Study on the Dynamic Process of the Onset of South China Sea Summer Monsoon

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ABSTRACT

By using the NCEP/NCAR reanalysis data from 1958 to 1997, we first looked into the atmospheric flow conditions in the one month immediately prior to the onset of the South China Sea summer monsoon (SCSSM) each year. A monthly-averaged zonal basic flow of 40-yr composite was then calculated. The stability of Rossby wave in the basic flow was studied based on the spherical barotropic vorticity equation. Furthermore, the spectral function expansion method was adopted to define and compute the evolution of a developing wave packet. The results indicate that there exists barotropic instability of spherical Rossby wave in the climatically-averaged flow field before the SCSSM onset. The instability is triggered by the westerly jet stream in the Southern Hemisphere, and the strongest instable perturbation lies to the south of the westerly jet stream. The peak of the developing spherical Rossby wave packet propagates from mid and high latitudes to low latitudes, though not crossing the equator, spurring the cumulus convection in the tropical zones. The eruption of the cumulus convection and its spread to monsoon regions help to speed up the adjustment of the general circulation and the SCSSM onset. It is concluded that elements that contribute to the SCSSM onset are on global scale, albeit the onset itself looks like a local phenomenon.

Key words: SCSSM onset, spherical barotropic instability, tropic cumulus convection, developing wave packet, dynamic process

1. Introduction

South China Sea (SCS) is an important place where the Pacific Ocean is connected with the Indian Ocean, and the South Asian summer monsoon interacts with the East Asian summer monsoon (EASM). Tao and Chen (1987) observed that the Asian summer monsoon onset first appears in the north of the SCS, and then advances northwestward step by step. Because the onset of the SCS summer monsoon (SCSSM) is an omen of that of the Asian summer monsoon, the study on the characteristics and mechanisms of the SCSSM onset is important for understanding the onset and progression of the whole Asian summer monsoon system.

Most studies on the characteristics and mechanisms of the SCSSM mainly focused on the change of large scale circulation and atmospheric heat source associated with the monsoon onset and the local conditions near SCS. Many researchers indicated that the SCSSM onset is not a local phenomenon. In the process of the monsoon onset, the atmospheric circulation experiences the first abrupt variation from spring to summer. The SCSSM onset depends on the large scale atmospheric circulation and associated seasonal proceeding of atmospheric heat source, along with the establishment of local conditions around SCS. The classic monsoon theory assumes the land-sea heat contrast to be the basic driving force for the monsoon onset. However, external heating will work through changing the structure of atmosphere. The changes of atmospheric flow activity centers or jet locations provide background fields for the formation of monsoon. The instability of background fields is sure to cause the instability of atmospheric wave patterns and the development and evolution of weather systems.

Nowadays, dynamic problems of the monsoon onset still call for further explanations. The monsoon onset mechanism has become a key for the full understanding of the EASM. The monsoon onset has long

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been recognized as a result of the global atmospheric adjustment (Zeng and Zhang, 1992) with the change of the local flow being less focused. The land-sea heat contrast caused by the seasonal variation of solar radiation is undoubtedly the essential reason for the monsoon onset. Such an external forcing is realized through atmospheric internal dynamic processes in which the instability plays an important role.

Lau and Yang (1996) studied the onset, abrupt transition, and evolution of the Asian summer monsoon. They calculated the instability of the monthly zonal symmetric basic flow along 100°–120°E in East Asia in April and May, and found that the number and growth rate of unstable modes increased significantly before the monsoon onset. Krishnakumar and Lau (1997) investigated the role of symmetric instability on the onset and abrupt variation of the monsoon onset. They calculated the instability of the monthly zonal symmetric basic flow along 100°E in East Asia in April and May, and found that the number and growth rate of unstable modes increased significantly before the monsoon onset. Krishnakumar and Lau (1998) used a spherical atmospheric Rossby wave model to calculate the effect of symmetrical instability on the onset and abrupt variation of monsoon. Their results revealed that the symmetrical instability is responsible for the monsoon onset.

2. Mathematic model and computing methodology

As the basic flow is regarded as a zonally-averaged band circulation, the horizontal non-divergent barotropic vorticity equation in spherical coordinate is:

\[
\left( \frac{\partial}{\partial t} + \frac{\pi}{\cos \varphi} \frac{\partial}{\partial \lambda} \right) \left[ \frac{1}{\alpha^2 \cos^2 \varphi} \frac{\partial^2 \psi}{\partial \lambda^2} + \frac{1}{\alpha^2 \cos^2 \varphi} \frac{\partial}{\partial \varphi} (\cos \varphi \frac{\partial \psi}{\partial \varphi}) \right] + \frac{1}{\cos \varphi} \frac{\partial \psi}{\partial \lambda} \left( \frac{1}{\alpha} \frac{\partial \bar{\psi}}{\partial \varphi} + \frac{2 \Omega \cos \varphi}{\alpha} \right) = 0, \tag{1}
\]

\[
\frac{1}{\alpha} \frac{d \bar{\psi}}{d \varphi} = \frac{1}{\alpha} \frac{\partial}{\partial \varphi} \left( -\frac{1}{\cos \varphi} \frac{\partial \psi}{\partial \varphi} \right), \tag{2}
\]

where \( \lambda \) is longitude, \( \varphi \) is latitude, \( \psi \) is stream function, and \( \frac{1}{\alpha} \frac{\partial \bar{\psi}}{\partial \varphi} \) is the vorticity gradient of basic flow shown in Eq.(2). Let \( \frac{\pi}{\cos \varphi} = U_m \) be the mean flow’s angle speed, \( \beta(\varphi) = \frac{d}{d \varphi} \cos \varphi + 2 \Omega \cos^2 \varphi \), and assume Eq.(1) has the solution shown as follows:

\[
\psi(t, \lambda, \varphi) = \Psi(\varphi) e^{i(m \lambda - \sigma t)}. \tag{3}
\]

After substituting Eqs.(2) and (3) into Eq.(1), we obtain:

\[
(s - U_m \cos^2 \varphi) \frac{d^2 \Psi}{d \varphi^2} - (s - U_m m) \frac{\sin 2 \varphi}{2} \frac{d \Psi}{d \varphi} - m [(s - U_m m) m + \beta(\varphi)] \Psi = 0. \tag{4}
\]

The boundary condition of Eq.(4) can be taken as:

\[
\Psi \big|_{\varphi = -\pi/2} = 0, \quad \Psi \big|_{\varphi = \pi/2} = 0. \tag{5}
\]

Therefore, Eq.(4) and boundary condition (5) compose an eigenvalue problem of a second-order variational coefficient linear ordinary differential equation (ODE). It can be found from Eq.(4) that, it is very difficult to obtain the analytic solution of a barotropic vorticity equation in spherical coordinate even if it only includes Rossby wave, thus its numerical solution...
will be obtained in the following approach.

We divide the latitudes from 90°S to 90°N into \( N \) equal sections with grid distance \( \Delta \varphi \), and write Eq.(4) in different latitudes. The final discretized equation can be written in a matrix form

\[
\sigma X = B^{-1} A X, \tag{6}
\]

where \( \sigma \) is the Rossby wave’s characteristic frequency, \( X = X(\Psi_2, \Psi_3, \Psi_4, \ldots, \Psi_N)^T \) is eigenfunction, which represents the stream functions of Rossby wave at different latitudes. \( A \) and \( B \) are \((N-1) \times (N-1)\) order real matrixes, the values of the matrixes are the coefficient of Eq.(4). When \( \sigma \) is a complex number, instability will occur. Its real part is the propagation frequency of unstable perturbation, its imaginary part is growth rate of instability, and the eigenfunction is the structure of Rossby wave. In this way, the frequency and structure of Rossby wave can be achieved by solving the eigenvalue and eigenfunction of matrix \( B^{-1} A \).

3. Spherical barotropic instability before the monsoon onset

3.1 Existence of instability

A basic flow shall be prescribed when the frequency and structure of Rossby wave are calculated using the barotropic vorticity Eq.(4) or (6). In order to study the mechanism of the SCSSM onset, the basic field before the SCSSM onset is configured first. The onset dates as given by Liang and Wu (2003) are shown in Table 1. The basic field shall be derived in the following approach. First, 30-day-averaged zonal wind before the monsoon onset in each year is calculated taking the onset day as a critical day. Then, a 40-yr composite is carried out. The basic field obtained this way reflects the climatic state of circulation before the monsoon onset, which is a result of the coaction of external forcing like the seasonal variation of solar radiation and the atmospheric internal process.

It is shown that the SCSSM mainly moves in the lower troposphere (Tao et al., 1983; Li and Qu, 2000), thus 40-yr composite zonal wind at 850 hPa before monsoon onset (Fig.1) are used as the basic field to solve Eq.(6). When the discretized resolution is taken as \( \Delta \varphi = 2.5^\circ \), 71 eigenvalues can be computed. The distribution of their real parts is shown in Fig.2.

It shall be noticed that there are 5 pairs of complex number within the 71 eigenvalues, the 20th, 30th, 38th, 50th, and 62th points denote the real parts of the complex numbers in Fig.2. The existence of complex eigenvalue indicates that there exist unstable modes of Rossby wave, i.e., perturbations are unstable, or the...
Fig. 2. Real part of eigenvalues as $\Delta \varphi = 2.5$. Abscissa denotes the number of computed eigen values, ordinate denotes the value of real parts, and unit is $10^{-5}$ s$^{-1}$.

Table 2. The growth rate of instability, propagation frequency, and $e$-folding time as $\Delta \varphi = 2.5^\circ$

<table>
<thead>
<tr>
<th>Mode</th>
<th>Growth rate ($\times 10^{-6}$s$^{-1}$)</th>
<th>Propagation frequency ($\times 10^{-6}$s$^{-1}$)</th>
<th>e-folding time (day)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.7767</td>
<td>0.1244</td>
<td>14.9</td>
</tr>
<tr>
<td>2</td>
<td>0.4876</td>
<td>-4.293</td>
<td>23.7</td>
</tr>
<tr>
<td>3</td>
<td>0.1240</td>
<td>-2.147</td>
<td>93.3</td>
</tr>
<tr>
<td>4</td>
<td>0.1008</td>
<td>-0.9162</td>
<td>114.8</td>
</tr>
<tr>
<td>5</td>
<td>0.0570</td>
<td>0.2470</td>
<td>203.1</td>
</tr>
</tbody>
</table>

unstable mode propagates eastward.

3.2 Structure of instable perturbations

Through analyzing five unstable modes’ structure, it can be seen that four of them have maximum values of stream function amplitude in the Southern Hemisphere. The maximum amplitude centers of unstable perturbations are all located at 60$^\circ$S and the region to the south of it. Their amplitudes are very small in the Northern Hemisphere. Only one mode shows the maximal amplitude in the vicinity of North Pole and the scale of its growth rate is an order smaller than the rest four modes. Figure 3 presents the stream function of the most unstable perturbation. If the negative area of the stream function is referred to as “trough” while the positive area as “ridge”, the trough-ridge of the unstable perturbation is tilted northwest-southeastward. Such a distribution pattern extends to the Northern Hemisphere. Then they are tilted northeast-southwestward there.

From the above analysis, the existence of the barotropic instability in the climatically-averaged flow before the SCSSM onset is proved. We infer that the barotropic instability can be regarded as a dynamic mechanism for the SCSSM onset. However, the most unstable perturbation mainly lies in the Southern Hemisphere, suggesting that the barotropic instability may trigger SCSSM onset in the Northern Hemisphere via the interaction between the Southern and Northern Hemispheres.

3.3 Role of the Southern Hemisphere basic flow

In order to reveal the role of the basic flow in the Southern Hemisphere, the zonal wind distribution in Fig. 1 is divided into four parts, which cover the South Pole easterly, Southern Hemisphere westerly, tropical easterly near the equator, and Northern Hemisphere...
It is shown in Table 3 that, for the climatic basic wind field before the monsoon onset, there is no instability when only the Northern Hemisphere westerly jet stream or the equatorial easterly jet stream exists there. Before and after the monsoon onset, the equatorial easterly jet stream makes no contribution to barotropic instability although its distribution changes dramatically. The appearance of barotropic instability mainly depends on the circulation pattern in the Southern Hemisphere. Instability can appear when there is only easterly jet stream or westerly jet stream. Instability growth rate is the largest when there exists only the Southern Hemisphere westerly jet stream. It is inferred that the SCSSM onset has a close relation to the circulation pattern in the Southern Hemisphere, and both the structure of such a circulation pattern and the magnitude of the basic flow will affect the value of the instability growth rate. When the strength of the westerly jet decreases to half (Experiment 5), the growth rate decreases by one third. Furthermore, the propagation direction of the unstable perturbations varies with different basic fields. The most unstable perturbation propagates westward (eastward) in the easterly (westerly) jet stream. All these conclusions indicate that the effect of the circulation in the Southern Hemisphere must be taken into account on the SCSSM onset. These conclusions are consistent with those from previous studies about the relationship between the atmospheric circulation and Asian monsoon using observation data (He et al., 1991; Xue and Zeng 1999; Xue, 2001; Xue et al., 2003). These studies show that the circulation variation of winter hemisphere plays a positive and active role in the interaction between the Southern and Northern Hemisphere. Therefore, when studying the summer climate variation in the Northern Hemisphere (especially the monsoon area), the variation of atmospheric circulation in the Southern Hemisphere during or prior to the monsoon onset must be paid attention to.

The structure of the unstable perturbations in Experiments 3 and 4 are similar to that in Fig. 2. The maximum amplitude appears in the region south of the westerly jet stream in the Southern Hemisphere. The stream function is very small in the Northern Hemisphere (figure omitted). Therefore, the instability of the background field before the SCSSM onset is mostly controlled by the westerly jet stream in the South Hemisphere, which also affects the structure and propagation of the unstable perturbations.
Table 3. Different circulation fields and their corresponding instability growth rates

<table>
<thead>
<tr>
<th>Experiment serial number</th>
<th>Experiment configuration</th>
<th>Number of instability</th>
<th>Maximal instability growth rate ($\times 10^{-6}$s$^{-1}$)</th>
<th>Propagation frequency ($\times 10^{-6}$s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Northern Hemisphere westerly jet only (same as the climatic basic flow in the region from 87.5° to 25°N, 0 at other latitudes)</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>2</td>
<td>Tropic easterly jet only (same as the climatic basic flow in the region from 25°S to 25°N, 0 at other latitudes)</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>3</td>
<td>Southern Hemisphere westerly jet only (same as the climatic basic flow in the region from 65° to 25°S, 0 at other latitudes)</td>
<td>1</td>
<td>0.9936</td>
<td>2.297</td>
</tr>
<tr>
<td>4</td>
<td>Southern Hemisphere westerly jet only (same as the climatic basic flow in the region from 87.5° to 65°S, 0 at other latitudes)</td>
<td>3</td>
<td>0.4588</td>
<td>-4.355</td>
</tr>
<tr>
<td>5</td>
<td>Only the Southern Hemisphere westerly jet, but half of the magnitude (same as experiment 3, except half of the magnitude)</td>
<td>1</td>
<td>0.3154</td>
<td>0.5739</td>
</tr>
</tbody>
</table>

4. Evolution of the developing wave packet

The unstable modes of perturbations are obtained by the normal mode approach. This approach can describe the increase of perturbation amplitude but not the change of perturbation structure with time. This is the limit of the normal mode method because the spatial structure is actually associated closely with time for real weather systems. To consider the perturbation evolvement with time, the perturbation wave packet must be calculated. Zeng (1983a) and Zhang and Zhang (2005) discussed the Rossby wave packet in detail and gave the formula to calculate the wave packet using spectral function expansion. In this paper, we use the same method to calculate the temporal evolvement of perturbation wave packet in the climatic basic field before the SCSSM onset. The developing wave packet consists of the sum of all eigenfunctions. The detailed formula is shown as follows:

$$
\psi'(\lambda, \psi, t) \approx \sum_j A_{s,j} \Psi_{s,j}(\phi)e^{i(m\lambda - \sigma_j t)} \\
+ \sum_j A_{I,j} \Psi_{I,j}(\phi)e^{i(m\lambda - \sigma_j t)},
$$

where $m$ is latitudinal wave number, $j$ is serial number of the eigenfunction and its corresponding weighted coefficient of a mode. $\Psi_s$ and $A_s$ are the eigenfunction and corresponding weighted coefficient of a stable mode, and $\Psi_I$ and $A_I$ are the eigenfunction and corresponding weighted coefficient of an instable mode. It is noticeable that when the time-averaged zonal flow is taken as basic flow, the wave packet retrieved by Eq.(7) is meridional but not of the whole space. Considering that the perturbation along a latitude circle is a single harmonic wave with the wave number $m$, all the latitudinal harmonic waves shall be superposed by summing up different $m$ in order to study the wave packet with different latitudinal wave numbers. For simplicity, harmonic wave with $m=3$ is considered here. Furthermore, the expansion coefficient of eigenfunction $A_s$ and $A_I$ are all taken as 1.

An analysis of Eq.(7) shows that the amplitude of the developing wave packet should grow with time because the instable mode grows exponentially with time. When the time grows infinitely, the evolution of the perturbation wave packet becomes unnoticeable compared to that of the instable growing mode. The
Fig. 4. Structure of the developing wave packet on the (a) initial time, (b) 3rd day, (c) 15th day, and (d) 90th day.
evolution of perturbations is also different from that of unstable modes in finite period.

Figure 4 shows temporal changes of the structure of the developing wave packet on the 0th, 3rd, 15th, and 90th day. At the initial time, larger amplitude wave packets appear at higher latitudes in both hemispheres. With the evolution of the developing wave packet, the region of the maximal amplitude in the two hemispheres become tilted. Along the slanting line of trough-ridge, the region of maximal wave packets amplitude expands to low latitudes, but without reaching or crossing the equator. As time goes by, the instable mode develops adequately. On the 90th day, the structure of the wave packet in the Southern Hemisphere is similar to that of unstable modes. The maximum amplitude appears at 60°S, while the amplitude in the Northern Hemisphere is very small.

The features of the developing wave packet are summarized as follows: 1) the trough-ridge of the wave packet is tilted; 2) the longitudinal scale along the tilted trough-ridge increases with time; 3) during the initial period, the maximum amplitude of the wave packet decreases as it expands to lower latitudes. Then, the amplitude will increase again with time. Because of the effect of instable modes, the structure of the wave packet develops into that of instable modes. This agrees well with the result of Zeng (1983a), who studied the development of Rossby wave packet using the Wentzel-Kramers-Brillouin-Jeffreys (WKBJ) method.

5. Role of the tropical cumulus convection

Though the barotropic instability of the basic flow is a kind of dynamic mechanism for the SCSSM onset, the interaction between the Southern and Northern Hemisphere is also required by the SCSSM onset. Considering that the developing Rossby wave packet cannot get across equator, it is inferred that there must be an intermediary at equator, through which the unstable energy in the Southern Hemisphere can get across equator, reach the SCS, and cause the SCSSM onset thereby. Zeng Qingcun (personal communication) pointed out that the effect of cumulus convection in the tropics may be the intermediary. Through analyzing the OLR isolines with the value smaller than 210 W m\(^{-2}\) before and after the monsoon onset, it can be found that, low values of OLR lie in the region from 0 to 10°S in New Guinea, Sumatra, and Kalimantan 30 days before the SCSSM onset. Then such values keep decreasing. New regions with OLR values less than 210 W m\(^{-2}\) appear again in the Pacific Ocean east of the New Guinea 15–20 days before the monsoon onset. Ten days before the monsoon onset, a large scale convective region appears in the Indian Ocean south of the Bay of Bengal, and then spreads northward and eastward, extending to Indo-China Peninsula and South China. Figure 5 shows the less than 210 W m\(^{-2}\) OLR isolines 15, 10, and 1 days before the onset; on the onset day; and 5 days after the onset.

Through analyzing the OLR distribution and evolution, it can be found that there is a close relationship between the SCSSM onset and the development of tropical convection. The convective region appears in the Pacific Ocean south of the equator firstly, then propagates to the equatorial Indian Ocean south of the Bay of Bengal, finally propagates northward and eastward into SCS.

In order to verify the relationship between the tropical convection and SCSSM onset, a numerical simulation is carried out using the T63L9 global spectral model. The initial field is the 40-yr-averaged global reanalysis data 5 days before the SCSSM onset. The model is integrated 10 days till the time 5 days after the monsoon onset. The results indicate (figures omitted) that, the model’s physical processes play a crucial role in the monsoon onset. The seasonal transition of winds from easterly to westerly, i.e., the SCSSM onset, cannot be simulated unless all the physical processes are taken into account. Among all the physical processes, cumulus convection has a substantial effect on the SCSSM onset. The westerly jet stream cannot sustain for a long time without cumulus convection, and the SCSSM onset cannot thus be predicted. In summary, numerical simulation results suggest that cumulus convection is a vital factor to the monsoon onset, in addition to solar radiation which provides the original momentum for the monsoon
Fig. 5. OLR distributions 15 days (a), 10 days (b), and 1 day (c) before the monsoon onset; on the onset day (d); and 5 days (e) after the monsoon onset.

6. Conclusions and discussion

The instability of climatic basic field, i.e., the instability of spherical barotropic Rossby wave before the SCSSM onset, is discussed by adopting wave packet dynamics. In May, the seasonal variation of solar radiation causes the seasonal variation of the atmospheric circulation. As a result of the seasonal variation, there exist two westerly jet steams in both the Southern and Northern Hemisphere, with that in the Southern Hemisphere being stronger. Before the SC-SSM onset, the spherical Rossby wave has barotropical instability in the monthly-averaged field. The instability is triggered by the westerly jet stream in the Southern Hemisphere. The largest instable perturbations lie to the south of the westerly stream in the Southern Hemisphere. The pattern of the basic flow determines the generation of instable perturbations. On the other hand, the basic flow is a response of atmosphere to external forcing, the distribution of which reflects the influences of external forcing. This emphasizes the importance of external forcing and it can be concluded that the external forcing in different seasons causes different basic flow patterns and thus produces
different instabilities of perturbations, which then lead to different climatic phenomena. Therefore, the eigenvalue issue in numerical equations, used to discuss the influence of basic flow on perturbations, actually reveals that the external forcing influence the atmosphere through the internal process in atmosphere.

The calculation and analysis of the developing wave packet of Rossby wave show that the wave-packet slopes with time and the largest amplitude expands from mid and high latitudes to low latitudes in the two hemispheres without getting across the equator. This indicates that the equator is the region of smaller wave-packet amplitude. Though the barotropic instability of basic flow is one of the dynamic mechanisms for the SCSSSM onset, it cannot propagate across the equator to complete the energy transport from the Southern Hemisphere to the monsoon area in the Northern Hemisphere. There must be an intermediary near equator which helps the unstable energy propagate across the equator to trigger the monsoon onset. Such an intermediary is found to be the tropical cumulus convection. The barotropic unstable energy propagates to low latitudes through meridional expansion of the developing wave packet and activates the tropical cumulus convection; the development of tropical convection accelerates the adjustment of atmospheric circulation and ultimately causes the monsoon onset. Therefore, although the SCSSSM onset is a local phenomenon, the contributive elements are at global scale.

It should be noted that the SCSSSM onset is a very complicated phenomenon, thus its dynamical mechanisms shall be explored from various aspects. Only one possible aspect is examined in this paper. Therefore, further studies on the dynamic mechanisms of the SCSSSM onset are still needed.

REFERENCES


