |
| 1999 | | *** |
| 1999 | | *** |
| 2000 | 10 August | Bay of Bengal |
| 2001 | 6 August | Bay of Bengal |
| 2001 | 11 September | Land Region |
| 2002 | | *** |
| 2002 | 1 August | Bay of Bengal |
| 2004 | | *** |
| 2004 | 26 July | Land Region |
| 2004 | | *** |
| 2005 | 11 September | Bay of Bengal |
| 2005 | 19 August | Land Region |
| 2007 | | *** |

| | | |
|---|---|---|
| 2007 | 18 August | Bay of Bengal |

**Table 3**: Conversion of Mean Kinetic Energy Values (Joule/second) at 200 hPa.

| **Year** | **Before break period** | **During break period** | **After break period** |
|---|---|---|---|
| 1979 | -533.85 | -324.67 | 568.11 |
| 1979 | -307.49 | -1668.31 | -1040.26 |
| 1980 | -902.14 | 419.32 | -483.19 |
| 1980 | -168.76 | 838.84 | 1241.05 |
| 1981 | -445.01 | 41.06 | 654.83 |
| 1982 | -206.31 | -104.81 | -873.12 |
| 1983 | -483.50 | 932.99 | 345.91 |
| 1984 | -381.30 | 386.58 | -334.05 |
| 1985 | -691.33 | 1736.20 | 864.34 |
| 1986 | -391.23 | 667.50 | 533.51 |
| 1987 | -845.24 | -15.09 | 1909.51 |
| 1987 | -845.24 | 266.40 | -92.27 |
| 1988 | -267.84 | -600.93 | 884.11 |
| 1989 | -134.70 | 1450.52 | 1326.88 |

| | | | |
|---|---|---|---|
| 1989 | -272.84 | 722.94 | -289.01 |
| 1992 | -209.65 | 551.18 | 796.84 |
| 1993 | -662.03 | 641.99 | 500.70 |
| 1993 | -27.15 | -134.37 | 132.38 |
| 1993 | -154.66 | 171.96 | 1059.52 |
| 1995 | -645.48 | 627.13 | -336.83 |
| 1995 | -155.81 | 1094.81 | -499.51 |
| 1996 | -410.98 | 1179.61 | 18.84 |
| 1997 | -348.92 | -435.35 | -1599.46 |
| 1997 | -424.11 | -249.73 | 147.87 |
| 1998 | -111.57 | 296.10 | 1457.89 |
| 1998 | -496.43 | 597.56 | -211.52 |
| 1999 | -409.16 | -320.18 | 178.70 |
| 1999 | -671.99 | 1026.46 | 727.02 |
| 1999 | -691.03 | 1662.09 | 646.58 |
| 2000 | -51.00 | 447.82 | -588.90 |
| 2001 | -547.99 | -1176.18 | -147.23 |
| 2001 | -348.58 | 224.42 | -108.66 |
| 2002 | -1218.62 | 357.86 | -950.32 |
| 2002 | -1218.62 | 818.58 | -950.32 |
| 2004 | -309.27 | 1192.99 | -520.30 |
| 2004 | -309.27 | 558.93 | -520.30 |
| 2004 | -442.18 | 462.54 | 952.01 |
| 2005 | -88.06 | 188.81 | 754.90 |
| 2005 | -418.94 | -341.93 | -275.85 |
| 2007 | -763.64 | 693.70 | 231.90 |

| | | | |
|---|---|---|---|
| 2007 | -384.59 | 1507.19 | 1175.04 |

**Table 4**: The channel width (D/2) between the subtropical westerly jet and tropical easterly jet (In degrees) at 200 hPa during the 1979-2007 period.

| **Year** | **Before break period** | **During break period** | **After break period** |
|---|---|---|---|
| 1979 | 35.43 | 28.25 | 26.11 |
| 1979 | 31.00 | 30.00 | 34.50 |
| 1980 | 42.25 | 37.25 | 34.83 |
| 1980 | 28.88 | 23.67 | 31.83 |
| 1981 | 35.00 | 31.75 | 34.29 |
| 1982 | 29.88 | 25.63 | 34.43 |
| 1983 | 33.29 | 35.33 | 36.38 |
| 1984 | 30.63 | 22.67 | 43.00 |
| 1985 | 31.00 | 20.00 | 36.14 |
| 1986 | 33.57 | 26.00 | 26.25 |
| 1987 | 26.10 | 29.83 | 22.67 |
| 1987 | 32.14 | 29.33 | 39.14 |
| 1988 | 38.00 | 32.75 | 30.83 |

| | | | |
|---|---|---|---|
| 1989 | 26.43 | 28.00 | 43.29 |
| 1989 | 48.86 | 40.20 | 44.00 |
| 1992 | 31.22 | 28.88 | 39.63 |
| 1993 | 30.43 | 25.75 | 42.43 |
| 1993 | 33.86 | 22.29 | 31.43 |
| 1993 | 30.14 | 28.57 | 38.43 |
| 1995 | 34.57 | 26.80 | 37.71 |
| 1995 | 35.29 | 28.67 | 37.83 |
| 1996 | 37.38 | 39.00 | 37.83 |
| 1997 | 41.71 | 39.20 | 34.29 |
| 1997 | 34.43 | 30.83 | 41.00 |
| 1998 | 29.00 | 23.00 | 30.67 |
| 1998 | 24.29 | 23.14 | 35.43 |
| 1999 | 24.43 | 25.80 | 26.14 |
| 1999 | 39.57 | 35.80 | 25.86 |
| 1999 | 31.43 | 20.50 | 39.43 |
| 2000 | 34.63 | 28.89 | 37.33 |
| 2001 | 25.43 | 26.33 | 37.43 |
| 2001 | 34.00 | 29.40 | 29.57 |
| 2002 | 29.86 | 31.43 | 38.86 |
| 2002 | 29.86 | 30.82 | 38.86 |
| 2004 | 28.13 | 32.25 | 40.20 |
| 2004 | 28.13 | 23.67 | 40.2 |
| 2004 | 29.70 | 23.33 | 31.00 |
| 2005 | 31.86 | 24.25 | 38.63 |
| 2005 | 32.71 | 33.63 | 34.86 |

| | | | |
|---|---|---|---|
| 2007 | 25.38 | 25.20 | 29.43 |
| 2007 | 37.57 | 37.00 | 40.57 |

**Table 5**: Calculated values of wavelength ($L_c$ Km) before, during and after break periods.

| **Year** | **Before break period** | **During break period** | **After break period** |
|---|---|---|---|
| 1979 | 9082 | 7310 | 6693 |
| 1979 | 7947 | 7690 | 8844 |
| 1980 | 10831 | 9549 | 8929 |
| 1980 | 7402 | 6067 | 8160 |
| 1981 | 8972 | 8139 | 8789 |
| 1982 | 7658 | 6569 | 8826 |
| 1983 | 8533 | 9057 | 9324 |
| 1984 | 7851 | 5810 | 11023 |
| 1985 | 7947 | 5127 | 9265 |
| 1986 | 8606 | 6665 | 6729 |
| 1987 | 6691 | 7648 | 5810 |
| 1987 | 8240 | 7519 | 10034 |
| 1988 | 9741 | 8395 | 7904 |
| 1989 | 6775 | 7178 | 11096 |

| | | | |
|---|---|---|---|
| 1989 | 12524 | 10305 | 11279 |
| 1992 | 8004 | 7402 | 10158 |
| 1993 | 7800 | 6601 | 10876 |
| 1993 | 8679 | 5713 | 8057 |
| 1993 | 7727 | 7324 | 9851 |
| 1995 | 8862 | 6870 | 9668 |
| 1995 | 9045 | 7349 | 9698 |
| 1996 | 9581 | 9997 | 9698 |
| 1997 | 10693 | 10049 | 8789 |
| 1997 | 8826 | 7904 | 10510 |
| 1998 | 7434 | 5896 | 7861 |
| 1998 | 6225 | 5933 | 9082 |
| 1999 | 6262 | 6614 | 6702 |
| 1999 | 10144 | 9177 | 6628 |
| 1999 | 8057 | 5255 | 10107 |
| 2000 | 8876 | 7405 | 9570 |
| 2001 | 6518 | 6750 | 9595 |
| 2001 | 8716 | 7536 | 7580 |
| 2002 | 7654 | 8057 | 9961 |
| 2002 | 7654 | 7900 | 9961 |
| 2004 | 7210 | 8267 | 10305 |
| 2004 | 7210 | 6067 | 10305 |
| 2004 | 7613 | 5981 | 7947 |
| 2005 | 8166 | 6216 | 9901 |
| 2005 | 8386 | 8620 | 8935 |
| 2007 | 6505 | 6460 | 7544 |

| | | | |
|---|---|---|---|
| 2007 | 9631 | 9485 | 10400 |

**Table A1:** A complete list of the symbols/notations representing various variables/parameters in the appendix.

| **Symbol** | **Definition** |
|---|---|
| $f$ | Coriolis parameter |
| $\phi$ | Latitude |
| $v_E = \nabla^2\psi$ | Relative vorticity |
| $\psi$ | Stream function for perturbation flow |
| $u$ | Zonal wind |
| $v$ | Meridional wind |
| U | Mean zonal wind |
| $\beta = \frac{df}{dy}$ | Rossby factor |
| $k = \frac{2\pi}{L}$ | Wave number |
| t | time |
| L | Wavelength |
| D | Channel width |
| Q | Any disturbance quantity |
| A,B | Amplitude |
| $K_r$ | Perturabation Kinetic energy |
| $K_{r_z}$ | Zonal kinetic energy |
| R | Radius of Earth<br>Incremental zonal & meridional distances |

| | |
|---|---|
| dx & dy | used in integration/differentiation |
| a,α,c | Wavelength |
| Ω | Vertical ‘p’ velocity |
| ω | Angular speed of the earth |
| $\Delta$ | Del operator applied to a quantity which varies on an isobaric surface |
| $\Delta^2$ | Laplacian operator |

**Figures**

Figure 1.

Observed Sea Level Pressure (SLP) distribution, after the break period (10 Aug, 2000).

Figure 2.

Conversion of Kinetic Energy anomaly values for the 1979-2007 period.

Figure 3.

(a) Observed U- wind at 200hpa on 4august, 2000, a typical break day (b) the corresponding Geopotential distribution (in Km).

Figure 4.

(a) Eddy momentum flux transfer during (1-9 August), before (14-23 July) and after (10-15 August) break periods of a break event in the year 2000. (b) Zonal wind at 200hpa during (4$^{th}$ August), before (23$^{rd}$ July) and after (11$^{th}$ August) break periods of a break event in the year 2000.

Figure 5.

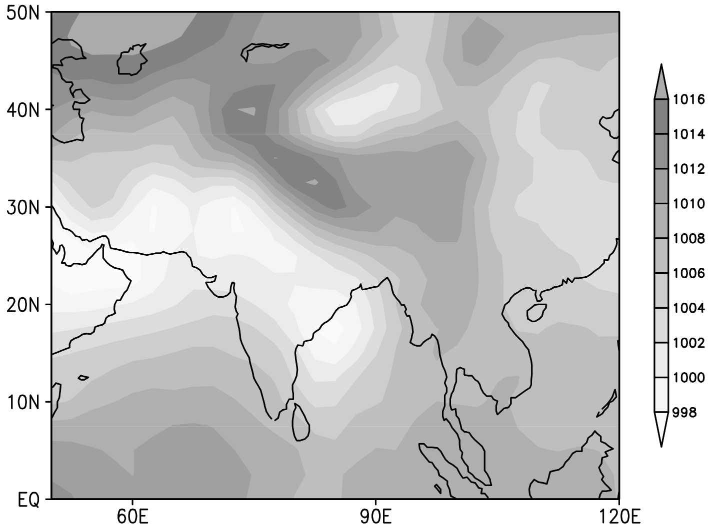
Composite of absolute vorticity profiles of break spells for the period of 1979-2007. (b) Multiple plot of absolute vorticity profiles of break periods.

Figure 6.

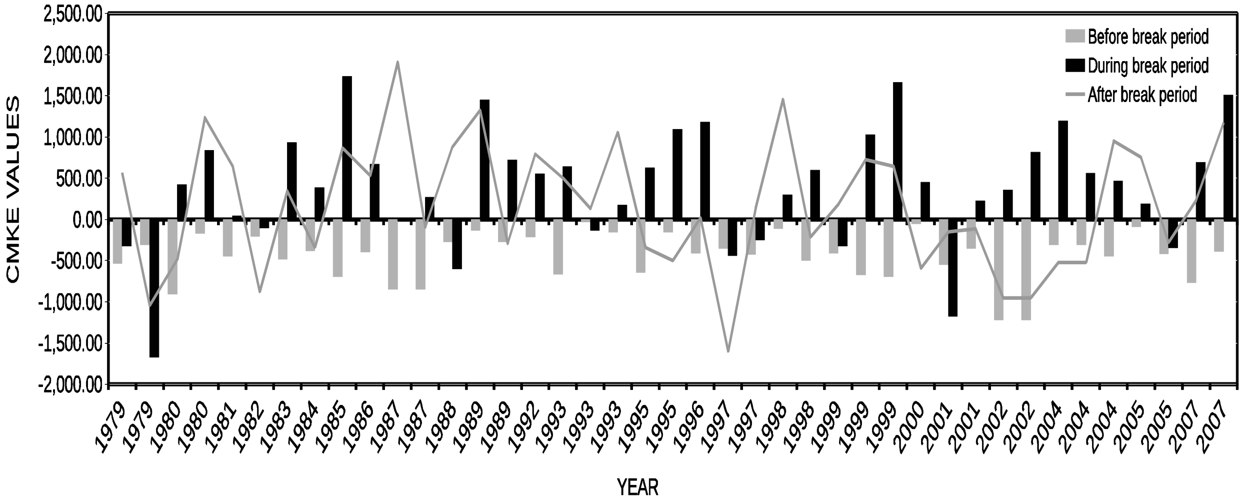
Latitudnal distance between westerlies and easterlies of the zonal wind at 200 hPa of the break events of each monsoon season.

Figure 7.

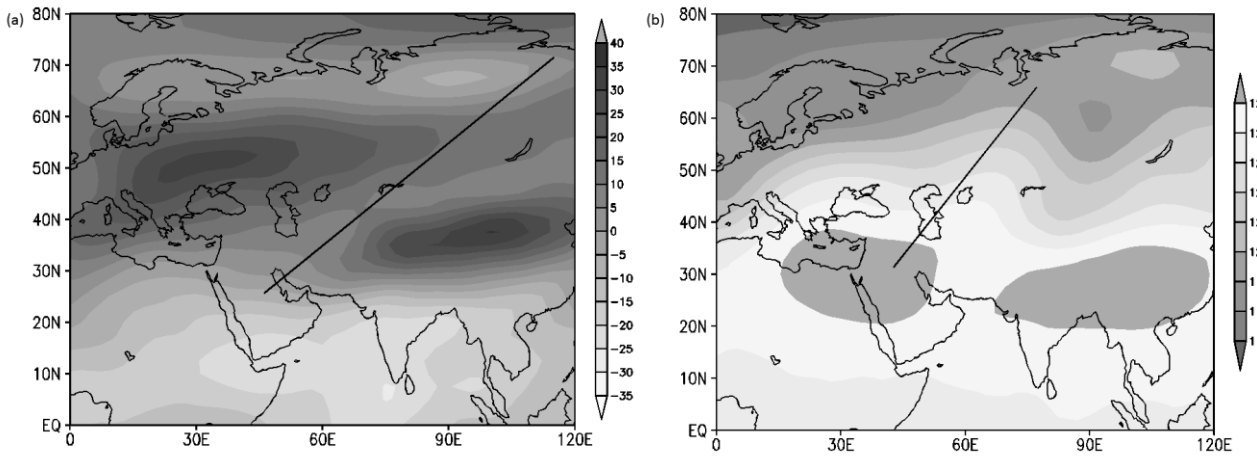
Wavelength anomalies of the zonal wind at 200 hPa averaged over the break events of each monsoon season.



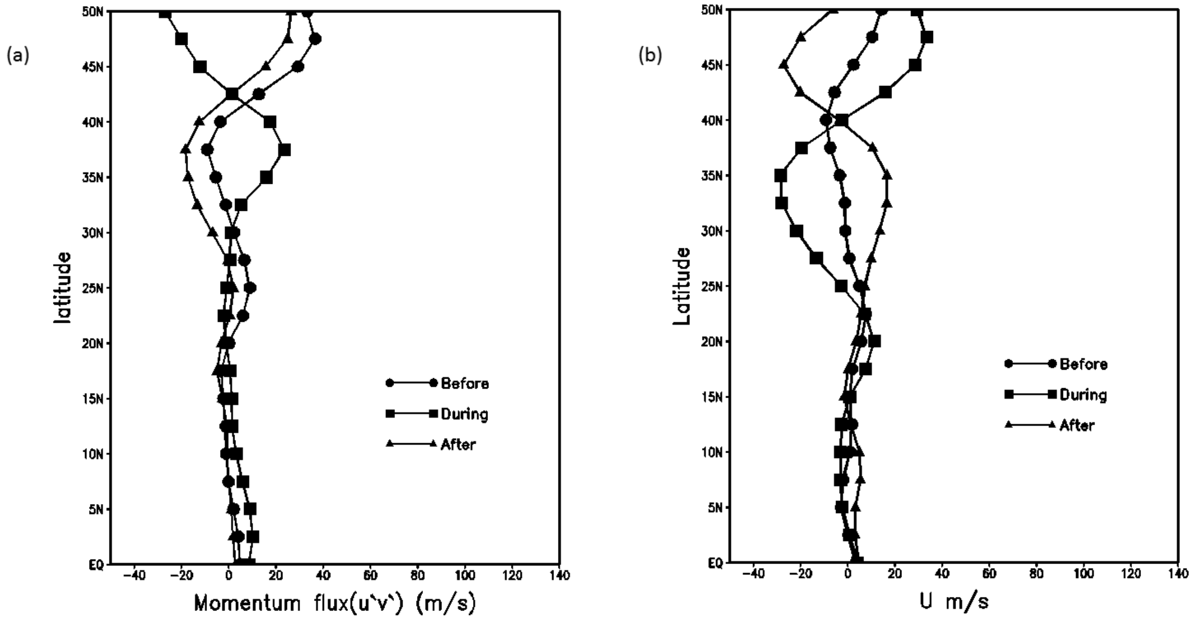

1. Observed Sea Level Pressure (SLP) distribution, after the break period (10 Aug, 2000).



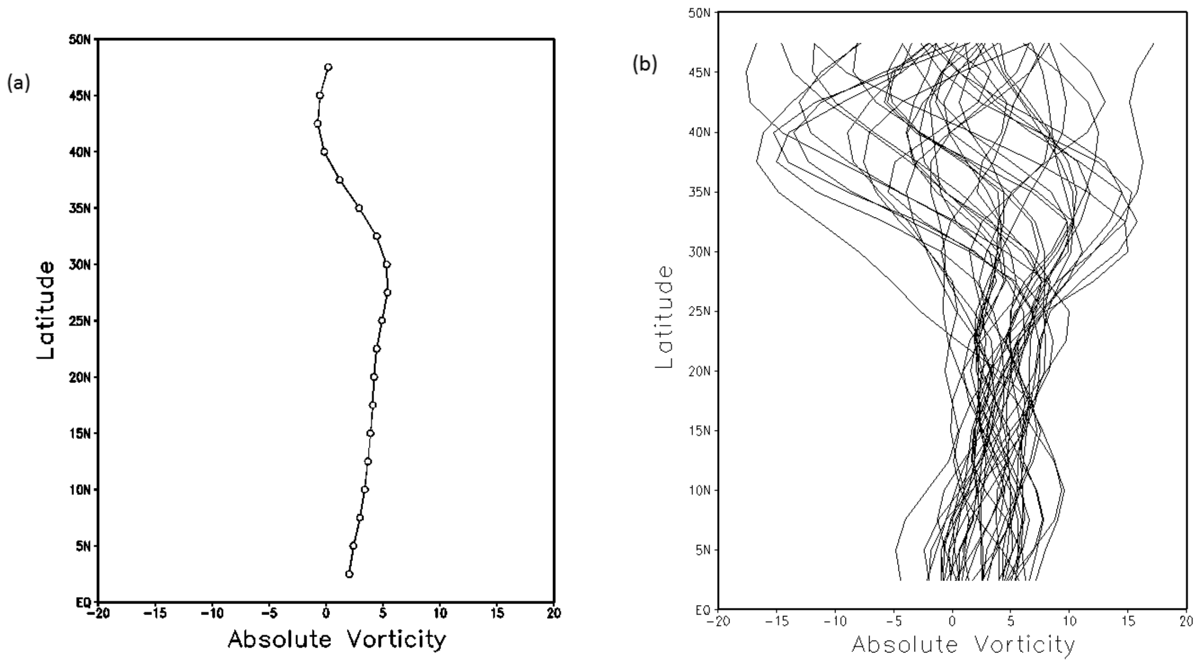

2. Conversion of Kinetic Energy (J/s) anomaly values for the 1979-2007 period.



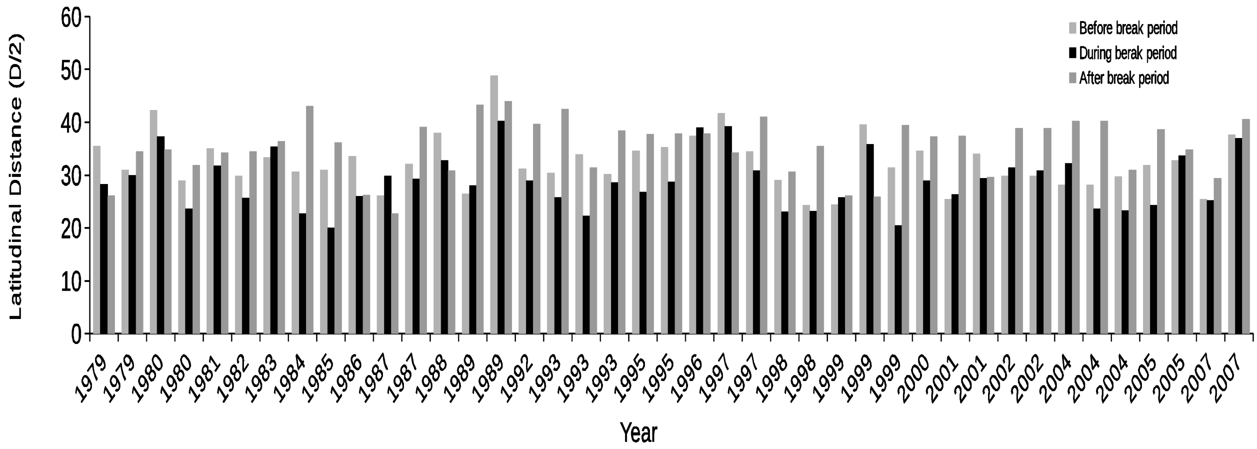

3. (a) Observed U- wind at 200hpa on 4august, 2000, a typical break day (b) the corresponding Geopotential distribution (in Km).



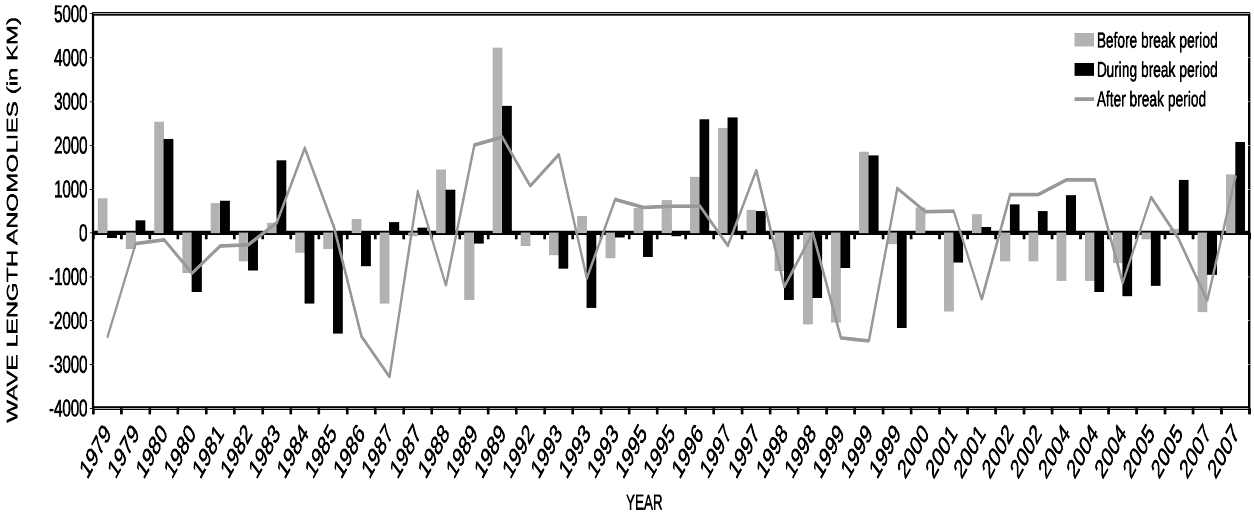

4. (a) Eddy momentum flux transfer during (1-9 August), before (14-23 July) and after (10-15 August) break periods of a break event in the year 2000. (b) Zonal wind structure of a typical break event at 200hpa, during (4th August), before (23rd July) and after (11th August) break periods in the year 2000.



5. (a) Composite of absolute vorticity profiles of break spells for the period of 1979-2007. (b) Multiple plot of absolute vorticity profiles of all 41 break periods.




6. Latitudnal distance (in degrees) between westerlies and easterlies of the zonal wind at 200 hPa of the break events of each monsoon season.




7. Wavelength anomalies of the zonal wind at 200 hPa averaged over the break events of each monsoon season.