Role of the Indian and Pacific Oceans in the Indian Summer Monsoon Variability

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By

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6.35 MSSA of the PO-CSST run: Lead/lag regression between the monthly means of S-PC1 of the RC8 and the monthly anomalies of the SST and winds ......................163
The role of the Indian and Pacific sea surface temperature (SST) variability in the intraseasonal and interannual variability of the Indian summer monsoon rainfall is examined by performing a set of regionally coupled experiments with the Climate Forecast System (CFS), the latest and operational coupled general circulation model (CGCM) developed at the National Centers for Environmental Prediction (NCEP). The intraseasonal and interannual variability are studied by isolating oscillatory and persistent signals, respectively, from the unfiltered daily rainfall anomalies using multi-channel singular spectrum analysis (MSSA). This technique identifies nonlinear oscillations, its variance and period without preconditioning the data with a filter and also helps to separate the intraseasonal and low frequency climate signals from the daily variability.

It is found that, although the model has large amount of daily variance in rainfall, the combined variance of coherently propagating intraseasonal oscillations is only about
7% while the corresponding number in the observations is 11%. The model has three intraseasonal oscillations with periods around 106, 57 and 30 days. The 106-day mode has a characteristic large-scale pattern extending from the Arabian Sea to the West Pacific with northward and eastward propagations. These features are similar to the northeastward propagating 45-day mode found in the observations except for the longer period. The 57-day mode is more dominant in the region, 60°E-100°E and is strictly northward-propagating. The 30-day mode appears to be equivalent to the northwestward propagating oscillation in the observations. The dominant low frequency persistent signal in the region is due to the El Niño-Southern Oscillation (ENSO). The ENSO-related rainfall anomalies, however fail to penetrate into the Extended Indian Monsoon Rainfall (EIMR) region, and therefore, the ENSO-monsoon relationship in the model is weak.

Regionally coupled simulations of the CFS have revealed that the northeastward propagating 106-day mode exists in the model with weak amplitude and reduced variance even when the air-sea interaction over the Indian Ocean is suppressed. However, this mode was not obtained when the Indian Ocean SST variability is reduced to climatology. The spatial structure and propagation of the 106-day mode appear to be unaffected by the Pacific SST variability; i.e., a simulation with climatological SST in the Pacific reproduced this mode. The 30-day northwestward propagating mode showed little change with respect to the Indian Ocean SST, but is dependent on the air-sea interactions over the west Pacific.

Simulations using prescribed SST in the Indian Ocean showed that the spatial structure of the ENSO mode in the Indian Ocean is dependent on the air-sea interaction in
that region. It is argued that the western Indian Ocean in this model is over-sensitive to atmospheric momentum fluxes and therefore cools down quickly in response to the ENSO-induced circulation anomalies. Further, this process creates a dipole pattern with cool (warm) western and warm (cool) eastern Indian Ocean during a La Niña (El Niño) event. This dipole prevents the ENSO anomalies from reaching the EIMR region and causes the incorrect ENSO-monsoon relationship. It is also found that such a dipole pattern, although with less variance is present even in the absence of the ENSO variability. The monsoon rainfall variability in the absence of the ENSO could be dictated by internal dynamics in this model.
1.1. Introduction

The Indian summer monsoon is a regular annual phenomenon, which brings heavy rainfall to India and adjacent countries during the months of June through September. The observed record of rainfall over India shows that the summer mean precipitation averaged over the Indian land region is about 923 mm (Parthasarathy and Mooley 1978). The climatological behavior of the summer monsoon is characterized by heavy rainfall over the South Asian region (60°E-140°E, 10°S-30°N) with maxima on either side of the Indian peninsula, and over the equatorial Indian Ocean. The corresponding low-level winds show a predominant southwesterly flow across the Arabian Sea and the Indian land area and a large-scale cyclonic flow extending from the Bay of Bengal to the northwest of India (see, e.g., Krishnamurthy and Kinter 2003; Goswami 2005). This low-level cyclonic vorticity and the associated surface pressure low possibly due to radiation balance are often termed the “monsoon trough” and the southwesterly flow, the “low-level westerly jet”.

Two main features of the summer monsoon are its intraseasonal and interannual variability. Early observational studies have shown that the monsoon rainfall over India occurs in intermittent spells of wet and dry phases (see, e.g., Ramamurthy 1969), which
are known as active and break cycles or the intraseasonal variability of the monsoon. The intraseasonal variability is associated with the north-south displacement of the monsoon trough as well as the formation of low-pressure systems over the Bay of Bengal and their westward propagation inland (Krishnamurthy and Shukla 2000, KS00 hereafter). The active (break) phase of the intraseasonal oscillation is characterized by above (below) normal rainfall over the western and the central parts of India and below (above) normal rainfall over the southeastern and northeastern parts of the country. It should be noted that topography-induced and other processes also play an important role in the rainfall over the Western Ghats. During an active phase, the southwesterly winds across the Arabian Sea and the Indian peninsula strengthen and the low-level cyclonic vorticity over the monsoon trough enhances, resulting in rainfall over most of India except in southeastern parts. During the break period, the monsoon trough moves northward to the foothills of the Himalayas producing rainfall over that region and southeastern India and subsidence and dry conditions over most of the India (Webster 1986). Apart from the intraseasonal variability, the seasonal mean rainfall undergoes noticeable changes from one year to the other. The standard deviation of the area-averaged rainfall over the Indian land region is about 88 mm/day, which is about 10% of the seasonal mean. However, this seemingly small standard deviation has resulted in several drought and flood seasons across India.

Understanding the mechanisms of the intraseasonal and interannual variability is crucial for improving the monsoon forecasts. While seasonal forecasts are useful for planning and disaster preparations, successful forecasts of the intraseasonal variability
may be helpful for planning activities such as planting occurring within the season. Predictability of these two components may be dependent on their relationships with the slowly varying components of the climate system such as sea surface temperature (SST).

Using atmospheric general circulation model (AGCM) experiments, Charney and Shukla (1981) showed that climate variability in the tropics is more dependent on the boundary parameters such as SST rather than the initial conditions. Recently, KS00 proposed a conceptual model by extending this idea, in which they consider the seasonal mean rainfall as a linear combination of externally forced, low frequency components and the mean of the intraseasonal oscillations. The low frequency components are found to have strong correlations with the SST variability, while the relationship between the intraseasonal oscillations and the SST is relatively weak (Krishnamurthy and Kirtman 2009, KK09 hereafter). It was also found that the externally forced, low frequency components determine the sign and amplitude of the seasonal anomalies of the monsoon in several years. But it is also recognized that the seasonal mean of the intraseasonal components or the other high frequency components may determine the seasonal anomalies in years where the external influence is weak. Nevertheless, the presence of the low frequency, possibly SST-dependent components, which bear relatively more potential for predictability than the higher frequency modes that may often be more dependent on the initial conditions, is encouraging.

An alternate hypothesis to explain the interannual variability of the monsoon is that the seasonal anomalies are formed as a result of the interannual variability of the intraseasonal oscillations (Palmer 1994). That is, the nature and extent of the active and
break cycles in a season determine its seasonal anomaly, such that a drought (flood) year is characterized by frequent or prolonged break (active). It is also argued that the role of SST or other slowly varying components may be in just altering the preference of the system to a certain active and break cycle pattern in each season. The potential for seasonal predictability in this scenario, is linked to the ability to predict the statistical nature of the active and break cycles ahead of a season. Goswami and Mohan (2001) and Sperber et al. (2000) have shown some evidence in support of this hypothesis using the reanalysis winds. However, studies by KS00 and Krishnamurthy and Shukla (2007; 2008, KS07 and KS08 hereafter) have found evidence for Charney-Shukla hypothesis in observed records of rainfall over India and outgoing longwave radiation (OLR).

The motivation for the present study comes from the findings of KS07 and KS08 and the conceptual model for the interannual variability proposed by KS00. The existence of the intraseasonal and low frequency modes in the observations and their dependence to SST variability is well documented in KS07, KS08 and KK09. The aim of this study is to understand the role of SST variability in the Indian and Pacific oceans in determining the intraseasonal variability and the seasonal mean monsoon using numerical model simulations. In this study, we perform numerical model simulations to examine the sensitivity of these modes to SST variability in the Indian and western Pacific Oceans. We provide a summary of previous studies in the following sections and list the objectives at the end.
1.2. Intraseasonal Variability

The first part of this study is to understand the role of SST in the intraseasonal variability of monsoon. A description of previous studies on the monsoon intraseasonal variability is provided next.

1.2.1. Background

Based on observational studies, the monsoon intraseasonal variability is believed to consist of two main oscillations with periods broadly falling around 30-60 days and 10-20 days (Krishnamurti and Bhalme 1976; Yasunari 1979; Hartman and Michelsen 1989; Lau and Chan 1986; Annamalai and Slingo 2001). Kirshnamurti and Bhalme (1976) reported that a quasi-biweekly oscillation exists in many elements of the monsoon system such as surface pressure, monsoon rainfall, low-level winds and monsoon cloudiness. Yasunari (1979) reported a similar result in the cloudiness data. Hartmann and Michelsen (1989), Yasunari (1979) and Lau and Chan (1986) have reported the presence of a 30-50 day oscillatory signal. The 10-20 day oscillations are generally associated with westward propagating events entering the Indian land region from the Bay of Bengal and the 30-60 day oscillations with the northward propagation of cloudiness and rainfall from the equatorial Indian Ocean to northern India.

Despite these studies, a comprehensive picture of the space-time structure of the intraseasonal oscillations was missing until recently. Applying multi-channel singular spectrum analysis (MSSA) on the daily OLR data over the domain 40°E-100°E, 20°S-35°N for the period 1975-2002, KS08 showed that the summer intraseasonal variability
consists of two dominant non-linear oscillations with periods centered at 45 and 28 days. A similar MSSA was performed on the gridded daily rainfall data over India for the period 1901-1970 (KS07), and the results are consistent. A summary of the spatial structure and propagation characteristics of the 45-day and 28-day oscillations is provided here. Phase composites of the 45-day oscillation shows that the convective anomalies originate over the equatorial Indian Ocean, propagate slightly eastward and then northward to the Indian subcontinent. The most notable feature of this oscillation is the large-scale structure extending from 60°E to 160°E in a northwest-southeast orientation. The 45-day oscillation has clear eastward propagation in the equatorial belt (5°S-10°N) and northward propagation over the Indian region (60°E-100°E) and East Asian region (100°E-160°E). The spatial structure of the 28-day mode shows a quadruple structure in its peak phase with same sign anomalies over India and Maritime Continent and opposite sign anomalies over the Indian Ocean and China Sea. The 28-day mode shows eastward propagation near the equator and westward propagation over the latitudes 10°N-25°N, and northward propagation over the Indian and western Pacific regions. The life cycles of the 45 and 28-day modes were shown to correspond to the active and break cycles over India.

Low-level wind anomalies corresponding to these modes can be obtained by regressing the Indian Monsoon Rainfall (IMR) index computed from the reconstructed components (RCs) of the 45 and 28-day modes to the daily anomalies of 850hPa wind anomalies from the reanalysis. A similar structure of the low-level circulation is obtained by performing a combined MSSA on the daily anomalies of the horizontal wind at
The period of the intraseasonal oscillations obtained from that MSSA is consistent with the KS08 results.

The above-mentioned results further verify the robustness of the 45 and 28-day modes in observations. The present study is an extension of KS08 with the objective to understand the relative roles of SST in the Indian, western Pacific and eastern Pacific Oceans in the evolution of the 45 and 28-day intraseasonal modes. Before describing the goals, a summary of the previous studies that have examined the role of SST in intraseasonal oscillations is provided in the next section.

1.2.2. Observational Evidence for Air-Sea Interaction

A large number of studies has examined the role of SST on the winter intraseasonal oscillation or the Madden Julian Oscillation (MJO) (see, review by Hendon 2005). Because of the lack of observational data over the Arabian Sea and Bay of Bengal and the complexities associated with the summer intraseasonal oscillation (ISO), relatively less number of studies has been published on the role of SST in the summer intraseasonal variability. Using high-resolution SST measurements from TRMM (Tropical Rain Measuring Mission) data, it is shown that there are fluctuations in the SST in the intraseasonal timescale in the Indian Ocean (Sengupta et al. 2001; Vecchi and Harrison 2002). Sengupta et al. (2001) note that the intraseasonal fluctuations in SST have amplitudes of 0.6°-0.8° and large horizontal structure similar to the ISO signal in convection. Vecchi and Harrison (2002) report that the basin wide average amplitude of the SST changes is about 1° to 2° C and that the air-sea interaction could be an important
factor in the evolution of active and break cycles. Observational study of Klingaman et al. (2008) also suggests that there exists a coherent intraseasonal oscillation in SST in association with the convection. The lead/lag relationship between SST and atmospheric fields indicate that the SST anomalies are generated by the atmospheric fluxes.

The role of SST specifically in the 45 and 28-day modes has been examined by KK09. Their results showed that the 45-day mode has moderate correlations with the daily SST. But there is evidence of an intraseasonal oscillation in SST extending from the Arabian Sea to the western Pacific similar to the spatial structure of the OLR anomalies of this mode. The correlations between the 28-day mode and the SST were found to be negligible indicating that this may be an atmosphere-only variability.

1.2.3. Model Experiments on Air-sea Interaction

A few studies have examined the role of intraseasonal SST variations in the ISO modes using general circulation model (GCM) experiments. Fu et al. (2003; 2004) showed that the coupled model produces more realistic intraseasonal oscillations and the phase relationship between SST and precipitation. Rajendran et al. (2004) also show similar results.

A generic picture of the air-sea interaction mainly based on the MJO can be summarized as follows (Hendon 2005). Enhanced convective anomalies are co-located with increased cloudiness and reduced incoming shortwave radiation while suppressed convection (seen east or north of it) is associated with clear skies and increased incoming shortwave radiation. Anomalous easterlies persist over the suppressed convective region
resulting in a reduction in amplitude of the mean westerlies, which in turn, results in reduced evaporation and latent heat flux. Reduction in the incoming radiation and evaporation acts in combination to increase the SST over the suppressed convective region. This warm SST ahead of the convective region reduces atmospheric pressure above it and increases surface convergence and humidity and destabilizes the atmospheric column for convection. It is believed that a similar mechanism may exist in the northward propagating summer oscillations as well.

1.3. Interannual Variability

The second part of the study is to examine the role of SST variability in the interannual variability of the monsoon. A summary of studies in this area is provided in this section.

1.3.1. Role of the Pacific Ocean

Since the interannual variability in the Pacific is dominated by the ENSO phenomenon (Bjerknes 1969), the impact of the Pacific SST on the monsoon variability is considered to be mainly through the ENSO. The relation between the Southern Oscillation and the Indian summer monsoon is one of the earliest observed teleconnections in the global climate studies (Walker 1924). Observed record of rainfall shows that India tends to experience below-normal monsoon during an El Niño or warm Pacific event and above-normal monsoon during a La Niña or cold Pacific event (Walker 1924; Sikka 1980; Angell 1981; Rasmusson and Carpenter 1983; Shukla and Paolino 1983; Ropelewski and Halpert 1987). The relationship between the monsoon rainfall and
ENSO is often presented as a lead/lag correlation plot between the monthly mean NINO3.4 index and the seasonal anomalies of Indian Monsoon Rainfall (IMR) index. The IMR index is the area averaged rainfall over the Indian land region and NINO3.4 index is the SST anomalies averaged in the region 170°W-120°W, 5°S-5°N. The correlation has a maximum value of about 0.6 during October-January following the monsoon season (see e.g. Kirtman and Shukla 2000). Since the maximum correlation appears after the monsoon season, the predictive value of this relationship is considered poor.

One of the mechanisms proposed to explain the observed negative correlation between the ENSO and the summer monsoon rainfall is the modulation of the Walker circulation in the Pacific (Shukla and Wallace 1983; Palmer et al. 1992; Ju and Slingo 1995; Soman and Slingo 1997). The climatological Walker circulation is characterized by large-scale ascending motion over the warm waters of the western Pacific and descending motion over the cold tongue region in the eastern and central Pacific. Observational studies and numerical model simulations have suggested that, during El Niño years, the ascending branch of the Walker cell gets shifted eastward as a response to the anomalous warming in the central and eastern Pacific. This eastward shift in the Walker cell is supposed to induce subsidence anomalies over the Indian region and deficient rainfall as a result. While several studies agree on the eastward shift in the Walker circulation during El Niño years, it is not clear how exactly this circulation anomaly induces subsidence over the Indian region. Based on their atmospheric general circulation model (AGCM) simulation, Ju and Slingo (1995) suggested that anomalous Walker circulation
might induce anomalies in the latitudinal position of the Inter Tropical Convergence Zone
(ITCZ), which in turn might cause subsidence over India. Another hypothesis is that the
anomalous Walker circulation produces convergence over the equatorial Indian Ocean
(IO) thereby encouraging divergence and subsidence over the Indian land region
(Goswami 1998).

Apart from modulating the east-west circulation in the tropics, the ENSO-related
anomalies can induce Rossby wave type responses over the western and north Pacific and
Indian oceans (Wu and Kirtman 2004a). As stated by Wu and Kirtman (2004a), “the
Rossby wave induces anomalous downward (upward) motion and low-level anomalous
anticyclonic (cyclonic) circulation over the eastern north Indian Ocean and South China
Sea. This would in turn induce changes in the shortwave radiation and evaporation and
produce an eastward shift of cold (warm) SST anomalies and low-level easterly
(westerly) anomalies through SST gradient changes.” Several authors have pointed out
the importance of the western Pacific SST anomalies in the ENSO-monsoon relation
(Soman and Slingo 1997; Lau and Wu 2001; Annamalai and Liu 2005). It was found that
the sign and the zonal extent of the western Pacific SST anomalies play a major role in
the ENSO-monsoon teleconnection.

Since the observed maximum lagged correlation between NINO3.4 and IMR
index is only about 0.6, the question arises as to why the monsoon-ENSO teleconnection
fails during certain years. As examples, the El Niño events of 1994 and 1997 were not
associated with deficient monsoon over India. Although an obvious explanation for this
would be the presence of internal dynamics in the system, several mechanistic
explanations are also put forward. One such hypothesis suggests that the atmospheric response to the ENSO forcing might depend on the amplitude of the ENSO-related anomalies (Slingo and Annamalai 2000). By comparing the atmospheric response to the El Niño events of 1982, 1987 and 1997, they showed that during the year 1997, apart from the eastward shift in the Walker circulation, a predominant suppression of the convection over the maritime continent and the Indian Ocean was also present. These dry convective anomalies might have induced a more active monsoon trough formation over India and thereby above-normal rainfall in that region. The response of the monsoon trough can be viewed as a modulation of the local Hadley cell. Based on this result, the authors argue that, during the year 1997, the amplitude of the El Niño was very large so that the atmospheric response or the anomalous Walker circulation entered a different regime and substantial modification of the local Hadley circulation was produced as a result. However, it is not clear whether remote responses of the ENSO are linearly dependent on the amplitude of the event. Another observational study suggests the importance of the regional processes in the Indian Ocean during 1997 (Lau and Wu 2001). A combined Singular Value Decomposition (SVD) analysis of SST and precipitation in that study showed that the above normal rainfall of 1997 is almost entirely contributed by the SVD mode that represents regional coupled processes in the Indian Ocean.

Recently, it has been proposed that the Indian monsoon is highly sensitive to the spatial distribution of the SST anomalies associated with the El Niño (Kumar et al. 2006). It is known that certain El Niño events are characterized by warm SST anomalies
extending to the central Pacific. Using composite analysis of SST anomalies, Kumar et al. (2006) argue that the drought years in India are associated with those El Niño conditions where the anomalous warming is extended to the west. Sensitivity studies using an AGCM seem to confirm this hypothesis. This result implies that the inclusion of the spatial distribution of the SST anomalies in the statistical prediction models should improve its ENSO-monsoon relation. However such an attempt by Rajeevan and Pai (2007) did not report any success. Their statistical model results conclude that the spatial pattern of the SST anomalies may not be critical in understanding the monsoon-ENSO relationship.

1.3.2. Role of the Indian Ocean

Relatively less number of studies have examined the influence of the Indian Ocean on the monsoon variability. Earlier studies showed that an excess (deficient) monsoon rainfall produces cold (warm) anomalies in the Arabian Sea in the spring following the monsoon season (Shukla 1987). Recently, as the use of the GCMs has become more common, several studies have examined the role of the Indian Ocean on the monsoon variability. Coupled GCMs allow us to examine the impact of the Indian Ocean by turning off the variability in that ocean basin.

An important result reported by several independent studies is the relevance of the atmospheric bridge-like mechanisms (Lau and Nath 1996, Klein et al. 1999) between the Indian and Pacific Ocean basins. It has been demonstrated that warm Pacific anomalies are well correlated with SST anomalies in the Indian Ocean approximately 3-6 months
after the peak phase of the warming in the Pacific. This delayed correlation can be explained as follows: the response of the atmospheric circulation to the El Niño anomalies produces changes in cloud cover and evaporation and modifies net heat flux over remote ocean basins such as Indian Ocean (Klein et al. 1999; Lau and Nath 2000). Therefore, it is possible that these SST anomalies in the Indian Ocean induced by the Pacific anomalies can significantly influence the monsoon variability. It has also been demonstrated that coupled air-sea processes in the Indian Ocean are critical to simulate the observed ENSO-monsoon correlation (Yu et al. 2002; Wu and Kirtman 2004b).

Recently, several studies have examined the independent influence of the Indian Ocean Dipole (IOD) mode (Saji et al. 1999; Webster et al. 1999) on the monsoon variability. However it is widely debated whether the dipole mode is an independent coupled mode of the Indian Ocean that has no relation with the ENSO (e.g., Allan et al. 2001; Krishnamurthy and Kirtman 2003). Sarkar et al. (2004) showed that the local dynamics in the Indian Ocean has become more active during the recent decades and has a dominant influence on the interannual variability of the monsoon. AGCM studies have shown that the combined influence of the ENSO and the Indian Ocean dipole mode determines the interannual variability of the seasonal mean monsoon rainfall (Ashok et al. 2001; 2004).

In summary, several independent studies agree on the modulation of the Walker circulation as an important mechanism through which the ENSO influences the monsoon variability. However, it is not clear whether the anomalous Walker circulation directly induces subsidence over India or the atmospheric and oceanic anomalies over the western
Pacific and Indian Ocean are critical in establishing this teleconnection. During the years when the traditional ENSO-monsoon teleconnection fails to manifest, the Walker circulation anomalies alone are not sufficient to explain the remote impact of the ENSO. Additionally, most studies have focused on the influence of El Niño conditions on the monsoon. For a comprehensive understanding of the remote influences of the Pacific Ocean, it is important to examine the influence of La Niña conditions as well.

A clear explanation of the mechanisms for the remote influence of the Pacific SST is lacking. Studies have indicated that the coupled processes in the Indian Ocean and western Pacific play important roles in determining the monsoon variability. Although the importance of the Indian Ocean dipole mode has been pointed out, numerical experiments using coupled GCMs have not been performed to address this problem. It is to be noted that all of the above-mentioned studies used seasonal or monthly mean quantities in their analyses. A different approach would be to isolate dominant modes of the monsoon variability from daily mean values and to examine the influence of the SST on each of these modes.

1.3.3. Seasonally Persistent Signals in Daily Variability

It was earlier noted that the observational studies by KS08 and KK09 show that externally forced, low frequency components can be isolated from the daily variability using the MSSA. KS08 refer these components as the seasonally persistent signals since these anomalies were found to persist with the same sign and similar amplitude throughout the season. As mentioned earlier, these studies report that two seasonally
persistent signals are found in the daily OLR data and their contribution to the seasonal anomalies of the monsoon is significant. The persistent signals were also found to have good correlations with the ENSO and the Indian Ocean Dipole variability. An important conclusion of KS08 was about the relative roles of the Indian and Pacific oceans in determining the seasonal mean monsoon. For example, in 1997 the impact of the El Niño was surpassed by a strong dipole event in the Indian Ocean, resulting in a normal monsoon season, while in 1987, the ENSO and dipole modes acted in concert. Based on these findings, the first step toward improving the seasonal predictability must be to understand the mechanisms of these persistent modes.

1.4. Objectives

The first objective of this study is to examine the role of the daily variability of SST and air-sea interaction in the Indian and Pacific oceans, specifically on the 45 and 28-day oscillations, using numerical experiments with a coupled general circulation model (CGCM). While studies have examined the role of air-sea interaction on the northward propagation of the summer ISO, no study, to our knowledge, has comprehensively examined the role of the SST on both the northward and eastward propagating modes. We believe that there is ample observational evidence that these are two modes that exist with distinct periods, spatial structure, propagation features and possibly driven by different mechanisms. Studies, so far, have largely looked at either the 30-60 day or the 10-20 day variability by pre-filtering the data in their respective frequency bands. In this study, we isolate the oscillatory modes from the unfiltered data.
using the MSSA technique and examine the relative roles of the Indian and Pacific SST variability in the spatial structure and propagation features of the 45 and 28-day modes. The objective is achieved by performing a series of regionally coupled simulations using one of the latest CGCMs. Details of the simulations are provided in the next chapter.

Within the context of modified Charney-Shukla hypothesis (KS00), the second objective of this study is to understand the role of Indian and Pacific Oceans in producing the seasonally persistent components that have a major role in determining the seasonal mean monsoon. We emphasize that this study focuses on the climate signals isolated from the daily data, and the benefits of this approach are described below. First of all, this is one of the very few methods to our knowledge that can be used to examine the relative importance of intraseasonal oscillations in determining the seasonal anomalies. Secondly, by using the MSSA, we are able to isolate the daily variability of the climate signal. It can be argued that an MSSA performed on monthly data will isolate the persistent signals although intraseasonal signal is not obtained. However, due the strong seasonality of the monsoon, such an analysis is technically difficult to perform. The specific questions addressed in this part are: What is the role of the Indian and Pacific SST variability in generating or determining the persistent signals, particularly the ENSO mode? Is the air-sea interaction over these basins crucial for the development of this mode? These questions are investigated by performing a series of regionally coupled simulations with prescribed SST over the Indian and Pacific Oceans.
Chapter 2 Methodology

2.1. Model

The numerical experiments in this study are conducted using the Climate Forecast System (CFS), the current operational coupled GCM developed by the National Centers for Environmental Prediction (Saha et al. 2006). The atmospheric component of the CFS is a coarse resolution version of the NCEP’s Global Forecast System (GFS) and the ocean component is the Modular Ocean Model version 3 (MOM3) developed by the Geophysical Fluid Dynamics Laboratory (GFDL) in Princeton, NJ. The GFS has a spectral triangular truncation of 62 waves in the horizontal (equivalent to 200km Gaussian grid) and a finite differencing in the vertical with 64 sigma layers. The Ocean model MOM3 uses spherical coordinates in the horizontal and \( z \) coordinate in the vertical. The zonal resolution is \( 1\degree \) and the meridional resolution is \( 1/3\degree \) between \( 10\degree S \) and \( 10\degree N \) and gradually decreases poleward till \( 30\degree N \) and \( 30\degree S \). The meridional resolution is fixed at \( 1\degree \) poleward of \( 30\degree S \) and \( 30\degree N \). The atmospheric and oceanic components exchange daily average quantities of heat and momentum fluxes once a day. No flux correction has been implemented in the CFS. The atmospheric and ocean model are coupled in the region between \( 65\degree S-50\degree N \). Observed climatological SST are used to force the model in the region poleward of \( 74\degree S \) and \( 64\degree N \). In the region between the
latitudes 74°S and 65°S and between 64°N and 50°N, the SST is obtained by weighted averages of the observed climatology and the SST from the ocean component of the CFS. The sea ice extent is prescribed from the observed climatology.

2.2. Design of the Experiments

The role of SST variability in different ocean basins is examined by performing regionally coupled simulations of the CFS. The basic idea is to prescribe SST in certain regions so that either the air-sea interaction or any SST variability in addition to the annual cycle is suppressed in that region. This allows an environment in which the sensitivity of different climate phenomena to SST variability of a specific region can be studied. A similar technique has been used in some earlier studies (see, e.g., Lau and Nath 2000; Huang and Shukla 2007). The methodology adopted in setting up the experiments is outlined below. The atmospheric and oceanic components of the model are initiated with their respective initial conditions. At the end of the first day, before the two models exchange fluxes and SST, the global SST field is modified in such a way that the prescribed values are inserted in grid points wherever the coupling needs to be suppressed. The prescribed SST values can be climatological or mean values from the observations or from a fully coupled simulation of the same CGCM. This procedure is repeated each day of the simulation. Therefore, in this design, the atmosphere feels the prescribed SST in one ocean basin, while the SST produced by the OGCM is felt in other basins. Note that, in this design, a global ocean is present throughout the simulation, and produces SST at each time step, but the SST is transferred to the atmosphere only over
certain regions. This means that there are no lateral boundaries inserted in the ocean to impose regional coupling. A potential problem of this design is that unrealistic oceanic fluxes may develop over time in the uncoupled regions, since the ocean model receives fluxes that are derived based on the prescribed SST values. These anomalies in the SST and fluxes can develop into circulation anomalies, which may propagate laterally and contaminate other basins where full coupling is employed. A method to reduce this error is to relax the SST to observed values so that the ocean model will remain stable throughout the simulation. In the experiments performed in this study, such a relaxation term is not employed.

This is due to the fact that we have used SST fields from a fully coupled simulation of the CFS as prescribed values as opposed to observed values. Additionally, our simulations are rather short, (30 years) to develop ocean circulation anomalies that reach other basins. We have also examined SST fields from the OGCM during the simulation and verified that no large anomalies are present.

A fully coupled 30-year long simulation of the CFS (Pegion and Kirtman 2008) is used in this study and will be referred as the control run. Details of the four regionally coupled simulations are provided below and shown in Table 2.1. In the first experiment, designated as IO-VSST (Indian Ocean forced by daily varying SST), daily mean SST from the control simulation are used to force the Indian Ocean (40°E-120°E) while the rest of the ocean basins are fully coupled. In this experiment, the atmosphere is allowed to respond to the Indian Ocean SST while the influence of atmosphere on the SST is suppressed. Comparing this run with the fully coupled control simulation, the influence
of the coupled air-sea feedbacks in the Indian Ocean can be studied. In the second simulation, designated as IO-CSST (Indian Ocean forced by daily climatological SST), the Indian Ocean is kept completely passive by prescribing daily climatology of SST from the control simulation while the rest of the regions are fully coupled. Comparing IO-VSST and IO-CSST with the control, the forced and coupled variability can be distinguished. In order to understand the influence of the western Pacific air-sea interaction, a third simulation in which daily SST is prescribed over the region (40°E-180°E) (designated as IOWP-VSST) is conducted. The fourth experiment, designated as PO-CSST, where the entire Pacific variability is suppressed by prescribing climatological SST in the region 120°E-90°W. the objective to isolate the Indian Ocean variability that is independent of the ENSO and its impacts on the monsoon.

In all the four simulations, the model is integrated for 30 years. Only one ensemble simulation is performed. The transition between the prescribed and coupled regions is smoothed by means of a 10° wide buffer zone (Fig 2.1) wherever appropriate. The SST values in the buffer zone are obtained by linear interpolation of prescribed and OGCM-produced values. In IO-VSST and IOWP-VSST, it is necessary to perturb the atmospheric initial conditions. We found that the initial conditions derived from the control run contained computer precision errors and they served as perturbations in our simulations. This was because the control and the regionally coupled simulations are carried out with different supercomputers.
2.3. Data

The observed datasets of OLR, precipitation, and SST are used to compare the simulation fields. The daily OLR data for the period 1979-2002 on a 2.5° x 2.5° grid was obtained from the National Oceanic and Atmospheric Administration (NOAA, Lieberman and Smith 1996). Observed rainfall came from the pentad data of the CPC (Climate Prediction Center) merged analysis of precipitation (CMAP) for the period 1979-2003 (Xie and Arkin 1996). The daily horizontal and vertical velocity fields are from NCEP-NCAR Reanalysis data for the period 1948-2000 (Kalnay et al. 1996). The SST data used is the weekly optimum interpolated fields by Reynolds and Smith (1994).

2.4. Analysis Methods

The main analysis used in this study is the multi-channel singular spectrum analysis (MSSA), which is the multivariate equivalent of the singular spectrum analysis (SSA). A detailed review of the SSA and the MSSA techniques from the perspective of the evolution of a nonlinear dynamical system and their advantages in the study of climatic time series can be found in Ghil et al. (2002) (see, also Plaut and Vautard 1994). A few important points from their review are reproduced here. The central objective of any climatic time series analysis is to identify regular occurrences of patterns, i.e., the periodic behavior underlying in the system, which bear the predictability. In the case of a chaotic system, although the variability appears to be completely random at first sight, it can be thought of as a combination of several quasi-periodic and quasi-stationary orbits that existed as perfectly periodic and stationary solutions in the system before it became
chaotic. That is, these solutions existed before the energy gained in the system crossed a certain threshold and triggered chaos. Those periodic and fixed point orbits remain in the chaotic attractor as unstable solutions or as “ghost limit cycles” and “ghost fixed points”, and considerable information of the behavior of the system can be obtained from them. One can think of these solutions forming the “skeleton” of the attractor, and the SSA or the MSSA as techniques to view the “skeleton” of the system.

Another way to understand them is as data-adaptive filters, which decompose the data into trends and oscillatory patterns that need not be linear. The oscillations can also be amplitude and phase modulated. Ghil et al. (2002) point out that “the SSA/MSSA can capture highly anharmonic oscillation shapes, which often require the use of many harmonics and subharmonics of the fundamental period when carrying out classical Fourier analysis”. In a simpler sense, the SSA isolates repeating sequences in a time series in their order of variance.

Mathematically, these methods are similar to the familiar empirical orthogonal function (EOF) and the extended EOF (EEOF) analyses. While the EOF provides spatial patterns of maximum variance and their temporal behavior, the MSSA provides information about the propagation of these patterns as well. The MSSA is identical to the Extended EOF (EEOF) analysis, such that both the methods are based on the eigenanalysis of the lagged covariance matrix. In EEOF analysis, the focus is given to the spatial information, and often a shorter temporal lag window is employed thus limiting the analysis of the spectral properties of the components. In MSSA, the lag windows are usually longer, common in non-linear dynamical system studies. The use of this method
to study the interannual and intraseasonal monsoon variability has been illustrated by KS07 and KS08.

Computational procedure of the MSSA can be briefly described as follows. Let the original data set contain $L$ spatial points (channels) at $N$ discrete time intervals. The lagged covariance matrices are constructed by choosing a certain lag window of length $M$ for each $L$ spatial point. The lagged covariance matrices for all spatial points are arranged to form a trajectory matrix of order $(LM, N-M+1)$, eigenanalysis of which yields $LM$ eigenvalues and $LM$ eigenvectors. The eigenvectors contain $M$ sequences of spatial maps and are referred as space-time EOFs (ST-EOFs). The space-time principal components (ST-PCs), each of length $N-M+1$, are obtained by projecting the original data on to the corresponding ST-EOFs. The component of the original data corresponding to each eigenvalue can be reconstructed by combining the ST-PC and its respective ST-EOF in a least square sense, and is referred as the reconstructed component (RC). The RCs share the spatial and temporal dimensions of the original dataset and can be considered as the filtered data corresponding to a particular mode.

Oscillatory signals, if exist, are resolved in two successive modes with same or close-by eigenvalues whose ST-PCs are in phase quadrature. The propagation of the oscillatory modes is represented by computing their phase composites based on the methodology by Moron et al. (1998). The spatial structure and amplitude of a non-oscillatory signal is examined by performing a spatial EOF (S-EOF) analysis of the corresponding RC.
Apart from the MSSA and EOF, correlation, regression and composite analyses are also utilized. Daily, monthly and seasonal anomalies are calculated by removing the respective climatology fields. The daily anomalies are smoothed by a 5-day running mean. This daily data, hereafter referred as unfiltered data, is used for most part of the analysis. A 20-100 day band pass filter is applied in certain cases where it was found to be difficult to isolate the intraseasonal modes in the unfiltered data.
Table 2.1: Summary of regionally coupled simulations

<table>
<thead>
<tr>
<th>Experiment name</th>
<th>Prescribed SST domain (ocean grid points within the region)</th>
<th>Prescribed SST type</th>
</tr>
</thead>
<tbody>
<tr>
<td>IO-VSST</td>
<td>40°E-120°E, 30°S-30°N</td>
<td>Daily mean from control run</td>
</tr>
<tr>
<td>IO-CSST</td>
<td>40°E-120°E, 30°S-30°N</td>
<td>Daily climatology from control run</td>
</tr>
<tr>
<td>IOWP-VSST</td>
<td>40°E-180°E, 30°S-50°N</td>
<td>Daily mean from control run</td>
</tr>
<tr>
<td>PO-CSST</td>
<td>120°E-90°W, 30°S-50°N</td>
<td>Daily climatology from control run</td>
</tr>
</tbody>
</table>
Figure 2.1. Regions of prescribed SST in IO-VSST and IO-CSST (top), IOWP-VSST (middle) and PO-CSST (bottom) simulations. The darker shaded area represents the 10° wide buffer zone.
Chapter 3 Summer Mean Climate and Variability in the Model

This chapter examines the ability of the coupled model to simulate the general features of the summer monsoon. The climatology and variability of the precipitation and circulation fields are computed from the control run and are compared with the observations. The objectives of this chapter are: (i) to demonstrate that the model simulates the mean climate and variability of the monsoon reasonably well and (ii) to point out any biases the model may have. All model fields shown in this chapter are from the control run.

3.1. Summer Climatology

The summer [June through September (JJAS)] climatology of the precipitation in the tropics is compared with observations in Fig 3.1. The model climatology is the mean of 30 summers and the observational field is based on the CMAP data for the period 1979-2003. In general, the model captures the main features of the summer precipitation climatology, such as the two maxima on either side of the Indian peninsula, rain shadow region over the southeastern India and the gross features of the inter tropical convergence zone (ITCZ) over the Pacific. A few deficiencies to note are the poor rainfall over the western Pacific (110°E-160°E, 0°-20°N), central India (north of 20°N and east of 75°E) and the relatively broad structure of the ITCZ over the Pacific. The model also has a
small region of enhanced precipitation over the equatorial Indian Ocean (west of 80°E), which is absent in the observations. The large anomalies of rainfall seen over the eastern part of the equatorial Indian Ocean (0°-10°S, 70°E-100°E) are also not reproduced well in the model.

Figure 3.2 depicts the JJAS climatology of horizontal winds at 850hPa and 200hPa levels. Although the direction of the low-level monsoon winds are well simulated in the model, the magnitudes are smaller compared to the Reanalysis fields (Fig 3.2a,b). The southwesterly winds that cross the west coast of India are about 10-12 m/s in the Reanalysis data, while the corresponding values in the model are 6-8 ms⁻¹. Two anticyclonic cells located along the equator at about 65°E and 95°E are missing in the model. This must be related to the cross equatorial flow which occurs between 40°-60°E according to the Reanalysis and between 60°E-80°E in the model. The meridional component of the southwesterly monsoon winds appears to be stronger in the model. Another point to note is the presence of a weak divergent flow in the model extending zonally over the northeast of India. The Reanalysis field shows no divergence in this region but has winds converging in the region (85°E-90°E, ~30°N). The upper-level airflow in the model is comparable to the reanalysis field although the model produces relatively weaker north-easterlies (Fig 3.2c,d).

The seasonal climatology of the SST in the model and its difference with the observation are shown in Fig 3.3. The model captures the main features of the SST variability such as the warm pool and cold tongue in the Pacific. However, the difference
maps show that the model has warm bias in the eastern Pacific and cold bias in north Pacific in all seasons. The western part of the Indian Ocean is warmer by 1-2 degrees during the summer and fall seasons. This must be related to the weak southwesterly monsoon winds discussed above.

Previous studies have suggested that an easterly vertical wind shear in the zonal mean wind is a critical factor in promoting the northward propagation of intraseasonal oscillations (Kemball-Cook et al. 2002). It has been suggested that the westward propagating intraseasonal oscillations are as a result of Rossby wave emanation from the western Pacific as Gill-type response to convection in that region (Wang and Xie 1997). The presence of mean easterly vertical wind shear was found to be important for the emanation of Rossby waves. Therefore, the vertical wind shear of zonal wind in the monsoon region computed as $U(200hPa) - U(850hPa)$ has been examined (Fig 3.4). Easterly wind shear is present over most of the monsoon region and western Pacific in both the observations and the model. However the wind shear is weaker in the model particularly over the western Pacific.

3.2. Annual Cycle

Figure 3.5a shows the annual cycle of the EIMR, which is the area-averaged rainfall over the region 70°E-110°E, 10°N-30°N. The annual cycle is computed from pentad climatology of rainfall in both observations and the model data. The annual cycle of the EIMR index is in good agreement with the observations. However, as noted in section 2, the seasonal climatology fields of the model showed considerable
overestimation of rainfall over the southwest and northeast of India and underestimation over the central India. This is further examined by plotting the area averages of pentad climatology of rainfall for the northeast (90°E-110°E, 20°N-30°N), southwest (73°E-76°E, 11°N-21°N) and the central India (79°E-85°E, 20°N-26°N) (Fig 3.5). These regions are marked in bottom panel of Fig 3.5. It is shown that in the model, the central Indian rainfall is underestimated by about 2 mm day$^{-1}$ while the northeastern and southwestern rainfalls are overestimated by about 8 and 4 mm day$^{-1}$ respectively. This suggests that the seemingly good correspondence of the EIMR index with the observations must be viewed with caution.

3.3. Intraseasonal Variability

The intraseasonal variability of summer precipitation is examined by computing the standard deviation of pentad rainfall anomalies and the active and break composites based on the EIMR index. Figure 3.6 shows the standard deviation of the pentad anomalies from the control run and the observations. The standard deviation of the daily precipitation anomalies from the model is also shown (Fig 3.6, middle panel). Firstly, there is no noticeable difference between the daily variability and pentad variability in the model simulation. Comparing the pentad variability of the model with that of the observations, the model captures the spatial pattern of the variability, which has maxima over the southwest coast of India, Bay of Bengal and the South China Sea area and minima over the rain shadow region in the southeast India. However, the variability is overestimated considerably over all the above regions. The maximum variability in the
CMAP data is 6-8 mm day$^{-1}$ in the Arabian Sea and Bay of Bengal, while the model has more than 10mm day$^{-1}$ standard deviation in those regions. The difficulty in simulating the central Indian rainfall is apparent in this analysis as well. The model has difficulty in simulating the narrow region of variability over the Pacific along the ITCZ.

Figure 3.7 shows the active and break composites of rainfall over the monsoon region. The active and break composites are computed following KS00. A period is considered active (break) when the standard deviation of the daily anomalies of the EIMR index stands above (below) $+(-) 0.5$ units for at least 5 consecutive days. Since the observed rainfall data is in pentads, the composites are calculated for the observations as follows. Active and break phases are identified from the daily anomalies of the OLR data for the period 1979-2002. The criterion discussed above is applied on the daily anomalies of the OLR to obtain the dates corresponding to active and break phases. The CMAP precipitation pentad anomalies corresponding to these Julian days are then averaged to obtain the composites. The OLR anomalies used here are 5-day running averages similar to all daily data used in this study. This method yields 1109 active and 1304 break days in the model and 669 active and 839 break days in the observations. Note that the observations have only 24 seasons while the model has 30 seasons.

The active composite in the observations is characterized by a northeast southwest tilted pattern of positive rainfall anomalies from 60°E-140°E and negative anomalies in the equatorial Indian Ocean. This spatial pattern is reversed during the break phase. The elongated rainfall belt is a characteristic of the northeastward propagating intraseasonal oscillations as presented by KS08. Note also that this pattern highlights the rain shadow...
area of the southeast India. The active and break composites of the model are comparable to the observations. However, the northeast southwest tilt of the rainfall anomalies is not captured well. The positive (negative) rainfall anomalies are limited to the region 60°E-120°E during the active (break) phase while the anomalies extend up to 160°E in the observations. The composites of the observations indicate that intraseasonal variability of the Indian monsoon rainfall is part of a large-scale variability present across the south Asian monsoon region. This is consistent with the spatial structure of 45-day mode described in KS08. In the model the anomalies are mainly over the Arabian Sea and Bay of Bengal. The space-time structure of the intraseasonal variability in the model is examined in detail in the next chapter.

3.4. Interannual Variability

The standard deviation of the JJAS seasonal anomalies of precipitation is presented in Fig 3.8. The model captures the two regions of high variability over India, namely, the western coastal belt and northeastern parts of the country although the variability is generally longer by 1 to 1.5 mm day\(^{-1}\) compared to observations. The variability over the northern parts (25°N-30°N, 75°E-80°E) is slightly underestimated in the model. The amplitude of the variability is largely overestimated over the Indian Ocean and most of the tropical Pacific. The model also shows very high standard deviation values over the west coast of Africa in the equatorial Indian Ocean. This error could be related to the coupled model’s simulation of the SST. A stand-alone simulation
of the GFS, the atmospheric component of the CFS, does not have this error in precipitation variability over the western equatorial Indian Ocean (Figure not shown).

Figure 3.9 shows the JJAS seasonal composites of strong and weak monsoon years based on the EIMR index. A season is categorized as strong (weak) when the standard deviation of the seasonal anomalies of the EIMR index is above (below) + (-) 1 unit. The seasonal composites from observed data shows that positive (negative) anomalies are present over most of the monsoon region in a strong (weak) monsoon season. The corresponding composites from the model data show patterns similar to the active and break composites. That is, positive anomalies are present over the Indian land region and negative anomalies over the Indian Ocean during a strong year and a revered pattern during a weak year. These results point out an important deficiency in the model in simulating the seasonal monsoon signals. Based on previous results (KS00), over the monsoon region the spatial structures of the interannual and intraseasonal variability are different.

Figure 3.10 shows the standard deviation of JJAS seasonal anomalies of the SST in the tropics. The model has a robust ENSO signature in the Pacific. However, the SST in the equatorial region extends slightly to the west of the dateline. The standard deviation of SST near the maritime continents and western Indian Ocean are also larger than observations.
3.5. Monsoon-SST Relationship

The influence of Pacific SST on the interannual variability of the monsoon is examined by computing composites of JJAS seasonal anomalies of precipitation based on the NINO3 index. The NINO3 index is the SST anomalies averaged over the region 90°E-150°W, 5°S-5°N. A season is classified as warm (cold) if the standardized JJAS mean NINO3 index is above +1 (below -1) standard deviation units. This criterion yields 4 warm and 2 cold seasons in the observed SST records for the period 1982-2002. The same criterion is applied over the 30 years of model data, which provided 5 warm, and 5 cold seasons. The warm and cold composites in Fig 3.11 are obtained by averaging the JJAS seasonal anomalies of precipitation over these selected seasons. Although there are exceptions, it is observed in general that a seasonal mean monsoon rainfall is below normal when eastern Pacific SST anomalies are abnormally warm and vice versa. This relationship can be seen in the warm and cold composites of observed precipitation shown in Fig 3.11. Similar composite maps using the model data, do not show this relation. It is found that in the model, during a warm (cold) ENSO event, precipitation anomalies are positive (negative) and therefore implies a positive correlation.

The ENSO-monsoon relation in the model is further examined by computing lead/lag correlations between JJAS seasonal anomalies of the EIMR index and monthly means of NINO3 index (Fig 3.12). This is a commonly used analysis in the literature (see, e.g., Krishnamurthy and Kinter 2003). It is known that the lead/lag correlation between the EIMR and the NINO3 reaches its maximum during the months October-November following the monsoon season. The simultaneous correlation between these
two indices is about -0.5. Here, we have used the Indian Monsoon Rainfall (IMR) index instead of the EIMR index to compute the observed correlation curve. The IMR index is the average of rainfall over the land grid points in India. An analysis using the EIMR index will yield similar results. It is noted that the model completely fails to reproduce the observed ENSO-monsoon relation. The correlation between the EIMR and NINO3 is weakly positive during and after the monsoon season.

This finding was verified by several independent analyses. For example, the lead/lag point correlations were computed between the JJAS seasonal anomalies of the EIMR and the tropical eastern Pacific SST anomalies in order to examine if there exists a negative correlation between the monsoon rainfall and SST anomalies of any part of the eastern Pacific (Figure not shown). The results were consistent with Fig 3.12. More sophisticated analyses using the MSSA are also conducted, the results of which are also consistent with the lag correlation analyses. These will be described in the next chapters.

3.6. Chapter Summary

This chapter examined ability of the coupled model used in this study to simulate the mean climate and the variability of the summer monsoon. The model captures the main features of the climatology and the intraseasonal variability. Two main problem found in the climatology are the excessive rainfall over the western equatorial Indian Ocean and poor rainfall over the western Pacific. The intraseasonal variability in the model is considerably overestimated and is more dominant over the Arabian Sea and the Bay of Bengal as opposed to a large-scale pattern that extends up to the western Pacific.
The spatial structure of the interannual variability of the monsoon is not consistent with that of the observations. The model appears to have difficulty in simulating a large-scale seasonal signal that is uniform over the equatorial Indian Ocean and the Indian land region. The spatial structure of the interannual variability appears to be similar to that of the intraseasonal variability. This indicates that the intraseasonal variability in the model is as a result of higher frequency oscillations and active and break cycles, which is not in agreement with the previous findings from observations (KS00). Another problem found is that the model fails to reproduce the observed negative correlation between monsoon rainfall and the eastern Pacific SST anomalies on the interannual time scale. The lack of a robust seasonal signal must be related to the poor ENSO-monsoon relation of the model.
Figure 3.1. JJAS seasonal climatology of precipitation in the CMAP observations (top) and the control runs of the model (bottom). The observations are based on the period 1979-2003 and the model climatology is for 30 years of simulation. Units are mm day$^{-1}$.
Figure 3.2. JJAS climatology of horizontal winds at 850hPa (top) and 200hPa (bottom) levels. Left panels are for the NCEP reanalysis (1948-2000) data and right panels are for the control run (30 years). Units are in $\text{m/s}$. 
Figure 3.3. Seasonal climatology of the SST in the control run (left) and the difference between control and observations (control-observations) (right). The units for the left panels are degree Celsius. The observed climatology is from the OISST for the period 1982-2002.
Figure 3.4. Contours are the JJAS climatology of vertical wind shear of the U wind ($U_{200}-U_{850}$) for the NCEP Reanalysis (top) and the control simulation (bottom). The reanalysis fields are for the period 1948-2000. Units are in ms$^{-1}$. Shading is JJAS climatology of rainfall for CMAP (1979-2003) (top) and the model control run (bottom). Units are in mm day$^{-1}$. 
Figure 3.5. Top panel: Pentad climatology of area averaged precipitation over (top left) EIMR region, (bottom left) central India, (top right) northeastern India, (bottom right) southwestern India for CMAP observations (black) and model control run (red). The y-axis is in mm day$^{-1}$. Bottom panel: The EIMR region (black box) and other three averaging domains (red boxes).
Figure 3.6. Standard deviation of daily (top) and pentad (middle) anomalies of precipitation for the control run and standard deviation of pentad anomalies of precipitation for the CMAP (1979-2003) data (bottom). Units are in mm day$^{-1}$. 
Figure 3.7. Active and break composites of precipitation for the model control run (top) and CMAP observations (bottom). Units are in mm day$^{-1}$. 


Figure 3.8. Standard deviation of JJAS seasonal anomalies of precipitation for the model control run (top) and the CMAP observations (bottom). Units are in mm day$^{-1}$. 
Figure 3.9. Strong and weak seasonal composites of precipitation based on the EIMR index for the control run (top) and the CMAP observations (bottom). Units are in mm day$^{-1}$. 
Figure 3.10. Standard deviation JJAS seasonal anomalies of SST for the control run (top) and the observations from OISST for the period 1982-2002 (bottom). Units are in mm day$^{-1}$. 
Figure 3.11. Composites of JJAS seasonal anomalies of precipitation based on JJAS NINO3 values for the control run (top) and CMAP observations (1982-2002) (bottom). Units are in mm day$^{-1}$. 

CFSCTL and CMAP: Seasonal composites of prec based on NINO3 index

mm/day

-3 -2 -1.5 -1 -0.5 0 0.5 1 1.5 2 3

-3 -2 -1.5 -1 -0.5 0 0.5 1 1.5 2 3
Figure 3.12. Lead/lag correlation between JJAS seasonal anomalies of the (E)IMR index and monthly anomalies of the NINO3 index for the control run (black) and observations (magenta). The observed curve is computed for the IMR index and the model’s curve is for the EIMR index. The observed analysis is for the period 1970-2001.
Chapter 4 Intraseasonal and Seasonally Persistent Modes in the Control Run

In this chapter, the intraseasonal and interannual variability in the control run are further examined using the MSSA. Technical details of the MSSA are provided in Chapter 2. Observational studies of KS07 and KS08 have illustrated the use of this analysis technique in isolating the intraseasonal and the seasonal signals from the daily data. Based on their results, the seasonal mean monsoon can be considered as a linear combination of the seasonally persistent components and the mean of intraseasonal signals. This finding is consistent with their earlier proposed conceptual model based on the Charney-Shukla hypothesis (KS00; Charney and Shukla 1981). In their analysis using the observed OLR data, KS08 found two seasonally persistent components and two intraseasonal oscillatory modes. The intraseasonal modes are of periods around 45 and 28 days, the spatial structure of which are described in Chapter 1. The two seasonally persistent (i.e., non-oscillatory within the span of the summer season) components were found to have good correlations with the ENSO and the Indian Ocean Dipole variability (KK09). The intraseasonal modes are only weakly correlated with the SST. The contribution of the persistent modes to the seasonal mean monsoon is considerably larger than that of the intraseasonal modes. However, the intraseasonal or high frequency components could be significant in years where the persistent signals are weak (e.g., a
normal eastern Pacific year). But, the MSSA provides us a way to dissect the different time scales of variability and their contributions to the seasonal anomaly.

In this chapter, we examine the dominant daily modes of precipitation in the monsoon region as simulated by the coupled model used in this study. The objectives of the analyses presented in this chapter are: (1) to examine whether the model successfully simulates the intraseasonal and seasonally persistent modes of the monsoon rainfall, and (2) to understand the relationship of these modes with the SST variability. It was mentioned that the model might have difficulty in producing the seasonal signals correctly over the monsoon region based on the composite and variance maps presented in the previous chapter. Therefore, it is essential to understand what contributes the seasonal signals in the control run, before we can interpret the results from regionally coupled experiments.

4.1. Dominant Daily Modes

Following the observational studies of KS08, we first performed an MSSA of lag window 61 days on the daily precipitation anomalies of the JJAS season from the model control run data. For a perfect model, this analysis should isolate persistent and intraseasonal signals dominant during the summer season. However, detailed examination of the modes obtained from the control run showed that this run failed to separate the persistent and oscillatory modes clearly. Spatial maps of the eigenmodes and variability of their time coefficients indicated the existence of a longer intraseasonal mode, which the current analysis could not resolve with a lag window of 61 days.
Therefore, a number of MSSA were performed on the same control run data by increasing the lag window length and consequently extending the dataset beyond the JJAS season. The results of these analyses were consistent such that they all isolated an oscillatory pair with period around 90-100 days. Results from one of these analyses that used a window length of 181 days and performed over all days of the year will be presented in the following sections. The focus of the present study is on the summer season, and therefore it is convenient to restrict our analyses to that season. It was also necessary to verify that the modes resolved in the all-season analysis mentioned above have significant amplitudes during the summer season. An all-season analysis may fail to isolate the summer signals since we conducted the analysis over a larger monsoon region (40°E-160°E, 35°S-35°N), which includes regions of high winter variability also. Due to these reasons, we conducted a second MSSA on the daily anomalies of rainfall of the JJAS season. The daily anomalies here are prepared by first removing the oscillatory signals from the all-season MSSA run. In summary, an all-season analysis was utilized as filter to remove the longer frequency oscillatory behavior seen in the model data. In addition to these two MSSA runs, another run on the 20-100 day band pass filtered daily data was also conducted for comparison purposes.

All the MSSA computations are conducted with daily rainfall anomalies. The domain of the study in all the analyses is 40°E-160°E, 35°S-35°N.
4.1.1. Power Spectra

Figure 4.1 (top panel) shows the power spectra of the space-time principle components (ST-PCs) of the dominant eigenmodes obtained from the all-season MSSA. Figure 4.1 (bottom panel) shows the singular values of the first 30 modes. The first mode is non-oscillatory within the 181-day lag-window length used here. The modes (2, 3) and (6, 7) constitute oscillatory signals with broad spectra centered at periods of 106 and 103 days respectively. The ST-PCs of the oscillatory modes are in near quadrature and their singular values are almost degenerate. The first 7 eigenmodes together constitute 5.5\% of the total variance in the daily data. The relatively low value of the variance explained is due to the fact that data is unfiltered.

Figure 4.2 shows the power spectra of the ST-PCs of the second MSSA performed in the JJAS data and their corresponding singular values. As mentioned in section 1, the data used in this analysis were obtained by removing the sum of the reconstructed components (RCs) of the modes 2, 3, 6 and 7 of the previous MSSA (all-season) from the total daily anomalies. The window length used in this second MSSA is 61 days, similar to that of KS08. The power spectra and singular value spectra show that modes 2, 3) and (4, 5) are oscillatory signals with periods around 57 and 30 days respectively. The first mode is again a seasonally persistent signal.

Finally, we present results from an MSSA performed on the 20-100 day band pass filtered data (Fig 4.3). The band pass filter was applied on the 5-day running means of total daily rainfall anomalies. The MSSA of this filtered data is limited to the JJAS season with a window length of 61 days. Similar to Fig 4.2, the power spectra show
evidence of two oscillatory modes with periods 60 and 30 days respectively. It will be shown in the following sections that the 60-day and 30-day modes obtained here are identical to the 57-day and 30-day modes presented in Fig 4.2.

4.2. Oscillatory Modes

In this section, we describe the spatial structure and propagation characteristics of the oscillatory modes obtained from the three MSSA computations described above.

4.2.1. Phase Composites of the Oscillatory Modes

The spatial structure of the oscillatory modes is described by computing the phase composites of the oscillation, details of which are provided in Chapter 2. Figure 4.4 shows the phase composite of the RCs of the 106-day mode isolated in the all-season MSSA (modes 3 and 4 in Fig 4.1). The phase composite depicts the evolution of the pattern in an average cycle of the nonlinear oscillation. Recall that the phase composite of the 45-day oscillation in KS08 accounts for part of the active and break cycle. The main feature of the spatial structure of the mode shown in Fig 4.4 is the large-scale rainfall anomalies from the Arabian Sea to the western Pacific which is evident during the peaks of the oscillations over India in phases 3 and 7. Anomalies are formed over the equatorial Indian Ocean shift slight eastward and then move northward to about 20°N. During the mature phase, three maxima are seen across the latitudinal belt 10°-20°N. This spatial structure is similar to that of the 45-day mode present in the observed OLR data.

The all-season MSSA presented in Fig 4.1 resolved another pair of oscillatory modes with a period of 103 days (n = 6, 7). The phase composites of this mode are shown
in Fig 4.5. It is evident that the precipitation anomalies are dominant over the southern and equatorial Indian Ocean in this mode as opposed to the Indian sub-continental area as in the modes 3 and 4. In one half cycle of the oscillation (i.e., phase 4 through 8), anomalies develop over the Indian Ocean between the longitudes 60°E-80°E and propagate eastward to the maritime continents (east of 120°E). Anomalies of opposite sign are developed at the eastern Indian Ocean in phase 8. The RCs of this mode were averaged over the regions north and south of the equator for all the 30 years of data in order to examine the relative amplitudes in the boreal summer and winter precipitation zones. It was clear that the amplitude of this oscillation is higher in the southern parts of the domain and during the boreal winter months. Similar averages computed for the modes 3 and 4 showed maximum amplitudes in the region north of the equator and during the summer months. Therefore, we conclude that the modes 3 and 4 are most likely a predominantly summer mode and the 6 and 7 are winter variability or the Madden Julian Oscillation (MJO). The rest of the study will focus on the modes 3 and 4 and they will be referred as the 106-day summer intraseasonal modes.

Figure 4.6 shows the phase composites of the RCs of 57-day oscillatory modes presented in Fig 4.2 (modes 2 and 3). The spatial structure of this mode bears several similarities to the 106-day summer mode shown in Fig 4.4 such as the northward propagation and the maxima on either side of the Indian peninsula. The amplitudes of the precipitation anomalies over the Arabian Sea and the Bay of Bengal are larger compared to the 106-day mode. Anomalies are formed over the Indian Ocean between 60°E-80°E (see phase 4 or 8) and grow in amplitude and spatial extent without any clear propagation
for parts of the cycle (phases 5, 6 or 1, 2). Some evidence of northward propagation is seen from phase 7 to 8 or 3 to 4. Large-scale east-west tilted anomalies are present in phases 4 and 8, near the end of the peaks of the cycle. Comparing the 106-day and 57-day modes, the latter has larger anomalies but a more local spatial structure predominant in the Arabian Sea and Bay of Bengal while the former is a more larger scale structure. Figure 4.7 shows the 60-day mode obtained from the MSSA of the 20-100 day filtered anomalies. It is clear that this mode is identical to the 57-day mode described above from comparing their respective phase composites.

In addition to the 106-day and 57-day modes, the model has another oscillatory component with period around 30 days. This mode was resolved in the two summer-season-MSSA runs presented in Figs 4.2 and 4.3. It was verified that the 30-day modes obtained in these two runs are identical. For brevity, we present the phase composites of the 30-day mode presented in Fig 4.2. The observational analysis of KS08 has shown the presence of a 28-day oscillatory mode in the OLR with a prominent quadruple structure in its anomalies. That is, positive anomalies were seen over India and the maritime continents while negative anomalies were seen over the equatorial Indian Ocean and western Pacific. We do not see a clear quadruple in the phase composites shown in Fig 4.8. However weak signs of a quadruple can be seen in phases 4 and 8, where positive (negative) anomalies are present over the Indian region and maritime continent and negative (positive) anomalies over the equatorial western Pacific (5°N-15°N, 100°E-130°E) and equatorial Indian Ocean (10°S-5°N, 60°E-80°E).
The robustness of all the above modes are further examined by plotting area averages of their corresponding RCs in the region 60°E-140°E, 0°-35°N against the total rainfall anomalies for the first 10 years of the simulation (Fig 4.9). These time series show that the 106-day mode captures variability in some years while the 57-day mode has more variability in some years.

Based on the discussion above, we summarize that the model has three dominant intraseasonal modes with periods around 106, 57 and 30 days. The propagation characteristics of these three modes are described in the next section.

4.2.2. Propagation of the Oscillatory Modes

The eastward and northward propagation of the modes are examined by averaging the phase composites of the RCs over two latitudinal and longitudinal belts, respectively. Figure 4.10 (left panels) show longitude-phase diagrams of the 106-day mode by averaging its RC across the latitudes 5°S-10°N and 10°N-25°N. There is clear eastward propagation near the equator from about 60°E to 160°E, while propagation signals are not present between 10°N-25°N. This is consistent with the behavior of the 45-day mode seen in the observations. Figure 4.10 (right panels) show latitude-phase diagrams of the 106-day mode by averaging across the longitudes 60°E-100°E and 100°-160°E. Northward propagation from 5°S to 20°N is over the Indian monsoon region (60°-100°E), which is consistent with the observations. However, the 106-day mode does not show northward propagation over the maritime continent and western Pacific while the 45-day mode in the observations shows a clear northward propagation in this region. The
lack of northward propagation in this region is evident in the phase composites of the 106-day model as well. Although this mode has an elongated stretch of anomalies from northwest India to the western Pacific, the northwest-southeast tilt of the pattern is not clear. From the phase composites of the 45-day mode as shown in KS08, the anomalies stretch from the latitude 15N in the Arabian Sea to 5N in the western Pacific. During the life cycle of the 45-day mode, the anomalies over the western Pacific move northward to about 20°N. However, the phase composites of the 106-day mode do not clearly show the northwest-southeast tilt of the rainfall anomalies. That is, the anomalies over the western Pacific are formed around 10°N and are seen to be expanding in spatial extent over the course of the oscillation with little northward movement.

The eastward and northward propagation of the 57-day mode is examined in Fig 4.11, with similar longitude-time and latitude-time cross-sections. The 57-day mode shows little evidence of eastward propagation except for the signals between 80°-100°E in Fig 4.11a. The northward propagation is more evident. Propagation is detected from the equator to 25°N over the Indian monsoon region (60°-100°E) and from 5°N to 15°N over the western Pacific (100°-160°E). These findings are consistent with the propagation properties of the observed 45-day mode.

These results show that the model has two northward propagating modes. The model’s 106-day mode exhibits a close resemblance to the observed 45-day mode in its spatial structure and the eastward propagation characteristics. However, this mode has a period longer than its observational counterpart, and also fails to capture the northward propagation over the western Pacific. The 57-day mode in the model shows some
evidence of northward propagation over the western Pacific. It is found that the amplitude of the 57-day mode is larger than that of the 106-day mode, particularly over the Arabian Sea and Bay of Bengal.

It is unclear which of these modes best represents the summer intraseasonal oscillation in the model. It is noteworthy that a similar all-season MSSA of the unfiltered daily anomalies of the observed OLR does not yield any lower frequency oscillations similar to the 106-day mode seen in the model (V. Krishnamurthy, personal communication). Therefore, it is clear that the existence of this low frequency oscillation is due to model deficiencies. The application of a time filter is a common practice in intraseasonal variability studies although it is not as good as the data-adaptive filtering of MSSA. Since it is well documented that the intraseasonal variability occurs within the time scale 20-60 days, one may find it convenient to use a filter in this range. However, as we found here, a model may have low frequency intraseasonal oscillations that would be completely eliminated by the application of a 20-100 day filter. That is, in this case, the application of the filter would restrict our analyses to the 57-day mode, which has poor eastward propagation. For example, Seo et al. (2007) point out that the intraseasonal oscillations of the CFS model has poor eastward propagation and attribute it to the biases in the Indian Ocean SST anomalies. They show that the propagation is improved by applying flux correction in the coupled model. While the SST errors may well be one of the sources for the poor eastward propagation, it is also important to note that the 106-day mode has better eastward propagation. The knowledge about the 106-day mode may
prove useful in improving the model’s intraseasonal variability and provides a cautionary note on the use of time filters.

The propagation features of the 30-day mode of the model are shown in Fig 4.12. The 30-day mode has clear westward propagation near the equator as well as north of 10°N and northward propagation over the Indian monsoon region and western Pacific. These figures are consistent with those of the 28-day mode found in the observed data (KS08) except for the westward propagation between 5°S-10°N. The observed 28-day mode has eastward propagation in that region. Another point of disagreement is that the northward propagation evident from the equator the 20°N in the model while it is dominant from 5°N to 20°N in the observations.

4.2.3. Oscillatory Modes and the SST

The relation between the intraseasonal modes and the SST is examined by computing daily point correlation between RCs averaged over the EIMR region and the daily anomalies of the SST during the JJAS season for the 30 summer seasons in the model run (Fig 4.13). For all the modes, the correlations are computed only for the daily anomalies of the JJAS season although the 106-day oscillation extends beyond the summer season. The 106-day mode shows moderate dependence on the eastern Pacific and Indian Ocean SST where correlation values range up to 0.3. Correlations greater than |0.03| are 5% significant. The 57-day mode has a similar correlation pattern as of the 106-day mode, but with even lesser values. Evidence of a northward propagation is seen in both of these modes as positive values over the northern Arabian Sea and Bay of Bengal.
and negative values over the equatorial Indian Ocean. The SST dependence of the 30-day mode is very weak where the maximum correlations are only 0.2. There is no Pacific influence for this mode. The weak correlations of the 30-day mode suggest that it is most likely an atmospheric mode of variability. These results are consistent with the findings of KK09, in which they show moderate correlations with the Indian and Pacific SST for the 45-day mode and negligible values for the 28-day mode.

### 4.3. Persistent Mode

The seasonally persistent signals resolved from the daily precipitation of the model run are described in this section. As presented in Figs 4.1 and 4.2, two non-oscillatory signals are resolved in the all-season and summer-only MSSA runs. For the reasons mentioned in section 2, we focus our analyses on the persistent signal obtained in the summer MSSA run (n=1 in Fig 4.2a). The persistence nature of this mode is verified by first examining the spatial maps of MSSA eigenvectors. The 60-day window length used in the MSSA gives 60 lagged spatial maps of the eigenvectors. It was verified that the anomalies of same sign persists throughout the window length. Alternately, following KS08, a new spatial EOF (S-EOF) analysis can be performed on the RC of this mode. For a persistent signal, the S-EOF will capture almost all the variance in the first mode.

#### 4.3.1. Spatial Structure of the Persistent Signal

Figure 4.14 (top panel) shows the first S-EOF of the RC1 of the persistent mode shown in Fig 4.1. The S-EOF captures 95% of the total variance in RC1. The high value of the fractional variance confirms that this is a single mode of variability. The spatial
structure of this mode has positive anomalies over the Indian Ocean (60°E-120°E, 15°S-
5°N) and negative anomalies over India. This structure is not exactly same as the ENSO
mode found in the observations which has a large-scale structure with same sign of
anomalies throughout the Indian monsoon region. The S-PC of the persistent mode and
daily means of NINO3 index are shown in Fig 4.14 (bottom panel). The correlation
between the time series is -0.54.

4.3.2. Persistent Mode and the SST

The relationship between the persistent signal and the global SST is examined by
computing point correlation between the S-PC1 of the RC1 and daily SST anomalies
during the JJAS season (Fig 4.15) The correlation pattern shows three centers of action:
negative correlation over the eastern Pacific, positive correlation over the western Pacific
and the eastern Indian Ocean and negative correlation over Western Indian Ocean. The
correlation values range from -0.6 to 0.6. This pattern is similar to the observed
relationship between the ENSO mode and the daily SST (KS09), although the east-west
dipole in the Indian Ocean is weaker in the observations. Another disagreement between
the observations and the model is the excessive westward extension of the negative
correlations in the Pacific, which must be due to the westward extension of the ENSO
anomalies in the model. It should be pointed out that the correlations are negative over
the eastern Pacific, which might appear inconsistent with the ENSO-monsoon
relationship on interannual timescale that was presented in Chapter 3. It was found that
the EIMR index is positively correlated with the NINO3 in that analysis. We would like

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to highlight the fact that the negative values over the eastern Pacific in Fig 4.15 arise because of the use of the S-PC1 instead of the EIMR index in the correlation analysis. It is also evident from the S-EOF1 of the persistent mode that the anomalies over the Indian region are of opposite sign of that over the Indian Ocean.

Although the model fails to capture the correct sign of the ENSO-monsoon relationship, it is clear that the persistent signal described above has strong dependence on the eastern Pacific SST. We computed point correlation values for the rest of the first 19 modes in order to examine whether there is any other ENSO related components in the daily data. No significant correlations were found for any of the 19 modes (Figure not shown). Therefore, the first mode is identified as the ENSO mode. We were unable to identify a mode an observational equivalent to the dipole mode in the model. EOF analysis of the SST shows that the dipole activity is predominant in the September-November season in the model as in the observations. The EOF analysis of the SST of the JJA months yields a pattern with same sign anomalies over most of the Indian Ocean.

4.3.3. Persistent Mode and the Seasonal Anomalies

The contribution of the ENSO mode to the seasonal anomalies is examined in this section. Figure 4.16a shows the EIMR index computed from the JJAS seasonal mean of the RC of the ENSO mode and the total seasonal anomalies of rainfall. It is noted that the ENSO mode fails to capture the seasonal variability of the monsoon in the EIMR region. Recall that the S-EOF of the RC1 captures the variance mainly over the eastern Indian Ocean region than over the EIMR region. The ENSO mode is therefore expected
contribute well to the seasonal anomalies over the eastern Indian Ocean. This is illustrated by defining an index called IOMR (Indian Ocean Monsoon region) by averaging the rainfall the region 60°-120°E, 20°S-0°. The IOMR index for the RC1 is compared with that of the total anomalies (Fig 4.16b). It is found that the seasonal anomalies over the IOMR region are solely contributed by the ENSO mode. The lack of a clear seasonal signal in the EIMR region is further addressed by computing standard deviation of the seasonal mean of the RCs of the first 20 MSSA modes and the residuals (Fig 4.17). The residuals are obtained by subtracting the sum of the first 20 modes from the total daily anomalies. No single mode among the first 20 modes was found to capture the interannual variability over India. The ENSO mode explains about 60% of the total interannual variance in the IOMR region. The first 20 modes combined explain only about 50% of the total interannual variability over the Indian subcontinent. It is however, the residuals of the MSSA, which is likely to be high frequency fluctuations that explain a large amount of interannual variability over the Indian subcontinent.

4.4. Chapter Summary

A detailed analysis of the interannual and intraseasonal variability in the control run has been presented in this chapter. It was found that the model has three modes of intraseasonal variability with broad spectra around 106, 57 and 30 days. Although the model has large daily variability in rainfall as shown in Chapter 3, the combined variance of coherently propagating oscillations is only 7% while the corresponding number in the observations is 11%. The 106-day mode has northward propagation and eastward
propagation and a large-scale structure extending across the south Asian monsoon region. The 57-day mode is northward propagating mode dominant over the Arabian Sea and Bay of Bengal. The 30-day mode has westward and northward propagation. The 106-day and 57-day modes have moderate correlations with the Indian and Pacific Ocean daily SST variability. A seasonally persistent component with clear ENSO influence is identified. The rainfall anomalies of the ENSO mode are dominant over the eastern equatorial Indian Ocean and weak over the Indian subcontinent. No single seasonal signal was obtained that captures the interannual variability of the rainfall over India.
Figure 4.1. MSSA of daily rainfall anomalies of all seasons combined for 30 years of the control run: (top) Power spectra of the ST-PCs of the eigenmodes 1 (red), 3, 4 (green) and 6, 7 (blue). The x-axis is period in days. (bottom) Singular value spectrum of the first 30 modes. Y-axis is singular values in log10 scale and x-axis is the order (n). Error bars denote 95% confidence interval for singular values.
Figure 4.2. MSSA of daily rainfall anomalies (that is devoid of modes 2, 3, 6, 7 in Figure 4.1) in the JJAS season for the control run: (top) Power spectra of the ST-PCs of the eigenmodes 1 (red), 2, 3 (green) and 4, 5 (blue). The x-axis is period in days. (bottom) Singular value spectrum of the first 30 modes. Y-axis is singular values in log10 scale and x-axis is the order (n).
Figure 4.3. MSSA of 20-100 day filtered daily rainfall anomalies in the JJAS season for the control run: (top) Power spectra of the ST-PCs of the eigenmodes 1, 2 (blue) and 3, 4 (green). The x-axis is period in days. (bottom) Singular value spectrum of the first 30 modes. Y-axis is singular values in log10 scale and x-axis is the order (n).
Figure 4.4. Phase composites of the 106-day oscillatory mode (n=3, 4 in Fig 4.1). Units are in mm day$^{-1}$. 
Figure 4.5. Phase composites of the 103-day oscillatory mode (n=6, 7 in Fig 4.1). Units are in mm day\(^{-1}\).
Figure 4.6. Phase composites of the 57-day oscillatory mode (n=2, 3 in Figure 4.2). Units are in mm day\(^{-1}\).
Figure 4.7. Phase composites of the 60-day oscillatory mode from the 20-100 day filtered anomalies (n=1, 2 in Fig 4.3). Units are in mm day$^{-1}$. 
Figure 4.8. Phase composites of the 30-day oscillatory mode (n=4, 5 in Fig 4.2). Units are in mm day$^{-1}$. 
Figure 4.9. Rainfall averages over the region 60-140E, 0-35N for total daily anomalies (grey, left axis), 106-day (red, right axis), 57-day (green, right axis) and for the sum of 100-day, 57-day and 30-day oscillatory modes (black, right axis) for the first 10 years of the control simulation. Simulation year is noted at the top right corner. Units are in mm day$^{-1}$. 
Figure 4.10. Longitude-phase cross-sections of the phase composites averaged between 5°S-10°N (top left) and 10°N-25°N (bottom left) for the 106-day mode. Latitude-phase cross-sections of the phase composites averaged between 60°E-100°E, (top right) and 100°E-160°E (bottom right) for the 106-day mode. Y-axis (x-axis) in left (right) panels represents phase angles for a complete oscillation. Units are in mm day$^{-1}$. 
Figure 4.11. Longitude-phase cross-sections of the phase composites averaged between 5°S-10°N (top left) and 10°N-25°N (bottom left) for the 57-day mode. Latitude-phase cross-sections of the phase composites averaged between 60°E-100°E, (top right) and 100°E-160°E (bottom right) for the 106-day mode. Y-axis (x-axis) in left (right) panels represents phase angles for a complete oscillation. Units are in mm day$^{-1}$. 

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Figure 4.12. Longitude-phase cross-sections of the phase composites averaged between 5°S-10°N (top left) and 10°N-25°N (bottom left) for the 30-day mode. Latitude-phase cross-sections of the phase composites averaged between 60°E-100°E, (top right) and 100°E-160°E (bottom right) for the 106-day mode. Y-axis (x-axis) in left (right) panels represents phase angles for a complete oscillation. Units are in mm day$^{-1}$. 
Figure 4.13. Daily point correlation between EIMR index of the RCs of the 106-day (top), 57-day (middle) and 30-day (bottom) oscillatory modes and daily anomalies of SST in the JJAS season for 30 years of the control simulation.
Figure 4.14. S-EOF1 of the RC of the ENSO mode (n=1 in Figure 4.2) (top) and its corresponding PC1 (bottom, black curve). The variance explained is 95%. Red curve in bottom panel is daily anomalies of NINO3. Daily values in the JJAS season are shown.
Figure 4.15. Daily point correlation between S-PC1 corresponding to the S-EOF1 of the ENSO mode (n=1 in Figure 4.2) and daily anomalies of SST in the JJAS season.
Figure 4.16. Top panel: EIMR index computed for the JJAS seasonal means of the RC corresponding to the ENSO mode (red) and total JJAS seasonal anomalies (black). Bottom panel: IOMR index computed for the JJAS seasonal means of the RC corresponding to the ENSO mode (red) and total JJAS seasonal anomalies (black). Y-axis is in mm day$^{-1}$. 
Figure 4.17. Standard deviations of JJAS seasonal means of RC1 (top), sum of first 20 RCs (middle) and residuals (bottom) plotted as a fraction of standard deviation of total JJAS seasonal anomalies. Regions where the total standard deviation is less than 1 mm day$^{-1}$ are left blank.
In this chapter, the role of the SST variability of the Indian, and Pacific Ocean basins in the monsoon intraseasonal variability is examined. The intraseasonal oscillations from four regionally coupled simulations are first discussed. The simulations are IO-VSST (Indian Ocean forced by daily varying SST), IO-CSST (Indian Ocean forced by climatological daily SST), IOWP-VSST (Indian Ocean and the western Pacific forced by daily varying SST) and PO-CSST (Pacific Ocean forced by climatological daily SST).

5.1. Climatology and Daily Variance

The JJAS summer climatology of rainfall from the four regionally coupled simulations and the control run are shown in Fig 5.1. There is no noticeable difference between the climatology of the control and the regionally coupled simulations. The annual cycles of the EIMR index in the experimental simulations show reasonable agreement with that of the control run (Fig 5.2). The seasonal climatology of the horizontal winds at 850hPa also shows close agreement among the five simulations (Fig 5.3). A small difference to note is that the winds crossing the Indian peninsula are slightly stronger (by 2mm day$^{-1}$) in the IO-CSST and IOWP-VSST runs. The climatology of the vertical shear of the zonal wind U is easterly over the monsoon region in all the five
simulations (Fig 5.4). The presence of mean easterly wind shear is documented to be critical for the northward propagation of the intraseasonal oscillations as discussed in Chapter 3. The standard deviation of 5-day running means and 20-100 day filtered rainfall anomalies are shown in Figs 5.5 and 5.6 respectively. The standard deviations are divided by that of the control run and plotted only for values greater than 6mm/day in the control run. In general, we find that the daily variability is increased over the Indian Ocean when SST is prescribed in that region (Fig 5.5). This is true whether the Indian Ocean is allowed to have daily variability or only the seasonal march of the SST. Over the western Pacific, the variability is increased when forced with daily varying SST, but the variability is slightly reduced when SST variability is suppressed to the annual cycle. The standard deviations plots of the filtered anomalies (Fig 5.6) show that in the regionally coupled simulations, there is very little increase in variability in the 20-100 day timescale over the Indian and western Pacific basins. This suggests that the increase in daily variability seen in Fig 5.5 must be due to high frequency fluctuations in the range of 5-20 days. The active minus break phase composites of the daily rainfall anomalies also do not show noticeable difference among the simulations (Fig 5.7).

5.2. Oscillatory Modes

The general procedure followed to identify the oscillatory signals is same as that used in the control simulation. As described in Chapter 4, three different MSSA runs are performed on the daily anomalies of precipitation over the domain 40°-160°E, 35°S-35°N. These MSSA runs are: 1) all-season analysis with a window length of 181 days, 2)
summer only analysis with a window length of 61 days, and 3) summer only analysis with a window length of 61 days performed on the 20-100 day filtered daily anomalies. Note that the second MSSA run utilizes a data set that is devoid of any low frequency intraseasonal oscillations that are resolved in the first MSSA run.

5.2.1. IO-VSST

The first MSSA run isolated an oscillatory component in eigenmodes 8 and 9. While the low ranking of these modes may raise questions about their significance, the oscillatory nature of these modes are verified by examining the phase shift of their ST-PCs and by computing the power spectra of the ST-PCs. The power spectra of the ST-PCs show a peak around 106 days, similar to that in the control simulation. The combined variance explained by modes 8 and 9 is 1%, which is comparable to the fractional variance of the 106-day mode in the control run.

The phase composites of the RC of the 106-day mode in IO-VSST (Fig 5.8) are somewhat similar to those of the 106-day mode found in the control run (Fig 4.4) although the amplitudes of the anomalies are weaker in IO-VSST. It is notes that the similarity is mainly over the northern parts of the domain while over the Indian Ocean and maritime continent features of the winter oscillations can be detected. We therefore suspect that this may be a mixed mode of variability that combines both summer and winter intraseasonal oscillations. Note that no other oscillatory signals were resolved as opposed to the control run where two pairs of oscillations with period around 100 days were obtained. It is also noted that the anomalies present over the maritime continents are
larger in amplitude compared to the 106-day summer mode found in the control run. A simple way to examine whether modes 8 and 9 represent both the summer and winter oscillations is to add the phase composites of the summer and winter modes obtained in the control run and examine whether the resulting pattern is similar to Fig 5.8. This procedure (Fig not shown) indicates that our speculation may be correct.

Hovmoller diagrams of the 106-day mode of the present experiment are shown in Fig 5.10. The longitude-phase composite averaged between 5°S-10°N shows that this mode has eastward propagation from 40°E to 80°E and then from 120°E to 160°E. There is some indication of eastward propagation from 120°E to 140°E between the latitudes 10°-25°N. Northward propagation is evident from the equator to 20°N over the Indian monsoon region but is absent over the western Pacific (Fig 5.10). These propagation features are similar to the 106-day summer mode obtained in the control run.

The second MSSA was performed on the daily precipitation anomalies of the JJAS season which was prepared by first removing the RCs corresponding of the 106-day mode of first MSSA from the total anomaly data. The power spectra of the ST-PCs of the first 20 modes show only modes 10 and 11 as an oscillatory pair with a period of 30 days. All the first 9 modes are non-oscillatory and appear to be persistent components. The third MSSA performed on the 20-100 day filtered anomalies resolve the leading two eigenmodes as an oscillatory pair with period around 27 days. By close examination of their power spectra and phase composites it is verified that the modes 10 and 11 obtained in the second MSSA and the modes 1 and 2 obtained in the third MSSA are identical. Since the oscillatory modes are better resolved in the filtered anomalies, we present the
phase composites and Hovmoller diagrams of the RC of 27-day mode obtained in that analysis (Fig 5.9). Positive anomalies are formed near the eastern end of the domain near the equator (phase 4) and propagate westward to the western Pacific and Bay of Bengal (phases 5 through 8) and then northward across the Arabian Sea and Bay of Bengal to reach about 15°N (phases 1-2). The Hovmoller diagrams of the phase composites also indicate westward and northward propagation across the domain (Fig 5.10). It is also noted that no mode equivalent to the 57-day, northward-propagating oscillations in the control run is obtained was this simulation.

Based on the discussion above, it can be summarized that the IO-VSST simulation has two intraseasonal signals with periods 106 and 30 days. It is noted that the 106-day mode was ranked as 8th and 9th eigenmodes indicating that this not the dominant variability in the 181-day time scale. However, we find that the eastward and northward propagation of this mode are coherent, as in the control. Similarly, the westward and northward propagation of the 30-day mode is similar to its counterpart in the control run. It is noted that the use of filtered data was necessary to isolate the 30-day mode as the first two eigenmodes since in the unfiltered anomalies they were resolved only as the 10th and 11th modes. These findings suggest that the air-sea interaction may not be an important factor in the propagation of these modes. There is no 57-day mode in this run equivalent to that found in the control run. In general, we found it was difficult to isolate significant intraseasonal signals from the unfiltered data.
5.2.2. IO-CSST

The first MSSA of the precipitation anomalies did not isolate any clear oscillatory modes for the IO-CSST simulation. The power spectra of the ST-PCs of the modes 4, 5, 6, 7 and 8 have peaks around 173, 167, 124, 108 and 75 days respectively. The power spectra of the modes 1 through 4 are red in nature. The second MSSA was performed by removing the RCs of the modes 4 through 8 from the total anomalies. It was found that there is no clear signal of oscillation in this analysis, although the modes 3 are 4 have peaks in their spectra. The third MSSA performed on the filtered anomalies show the first two modes have oscillations around 30 days. The phase composites of this 30-day mode are presented in Fig 5.11, which is very similar to the 30-day and 27-day modes obtained in the control and IO-VSST simulations. The Hovmoller diagrams also indicate westward and northward propagation as seen in the two previous simulations (Fig 5.12).

The results indicate that model may have difficulty in producing coherent signals of the eastward and northward propagating modes in the absence of daily varying SST. Note that the only intraseasonal signal obtained in this run is from the filtered data. It may be concluded that the daily variability of the SST is critical for the northeastward propagating modes. But it is noteworthy that the westward propagating signal with comparable spatial structure and propagation features is identified even with the climatological SST in the Indian Ocean. However, it should be mentioned that the westward propagation of this mode not evident to the west of 100°E where the SST is climatology in this run while the northward propagation is evident in the Indian and west Pacific sectors.
5.2.3. IOWP-VSST

As in the case on IO-CSST simulation, no clear oscillatory modes are obtained in this run also. The modes 5, 6, 7 and 8 have spectral peaks around 132, 106, 131 and 75 days in their respective power spectra but can not be considered as clean oscillations. The modes 1 through 4 are persistent in nature. The second MSSA, performed by removing the above-mentioned modes from the total anomalies shows that the modes 3 and 4 are in phase quadrature with an oscillation around 57 days. The phase composites of the RC of this model and their Hovmoller diagrams are shown in Figs 5.13 and 5.14. The spatial structure of the anomalies resemble the 106-day and 57-day modes obtained in the control simulation such that the anomalous rainfall extending from the Bay of Bengal to the western Pacific. However, this mode has no eastward propagation near the equator or north of 10°N. Evidence of northward propagation is seen between 100°-160°E, while standing oscillations are found between 60°-100°E. The third MSSA on the filtered anomalies show a similar mode as the one described above, although the period of the oscillations are around 45 days. The phase composites and Hovmoller diagrams of this mode are in agreement with that described above and therefore are not shown.

These results indicate that there is no clear propagating pattern in this run except for the 57-day mode, which has only northward propagation. The spatial structure of this mode is degraded when compared to the 57-day mode in the control run. Another point is that no westward propagating 30-day mode is found in this run, which indicates that the air-sea interaction over the western Pacific is crucial for its existence.
5.2.4. *PO-CSST*

In PO-CSST simulation, the all-season MSSA of 181-day lag window resolved an oscillatory mode (modes 1 and 2) with period around 85 days. In addition to this modes other oscillatory signals are obtained in modes (6, 7) and (11, 12). The power spectra of modes 6 and 7 show that they may not be considered as oscillatory pairs. Modes 11 and 12 have a period about 93 days and they appear to be the winter intraseasonal oscillations. The phase composites and Hovmoller diagrams of modes 1 and 2 are presented in Figs 5.15 and 5.17, respectively. The spatial structure of the 85-day mode is similar to the 106-day mode obtained in the control and IO-VSST simulations. There is evidence of eastward propagation between the longitudes 60°-100°E, while it is less coherent from 100°-160°E. Note that in this simulation, the SST variability is suppressed over the entire Pacific basin. Northward propagation is evident both in the Indian monsoon region and the western Pacific.

The second MSSA performed by removing the RCs of the modes 1 and 2 mentioned above yields an oscillation with period around 45 days, which are resolved in modes 2 and 3. A similar oscillation with period 38 days is found in the 20-100 day filtered anomalies. Although the period is slightly different, the spatial structure and propagation features of these two modes were found to be very similar. The Hovmoller diagrams of the 45-day mode show eastward propagation from 50°-80°E and westward propagation from 100°-160°E near the equator (Fig 5.17). Westward propagation is also evident north of 10°N. Northward propagation from the equator to 25°N is present both in the Indian Ocean and the western Pacific. It is difficult to interpret these results since
this simulation does not seem to have a clear westward propagating mode. It is unclear whether the 45-day mode shown here is a slower version of the 30-day northwestward propagating or a faster version of the 57-day northeastward propagating modes.

The results shown here suggest that the northeastward propagating mode exist with out Pacific SST variability. The spatial structure of this mode compares well with that of the control except over the China Sea, which includes the prescribed SST region in this simulation. But eastward propagation and northward propagation are consistent with its counterpart in the control. However, with climatological SST prescribed over the entire western Pacific, the westward propagating mode is less coherent. The phase composites and Hovmoller diagrams indicate that the 45-day mode obtained here could be a mixed mode of eastward and westward propagating oscillations.

5.3. Chapter Summary

This chapter examined the intraseasonal modes in the four regionally coupled simulations. In general, we found it difficult to isolate intraseasonal signals from the unfiltered data when the Indian or the west Pacific is uncoupled (i.e., in IO-VSST, IO-CSST and IOWP-VSST). This does not however indicate that the intraseasonal variability is reduced in the uncoupled versions. As shown in Figs 5.5 and 5.6, the daily variability is slightly increased or remained the same in the 20-100 day range in all the three simulations mentioned above. One argument that can be made is that the uncoupled versions lack coherent signals of propagation and the increase in the daily variability is due to high frequency fluctuations with no signal. Additionally, the interannual
variability is increased considerably in all these runs compared to the control. This will be shown in Chapter 6. Therefore, the first several dominant signals in the unfiltered MSSA computations are often non-oscillatory or persistent in nature, which indicates that the variance of the intraseasonal signals are reduced in the regionally coupled simulations. Nevertheless, oscillatory signals of varying periods are obtained in the experiments based on which some general conclusions can be drawn.

1) The northwestward propagation is consistently obtained in periods close to 30 days in the IO-VSST and IO-CSST runs. No noticeable differences were found either in the spatial structure or the propagation features of this mode whether the Indian Ocean is coupled or forced with varying or climatological SST. This suggests that the SST variability in the Indian Ocean is not important for the 30-day mode. This mode completely disappears when the SST is prescribed over the Indian and west Pacific Oceans, which suggests that the western Pacific air-sea interaction is crucial for its existence. Based on these results, one would expect that the 30-day mode would be absent in the PO-CSST simulations in which the entire Pacific is prescribed with climatology. Although no 30-day mode was isolated, evidences of westward propagation are seen in the 45-day mode obtained in that run. Based on these results it is difficult to state any firm conclusions about the dependence of the 30-daymode on the west Pacific SST.

2) Evidence of eastward propagation is found in the IO-VSST and PO-CSST simulations as oscillations of periods 106 and 85 days respectively. The phase composite of these modes compare well with the 106-day mode in the control run. It is found that
the eastward propagation signal coherent even without the air-sea interaction over the Indian Ocean. However the propagation appears to be weak when the underlying SST is climatology, i.e., in the PO-CSST run, the propagation is relatively weak to the east of 100°E (Fig 5.17). This is also consistent with the fact that no eastward propagating signals are identified in the IO-CSST run. However, the results from the IOWP-SST are more or less in conflict with the above discussion. In that run, no eastward propagation was found. Another point to note is that the period of the oscillation is decreased to 85 days.

3) The northward propagation over the Indian and western Pacific sectors is found in almost all cases examined in this chapter and therefore do not seem to be dependent on the underlying SST variability or air-sea interaction.
Figure 5.1. JJAS seasonal climatology of rainfall for 30 years of the control, IO-VSST, IO-CSST, IOWP-VSST and PO-CSST simulations (from top to bottom). Units are in mm day$^{-1}$. 

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Figure 5.2. Daily climatology of the EIMR index in mm day$^{-1}$ for the control and four regionally coupled simulations.
Figure 5.3. JJAS seasonal climatology of horizontal winds at 850hPa for the control, IO-VSST, IO-CSST, IOWP-VSST and PO-CSST simulations (from top to bottom starting from left panels). Units are in ms$^{-1}$. 
Figure 5.4. JJAS seasonal climatology of vertical shear of U wind for the control, IO-VSST, IO-CSST, IOWP-VSST and PO-CSST simulations (from top to bottom starting from left panels). Units are in ms$^{-1}$. 

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Figure 1. Standard deviation of the daily rainfall anomalies in the JJAS season for IO-VSST, IO-CSST, IOWP-VSST and PO-CSST simulations (from top to bottom) divided by corresponding values in the control run. Regions where the values are less than 6 mm day$^{-1}$ in the control run are left blank.
Figure 5.6. Standard deviation of the 20-100 day filtered daily rainfall anomalies in the JJAS season for IO-VSST, IO-CSST, IOWP-VSST and PO-CSST simulations (from top to bottom) divided by corresponding values in the control run. Regions where the values are less than 6 mm day$^{-1}$ in the control run are left blank.
Figure 5.7. Daily composites of active and break phases of the rainfall based on the EIMR index for the control, IO-VSST, IO-CSST, IOWP-VSST and PO-CSST simulations (from top to bottom starting from the left). Units are in mm day$^{-1}$. 
Figure 5.8. MSSA of daily anomalies of all seasons combined for the IO-VSST simulation: Phase composites of the 106-day oscillatory mode. Units are in mm day\(^{-1}\).
Figure 5.9. MSSA of daily anomalies of 20-100 day filtered anomalies in JJAS season for the IO-VSST simulation: Phase composites of the 30-day oscillatory mode. Units are in mm day$^{-1}$.
Figure 5.10. MSSA of rainfall for the IO-VSST simulation: Longitude-phase cross-sections of the phase composites averaged between 5°S-10°N and 10°N-25°N (four left panels) and Latitude-phase cross-sections of the phase composites averaged between 60°E-100°E and 100°E-160°E (four right panels). The top four panels are for the 106-day mode and the bottom four panels are for the 30-day mode. Units are in mm day$^{-1}$. 

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Figure 5.11. MSSA of daily anomalies of 20-100 day filtered anomalies in JJAS season for the IO-CSST simulation: Phase composites of the 30-day oscillatory mode. Units are in mm day$^{-1}$. 
Figure 5.12. MSSA of rainfall for the IO-CSST simulation: Longitude-phase cross-sections of the phase composites averaged between 5°S-10°N (top left) and 10°N-25°N (bottom left) for the 30-day mode. Latitude-phase cross-sections of the phase composites averaged between 60°E-100°E, (top right) and 100°E-160°E (bottom right) for the 106-day mode. Y-axis (x-axis) in left (right) panels represents phase angles for a complete oscillation. Units are in mm day$^{-1}$. 
Figure 5.13. MSSA of daily anomalies of 20-100 day filtered anomalies in JJAS season for the IOWP-VSST simulation: Phase composites of the 57-day oscillatory mode. Units are in mm day$^{-1}$. 

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Figure 5.14. MSSA of rainfall for the IOWP-VSST simulation: Longitude-phase cross-sections of the phase composites averaged between 5°S-10°N (top left) and 10°N-25°N (bottom left) for the 57-day mode. Latitude-phase cross-sections of the phase composites averaged between 60°E-100°E, (top right) and 100°E-160°E (bottom right) for the 106-day mode. Y-axis (x-axis) in left (right) panels represents phase angles for a complete oscillation. Units are in mm day$^{-1}$. 
Figure 5.15. MSSA of daily anomalies of all seasons combined for the PO-CSST simulation: Phase composites of the 85-day oscillatory mode. Units are in mm day$^{-1}$. 

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Figure 5.16. MSSA of daily anomalies of all seasons combined for the PO-CSST simulation: Phase composites of the 45-day oscillatory mode. Units are in mm day$^{-1}$. 
Figure 5.17. MSSA of rainfall for the PO-CSST simulation: Longitude-phase cross-sections of the phase composites averaged between 5°S-10°N and 10°N-25°N (four left panels) and Latitude-phase cross-sections of the phase composites averaged between 60°E-100°E and 100°E-160°E (four right panels). The top four panels are for the 85-day mode and the bottom four panels are for the 45-day mode. Units are in mm day$^{-1}$.
Chapter 6 Role of the Indian and Pacific Oceans in the Seasonally Persistent Signals

In this chapter, the role of the Indian and Pacific oceans in the interannual variability of the monsoon is examined using the regionally coupled simulations. First, the difference between the simulations and the control are examined by means of variance, correlation and regression analyses using the monthly or seasonal anomalies. Next, the seasonally persistent modes from the MSSA are used to understand the interannual variability of the monsoon. Specifically, we examine 1) whether and how the ENSO mode found in the control run changes in the experiments where the Indian Ocean and western Pacific SST is prescribed (i.e., in IO-VSST, IO-CSST and IOWP-VSST runs) and 2) whether there is any dominant seasonally persistent signal present in the monsoon rainfall in the absence of Pacific variability (i.e., in PO-CSST run). In the process, an explanation for the wrong ENSO-monsoon relation in the control run is provided.

6.1. Variance

Figure 6.1 shows the standard deviation of the JJAS seasonal anomalies of the precipitation for the four regionally coupled simulations. The standard deviation values are divided by that of the control simulation. In general, it is found that the variability over the Indian Ocean and the western Pacific is increased when the SST is prescribed
over that region. The increase in variability is mainly over the north of the equator in these oceans such as in the Arabian Sea, Bay of Bengal and northern west Pacific, while in the southern part of these basins, a moderate reduction or little change is seen. The rainfall variability is increased over the Pacific Ocean also when the Indian Ocean is uncoupled (IO-VSST). In the PO-CSST run, the variability increases mainly over the northern west Pacific and the eastern Pacific while it decreases over the eastern Indian Ocean. It has been shown that the atmospheric variability increases in general when the air-sea interaction is suppressed (Wu and Kirtman 2005) and is therefore consistent in the above results from IO-VSST and IOWP-VSST. That is, in the uncoupled run, an atmospheric variability that would have been damped in the coupled version due to the impact of atmospheric fluxes on the ocean (for example, through SST-cloud-radiation feedback) may be left to grow until the underlying SST forcing changes.

Figure 6.2 shows the standard deviation of the JJAS SST anomalies. Note that the ENSO variability is increased and extends farther into the west in all three simulations (IO-VSST, IO-CSST and IOWP-VSST). This appears to be in disagreement with the results of Wu and Kirtman (2005) where they found the ENSO variability is reduced by 20% in their experiment where the Indian Ocean was prescribed. However there are differences between the coupled model used in their study and the present one. They used a coupled model, which has the ENSO variability with a dominant biennial scale and correct ENSO-monsoon correlation. Using that, they found that the ENSO-monsoon correlation became positive when the Indian Ocean is uncoupled. As noted earlier, the coupled model used in this study has positive ENSO-monsoon correlation, and it will be
later demonstrated that the correlations become negative when the Indian Ocean is uncoupled. An argument can be made that the increase in the Pacific variability seen here is connected with the improvement in the ENSO-monsoon relation. This argument utilizes the results of Kirtman and Shukla (2000), in which they found that the seasonal anomalies of the monsoon winds act to strengthen an existing ENSO event, i.e., a weak monsoon strengthens an already developing El Niño and a strong monsoon a La Niña. Therefore, in a perfect model (here, one that simulates the ENSO-monsoon link correctly), the Pacific SST variability may have a component that is due to the monsoon anomalies. So in the model used by Wu and Kirtman (2005), the ENSO-monsoon link degrades when the Indian Ocean is uncoupled and therefore Pacific SST variability decreases. This argument can be reversed to explain the increase in Pacific variability in the uncoupled runs shown in Fig 6.2. That is, in our uncoupled runs, the ENSO-monsoon connection is improved which leads to an improvement in the monsoon’s impact on the ENSO, which is reflected in the increase in the SST variability there. The standard deviation of JJAS SST seasonal anomalies in the PO-CSST run plotted in Fig 6.3, shows a reduction in the variability over most parts. This indicates that the Indian Ocean SST variability is dependent on the ENSO signal in this model and is consistent with the atmospheric-bridge argument of Lau and Nath (2000).

6.2. ENSO-Monsoon Relation

Figure 6.4 shows the lead/lag correlation between the EIMR index of the summer (JJAS) seasonal anomalies of precipitation and monthly means of the NINO3 index. As
shown in Chapter 3, the correlations are weakly positive in the control run. It is found that the experimental runs produce the sign of the correlation correctly during the JJAS season although the details of the curve are far from perfect. For example, the correlations are unrealistically large before the monsoon season, where in nature they remain close to zero. However, the fact that the correlations are negative in the uncoupled simulations demands further investigations. Again, this result seemingly contradicts with the study of Wu and Kirtman (2004b), where they found that the ENSO-monsoon relationship is weakened when the air-sea interaction over the Indian Ocean is suppressed whereas in our study the relationship is improved when the Indian Ocean is uncoupled. An explanation for this apparent difference may be that in the model used in the present study, certain parts of the Indian Ocean is over-sensitive to the atmospheric fluxes (or the air-sea interaction mechanisms over the Indian ocean are wrong) and therefore the ENSO-induced anomalies are suppressed by air-sea interaction processes in the Indian Ocean. Therefore, the results presented here are not necessarily inconsistent with Wu and Kirtman (2004b), in that both the studies emphasize the role of air-sea interaction in the ENSO-monsoon connection. This point will be further illustrated using the eigenvectors from the MSSA.

6.3. Air-sea Interaction in the Indian Ocean

The impact of the air-sea interaction is examined by computing the point-to-point temporal correlation between the monthly precipitation anomalies of the control run and that of the IO-VSST and IOWP-VSST simulations (Fig 6.5) (Wu and Kirtman 2005).
Note the in these experiments, the prescribed SST consists of daily mean values of the control run. If the rainfall is a slave to the SST variability, both the experiments should produce exactly the same variability as in the control, which should be reflected as high positive correlations in Fig 6.5. It is seen that positive correlations of magnitude 0.4~0.5 are limited to regions over the eastern and western equatorial Indian Ocean. Over the rest of the Indian Ocean, particularly the Arabian Sea and the Bay of Bengal the correlations are less than 0.3. This indicates that the rainfall variability over these regions in the control simulation is a result of two-way air-sea interaction processes rather than an SST-forced atmospheric response.

Another method to examine the air-sea interaction is by computing the correlation between the SST and the precipitation (Wu and Kirtman 2005). The idea behind this analysis is that if the atmosphere acts only in response to the ocean, there should be simultaneous positive correlation between the SST and the precipitation. Figure 6.6 shows the point-to-point temporal correlation between monthly anomalies of the precipitation and the SST in the control, IO-VSST and IOWP-VSST simulations. In the control simulation, where air-sea interaction was allowed in all ocean basins, the correlation is strong over a broad region of the east Pacific. Weaker positive and negative correlations exist over the Indian Ocean, particularly over the Arabian Sea and Bay of Bengal and over the northern west Pacific, indicating the role of atmospheric fluxes in SST anomalies in those regions. In the IO-VSST and IOWP-VSST runs in which air-sea interaction is suppressed, the correlations over these regions are nearly 0.4~0.6. This is not surprising since these simulations would have only SST-forced response by design.
But, this illustrates the point that air-sea interaction plays a role in the rainfall variability in certain regions of the ocean and emphasizes the need for the coupled model.

6.4. Seasonally Persistent Signals

The seasonally persistent signals obtained from the MSSA in the control and four regionally coupled simulations are examined in this section. All modes described here are obtained from the MSSA of daily rainfall anomalies in the JJAS season, after removing the oscillatory modes found in the all-season MSSA. The analysis is performed over the region 40°E-160°E, 35°S-35°N with a lag window of 61 days.

6.4.1. Control

As described in Chapter 3, the control simulation has one seasonally persistent signal, which was identified as the ENSO mode. It was also noted that there is no single mode in the first 20 eigenmodes that captures the variability of the total seasonal anomalies. It was also found that the residuals of the MSSA defined as the sum of all modes higher than 20, captures the variability over the monsoon region. The EIMR index computed from RC1, sum of the first 20 RCs and the residuals are compared with that of the EIMR index of the total anomaly to restate this point (Fig 6.7). The sum of the first 20 RCs has a correlation of 0.59 with the total EIMR and the residuals 0.77. This shows that the monsoon seasonal anomalies over the EIMR region may be determined by higher frequency fluctuations in this model as opposed to a single low frequency climate signal.

The point correlation between the EIMR index from each of the first 20 modes and the daily SST were computed to determine if there are any SST-dependent signals.
other than the ENSO mode (Figure not shown). It was found that, apart from the ENSO mode, there is no seasonally persistent, SST-dependent signal in the control run. Recall that the MSSA of the observed rainfall had an Indian Ocean Dipole signal in addition to the ENSO mode (KS08). Point correlations between the daily DMI and RCs were computed and we find that the DMI has the highest correlation with the RC1. This indicates that the ENSO and dipole signature are in a single mode in the model. It can be interpreted from this that unlike the observations, the model has a dipole mode that always varies in conjunction with the ENSO mode. The lead/lag correlation between the JJAS seasonal mean EIMR index and the monthly NINO3 are computed for the RC1 (ENSO mode), sum of first 19 RCs ranked below the first, and the residuals (Fig 6.8). Note that the correlation curve for the residuals is in close match with the total anomalies. The RC1 has high correlations with the NINO3 although the sign is positive.

Based on the above discussion, it is argued that the failure of the ENSO-monsoon relation in the control run can be attributed to two factors: 1) The precipitation anomalies due to the ENSO do not reach the Indian subcontinent, rather concentrate around the eastern Indian Ocean. 2) The seasonal anomalies over India in this model are determined by high frequency fluctuation, which show no correlation with the SST.

The evolution of the ENSO mode is examined by computing lead/lag regression between the monthly mean S-PC1 of the RC1 and monthly anomalies of the precipitation, SST and winds at 850hPa. Figure 6.9 shows the regressed fields of precipitation for lags -1 to +4. Lag 0 indicates simultaneous regression for the months in JJAS while lag +1(-1) indicates the June S-PC1 regressed to July (May) fields and so on. Figure 6.9 shows that
the rainfalls anomalies form over the eastern Indian Ocean and grow in amplitude and spatial extent during the summer. Regressed fields up to lag +11 were examined (Figures not shown) and it was found that the precipitation anomalies over the eastern Indian Ocean persist up to the following winter and start to weaken after the peak phase of the ENSO. It is also noted that rainfall anomalies of opposite sign develop over the western Indian Ocean and peaks during September and October. Figure 6.10 shows similar regressed fields of the SST and horizontal winds at 850hPa. The SST anomalies are consistent with the rainfall anomalies such that warm SST anomalies co-locate with convection and vice versa. Easterly wind anomalies associated with the Pacific cooling flows to the Indian land area across the latitudes 5°N-15°N. Over the equatorial Indian Ocean, westerlies are present from the suppressed convection region in the western Indian Ocean to the enhanced convection region in the west Pacific. These westerly anomalies strengthen from lag +1 to lag +4 cooling the western Indian Ocean. Southeasterly flow across the southeastern Indian Ocean is also seen to be cooling the ocean along its path. Note that the westerly anomalies over the eastern Indian Ocean increases the southwesterly mean flow while the westerlies over the eastern Indian Ocean reduces the mean easterlies in that region. This difference in wind speed amplifies an already existing dipole pattern in the SST by air-sea interaction by inducing more mixing and cooling in the west and less mixing and thereby warming the ocean in the east. Due to the strong dipole in the SST, the winds flow parallel to the equator along the Indian Ocean producing no rainfall along the southwestern parts of the India. It is suspected that
this strong air-sea interaction mechanism over the western Indian Ocean may be the reason why the ENSO mode fails to reach the Indian subcontinent in the coupled model.

6.4.2. IO-VSST

The persistent signals from the IO-VSST run are examined in this section. The MSSA of the daily rainfall anomalies isolated two dominant modes in eigenmodes 1 and 2 as seasonally persistent signals. The persistent nature of these modes is verified by examining their eigenvectors and computing the power spectra of their ST-PCs. The S-EOF1 of the RCs of these modes are shown in Fig 6.11. The first EOF explains more than 90% of the variance in their corresponding RCs, which further confirms the persistent nature of these modes. The spatial pattern of the RC1 has maximum variance over the equatorial eastern Indian Ocean, maritime continents and Arabia Sea. Although the variance over India is small, this structure is noticeably different from the S-EOF1 of the ENSO mode in the control run (Fig 4.14). The amplitude of the negative anomalies over the Indian land region reaching across the western Pacific is small compared to the control run. The S-EOF1 of the RC2 shows maximum variance over the Arabian Sea and Bay of Bengal. The corresponding S-PCs show that the first mode has a correlation of about 0.7 with the NINO3 while the second mode has little correlation with the NINO3 (Figure not shown). Point correlation between the S-PCs and the daily SST (Fig 6.12) also indicates a strong ENSO signal in the first mode. It is noted that there is weak positive correlation between the S-PC1 of the RC2 and the SST, which is evident in the western Pacific. It is unclear at this point, whether this mode represents any physical
mechanisms. Since the RC1 appears to have a robust ENSO signal, we focus on this mode.

The contribution of RCs 1 and 2 to the total seasonal anomalies is examined by computing the EIMR index. Figure 6.13 shows the EIMR index for the RC1, sum of the RCs 1 and 2 (RC(1, 2)), sum of first 20 RCs and the residuals. It is evident that the RCs 1 and 2 combined captures most of the variability in the total anomalies. The correlation between these two time series is 0.7. The residuals have relatively small amplitude although they have moderate correlation with the total anomalies. This is different from what was seen in the control simulation. This point was further examined by computing the standard deviation of seasonal means of the RC1, sum of first 20 RCs and the residuals. Over the Arabian Sea and eastern Indian Ocean, the RC1 captures more than 60% of the interannual variability in the total anomalies. The sum of the first 20 modes captures more than 80% of the total variability in these regions. The impact of the residuals is minimal compared to what was observed in the control run (Figure not shown).

The ENSO-monsoon relation is examined by computing the lead/lag correlation between the JJAS seasonal EIMR and monthly NINO3 for RC1, RC2, sum of first 20 modes and the residuals (Fig 6.14). It is clear that the ENSO mode has improved in this simulation. The correlation curve of the RC1 shows a similar behavior as the observed ENSO monsoon relationship and also shows strong relation.

Based on the results discussed so far we can state that the MSSA captures the seasonal signals more effectively in the IO-VSST than in the control simulation. This is
consistent with the fact that interannual variability is increased in this simulation compared to the control.

Regressed fields of the precipitation, SST and the winds at 850hPa level on to PC1 of RC1 are shown in Figs 6.15 and 6.16. Strong rainfall anomalies are seen in the Arabia Sea and Bay of Bengal in lag -1, one month prior to the monsoon onset. Rainfall anomalies begin to weaken in July and disappear from the Arabian Sea by August. Similar fields of winds and SST show a consistent picture (Fig 6.16). Warm SST anomalies are present in most parts of the Indian Ocean throughout the season as opposed to a strong dipole in the control run (Fig 6.16). It is also noted that the wind anomalies blow from the from the eastern Indian Ocean to the southwest India as opposed to a near east-west flow found in the control run. In the case of the control run, ocean surface cooling by wind mixing and associated dipole formation was seen. In the IO-VSST, where the air-sea interaction is suppressed, the dipole pattern in the SST and precipitation has disappeared. These findings favor our earlier hypothesis that the stronger than normal air-sea interaction activity over the Indian Ocean may suppress the ENSO signal reaching India.

6.4.3. IO-CSST

The same set of computations as presented in the previous subsection is carried out for the IO-CSST simulation where the Indian Ocean is prescribed with climatological SST. It is found that the results are very similar to that seen in the IO-VSST simulation. As in the case of the IO-VSST, two persistent signals are isolated in modes 1 and 2 (Fig
Compared to the IO-VSST (Fig 6.11), one main difference is that the extent of the anomalies is smaller in the eastern Indian Ocean and the Arabian Sea in the RC1. The RC1 is identified as the ENSO signature by computing the correlation between its S-PC1 and SST (Fig 6.18). From Fig 6.19, which shows the seasonal means of the RCs over the EIMR region, it is found that the RCs 1 and 2 in combination captures most of the variability in the EIMR region. The correlation between the EIMR index from RC(1, 2) and the total anomalies is 0.8. the lead/lag correlation between the EIMR and NINO3 shows the RC1 captures he correct sign of the ENSO-monsoon relation (Fig 6.20) but with less resemblance to the observations than in IO-VSST case (Fig 6.14). Regressed fields of rainfall, SST and winds show a similar picture as seen in the IO-VSST run (Figs 6.21, 6.22). The rainfall anomalies associated with the ENSO persist mainly over the Arabian Sea and Bay of Bengal while the eastern Indian Ocean anomalies are almost nonexistent. As opposed to the control run, where the eastern Indian Ocean anomalies grow in amplitude through JJAS and persist beyond that season, in the IO-VSST and IO-CSST runs, the anomalies quickly weaken in that region. This indicates that air-sea interaction is critical for the growth of the ENSO-induced anomalies over that region.

6.4.4. IOWP-VSST

Results from the IOWP-VSST run are shown in Figs 6.23 through 6.28. These results are presented with a cautionary note. Recall that in this experiment, daily mean SST values from the control run are prescribed over the Indian and as well as western Pacific Oceans extending up to the date line. Since the coupled model has ENSO
anomalies that extend often to the west to the dateline, this experimental design may produce sharp transitions in the SST across the Pacific despite the 10° wide blending zone. The SST anomalies plotted for the JJAS season did not reveal such a discrepancy except in a few years and therefore we present the findings from that run.

The first two modes from the MSSA are examined by plotting their spatial structure and relationship with the SST (Figs 6.23, 6.24). The S-EOF1 of the RC1 has a structure with variance on either side of the Indian peninsula. Although this structure is similar to the patterns of the RC1 in the IO-VSST and the IO-CSST runs, an important distinction is the complete absence of variance over the eastern Indian Ocean. The spatial pattern of the RC2 is somewhat similar to the ENSO mode in the control simulation. Note that the variance is maximum over the eastern Indian Ocean. The correlation between the NINO3 and S-PC1 is 0.4 and the corresponding value for the S-PC2 is only -0.25. The point correlation between the S-PCs and the global SST shows that the RC1 has moderate ENSO dependence with the correlations in the eastern and the central Pacific reaching to 0.4. However, it is also found that the RC2 has strong dependency with the SST anomalies over the west Pacific. The lead/lag correlation between the EIMR index from the RCs and the monthly mean NINO3 (Fig 6.25) shows that the RC1 captures the correlation in the total anomalies. Figure 6.26 compares the EIMR index of the RCs with that of the total anomalies and it is seen that the RC1 captures the almost all of the variability in the total anomalies in the EIMR region.

The RC2 must be a mode that exists in the simulation due to the flaw in the experimental design that was mentioned earlier in this section. It appears from the point
correlation analysis that the RC2 is the weak forced response in precipitation due to the SST tri-pole structure existed across the Indian and western Pacific in the control run. Note that the SST from the control simulation is used to force the model in that region. Since air-sea interaction is suppressed over the Indian and west Pacific, the precipitation response is weak and does not grow into a strong dipole structure as in the case of the control.

Regressed fields of precipitation, SST and winds are shown for the RC1 in Figs. 6.27 and 6.28. The precipitation anomalies are formed in May (lag-1) and grow in amplitude in June and July and start weakening in August before almost disappearing in September. This evolution is similar to what was seen in the IO-VSST and IO-CSST runs but is different from the control run. As mentioned in previous sections, air-sea interaction is crucial for the growth of the ENSO-induced precipitation anomalies in the Indian Ocean sector. The regressed SST fields are consistent with the precipitation in that the warm anomalies co-locate with rainfall in May and June. As the SST starts to cool down over the eastern Indian Ocean in July (lag +1), the rainfall responds with weaker anomalies. The wind vectors show that southwesterly flow across the west coast of Africa is strong during lags -1 through +1 forming two cyclonic cells on either side of the Indian peninsula. By lag +3, the southwesterly winds become westerly and blow from the eastern Indian Ocean to the west thereby decreasing the rainfall over the Arabian Sea and Bay of Bengal.

One important distinction between the wind fields shown here and that in the previous runs of IO-VSST and IO-CSST is that the easterly anomalies from the Pacific
fails to reach the Bay of Bengal in this run. It is recalled that in the control run, easterly anomalies from the Pacific blow across the Bay of Bengal and central India weakening the mean monsoon flow. This feature was present in the IO-VSST and to some extent in the IO-CSST run. In the IO-CSST run, the easterly anomalies are weaker in comparison and do not persist throughout the monsoon season. Since the ENSO anomalies are limited to the region east of the dateline in the IOWP-VSST run, the wind anomalies from the Pacific reach only around 130°E. This helps in improving the ENSO-monsoon link in this run.

6.4.5 PO-CSST

The results presented so far have pointed to the direction that the air-sea interaction over the Indian Ocean is important for the amplification of the SST and rainfall anomalies in the Indian Ocean particularly in the eastern parts. It was also found that the western Indian Ocean is over-sensitive to the atmospheric fluxes and as a result a stronger than observed dipole-like pattern is formed in the SST and precipitation, which is responsible for the weak ENSO-monsoon link in the control.

In this section, results from the PO-CSST run are examined. The objective is to find whether there are any persistent modes of variability due to the SST variability inherent to the Indian Ocean and possibly due to air-sea interaction. The MSSA result of this run shows among the first 10 modes examined, there are only two components that have some dependency with the SST and they are modes 1 and 8. The S-EOF1 of the RCs 1 and 8 are shown in Fig 6.29. The RC1 has a pattern similar to an intraseasonal
mode with positive anomalies in the west coast of India, Bay of Bengal and parts of west Pacific. The RC8 shows a dipole-like pattern in the Indian Ocean but has little variance over India. The point correlation between the S-PCs of the RCs and SST is shown in Fig 6.30. A dipole structure is evident for the RC8 in the correlation pattern. The RC1 has mainly negative correlations over most of the basin except over a small region in the eastern Indian Ocean indicating a weak dipole-like pattern. The contribution of these modes to the total seasonal anomalies is examined by computing the EIMR index for the respective RCs and total anomalies (Fig 6.31). It is seen that RC1 captures almost all of the variability in the total seasonal anomalies over the EIMR region, while RC8 has negligible seasonal anomalies.

The evolution of the RC1 during the monsoon season is examined by computing the lead/lag regression between the monthly means of S-PC1 of the RC1 with the monthly anomalies of precipitation, SST and winds at 850hPa (Fig 6.32). The regressed fields of precipitation show relatively weak anomalies over India and strong anomalies over the Bay of Bengal and west Pacific. The anomalies start to weaken by August (lag +2) and completely disappear thereafter. The corresponding fields of winds (Fig 6.33) show southwesterly anomalies in lag -1, which weakens and become westerlies by lags +1 and 0. The SST field shows a cooling over most of the basin in response to the wind mixing and cloudiness from convection.

Figures 6.34 and 6.35 show similar regressed fields for the RC8. The precipitation and SST fields show a dipole structure developing as the summer season progresses. An east west dipole structure with warm SST and positive rainfall anomalies in the east and
cold SST and negative rainfall in the west is seen in month of May (lag +1). Correspondingly winds blow from the east to the west parallel to the equator. This pattern grows in amplitude and extent during the season with stronger wind anomalies and persists beyond the summer season. The amplification of these anomalies indicates air-sea interaction mechanism as seen in the case of the ENSO modes presented in the previous runs. Note that the westward anomalies seen in lag +1 increases the total wind speed in the eastern Indian Ocean since the climatological winds in that region is easterly. But the easterly anomalies in the western Indian Ocean, reduce the total wind since the monsoonal winds in that region is westerly. Therefore, once an east west dipole in SST and associated wind anomalies are established in the Indian Ocean as seen in lag +1 of Fig 6.35, the SST anomalies in the east and the west grow in amplitude due to more wind induced mixing in the east and less in the west. The amplified dipole structure in the SST increases the atmospheric pressure gradients and further increases the east west flow.

6.5. Chapter Summary

This chapter examined the interannual variability of the monsoon rainfall in a set of regionally coupled simulations. The objective was to examine the relative roles of the Indian and Pacific oceans in the seasonal mean monsoon. Our results indicate that the impact of the ENSO on the Indian Ocean is strongly dependent on the air-sea interaction in the Indian Ocean. It is demonstrated that the coupled model failed to produce the correct ENSO-monsoon relationship because of an over-active western Indian Ocean. Our results indicate that the western Indian Ocean in this model cools down quickly in
response to the ENSO-induced circulation anomalies thereby forming an east-west dipole in SST and precipitation in the Indian Ocean. This dipole prevents the ENSO’s impact from reaching the Indian monsoon region. The ENSO-monsoon relationship is more comparable to observations when the air-sea interaction is suppressed over the Indian Ocean. It is also found that the growth of the ENSO-induced convective anomalies over the eastern Indian Ocean depends on the air-sea interaction in that region. These anomalies, which persist beyond the summer season in the coupled simulation, disappear quickly in the uncoupled runs. In the absence of Pacific variability, monsoon rainfall appears to be controlled by atmospheric internal dynamics. A dipole mode is found in the Indian Ocean even in the absence of the ENSO forcing.
Figure 6.1. Standard deviation of the JJAS seasonal anomalies of the rainfall in the four regionally coupled simulations divided by the corresponding values in the control run. Regions where the standard deviation in the control is less than 2 mm day$^{-1}$ are masked. Unit is in mm day$^{-1}$. 
Figure 6.2. Standard deviations of the JJAS seasonal anomalies of the SST in the control and the three regionally coupled simulations.
Figure 6.3. Standard deviations of the JJAS seasonal anomalies of the SST in the PO-CSST simulation.
Figure 6.4. Lead/lag correlation between the JJAS seasonal anomalies of the EIMR index and monthly NINO3 in the control and the regionally coupled simulations. Grey shading represents the JJAS season.
Figure 6.5. Point correlation between monthly anomalies of precipitation during the JJAS season in the control and the IO-VSST (top) and between the control and IOWP-VSST (bottom) simulations.
CFS control and experiments

correlation between monthly Prec and SST for JJAS

Figure 6.6. Point correlation between monthly anomalies precipitation and SST during the JJAS season for the control, IO-VSST and IOWP-VSST simulations.
Figure 6.7. MSSA of the Control run: EIMR index computed from the JJAS seasonal means of the RC1 (top), sum of first 20 RCs (middle) and the residuals (bottom) (all in red) compared with the total seasonal anomalies of the EIMR index (black). Units are in mm day$^{-1}$. 

CFSCOLA (2007–2036) Prec: MSSA_FLT (ISO REMOVED) of daily (rm5) anomalies (JJAS) lag 61 JJAS EIMR index from RCs
Figure 6.8. MSSA of the Control run: Lead/lag correlation between the JJAS mean EIMR indices computed from the total precipitation anomalies (black), RC1 (red), sum of the first 20 RCs (green) and the residuals (blue) and the monthly NINO3 index. Grey shading represents the JJAS season.
Figure 6.9. MSSA of the Control run: lead/lag regression between the monthly means of S-PC1 of the RC1 and the monthly anomalies of the precipitation for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag 1(-1) denotes the June S-PC1 regressed to July (May) precipitation anomalies.
Figure 6.10. MSSA of the Control run: lead/lag regression between the monthly means of S-PC1 of the RC1 and the monthly anomalies of the SST (shading) and winds at 850hPa (vectors) for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) SST or wind anomalies.
Figure 6.11. MSSA of the IO-VSST run: S-EOF1 of the RC1 (top) and RC2 (bottom) normalized by their respective S-PCs. Units are in mm day$^{-1}$. Variance explained is noted in the top right corner.
Figure 6.12. MSSA of the IO-VSST run: Point correlation between the S-PC1 of the RC1 (top) and RC2 (bottom) and the daily SST anomalies.
Figure 6.13. MSSA of the IO-VSST run: EIMR index computed from the JJAS seasonal means of the RC1, sum of first two RCs, sum of first 20 RCs and the residuals (top to bottom, all in red) compared with the total seasonal anomalies of the EIMR index (black). Units are in mm day$^{-1}$. 
Figure 6.14. MSSA of the IO-VSST run: Lead/lag correlation between the JJAS mean EIMR indices computed from the total precipitation anomalies (black), RC1 (red), RC2 (green), sum of the first 20 RCs (blue) and the residuals (purple) and the monthly NINO3 index. Grey shading represents the JJAS season.
Figure 6.15. MSSA of the IO-VSST run: Lead/lag regression between the monthly means of S-PC1 of the RC1 and the monthly anomalies of the precipitation for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) precipitation anomalies.
Figure 6.16. MSSA of the IO-VSST run: Lead/lag regression between the monthly means of S-PC1 of the RC1 and the monthly anomalies of the SST (shading) and winds at 850hPa (vectors) for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) SST or wind anomalies.
Figure 6.17. MSSA of the IO-CSST run: S-EOF1 of the RC1 (top) and RC2 (bottom) normalized by their respective S-PCs. Units are in mm day$^{-1}$. Variance explained is noted in the top right corner.
Figure 6.18. MSSA of the IO-CSST run: Point correlation between the S-PC1 of the RC1 (top) and RC2 (bottom) and the daily SST anomalies.
Figure 6.19. M SSA of the IO-CSST run: EIMR index computed from the JJAS seasonal means of the RC1, RC2, sum of first 10 RCs and the residuals (top to bottom, all in red) compared with the total seasonal anomalies of the EIMR index (black). Units are in mm day\(^{-1}\).
Figure 6.20. MSSA of the IO-CSST run: Lead/lag correlation between the JJAS mean EIMR indices computed from the total precipitation anomalies (black), RC1 (red), RC2 (green), sum of the first 10 RCs (blue) and the residuals (purple) and the monthly NINO3 index. Grey shading represents the JJAS season.
Figure 6.21. MSSA of the IO-CSST run: Lead/lag regression between the monthly means of S-PC1 of the RC1 and the monthly anomalies of the precipitation for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) precipitation anomalies.
Figure 6.22. MSSA of the IO-CSST run: Lead/lag regression between the monthly means of S-PC1 of the RC1 and the monthly anomalies of the SST (shading) and winds at 850hPa (vectors) for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) SST or wind anomalies.
Figure 6.23. MSSA of the IOWP-VSST run: S-EOF1 of the RC1 (top) and RC2 (bottom) normalized by their respective S-PCs. Units are in mm day$^{-1}$. Variance explained is noted in the top right corner.
Figure 6.24. MSSA of the IOWP-VSST run: Point correlation between the S-PC1 of the RC1 (top) and RC2 (bottom) and the daily SST anomalies.
Figure 6.25. MSSA of the IOWP-VSST run: Lead/lag correlation between the JJAS mean EIMR of RCs and monthly NINO3 index. Grey shading represents the JJAS season.
Figure 6.26. MSSA of the IOWP-VSST run: EIMR index computed from the JJAS seasonal means of the RC1, RC2, sum of first 10 RCs and the residuals (top to bottom, all in red) compared with the total seasonal anomalies of the EIMR index (black). Units are in mm day$^{-1}$. 
Figure 6.27. MSSA of the IOWP-VSST run: Lead/lag regression between the monthly means of S-PC1 of the RC1 and the monthly anomalies of the precipitation for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) precipitation anomalies.
Figure 6.28. MSSA of the IOWP-VSST run: Lead/lag regression between the monthly means of S-PC1 of the RC1 and the monthly anomalies of the SST (shading) and winds at 850hPa (vectors) for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) SST or wind anomalies.
Figure 6.29. MSSA of the PO-CSST run: S-EOF1 of the RC1 (top) and RC8 (bottom) normalized by their respective S-PCs. Units are in mm day$^{-1}$. Variance explained is noted in the top right corner.
Figure 6.30. MSSA of the PO-CSST run: Point correlation between the S-PC1 of the RC1 (top) and RC8 (bottom) and the daily SST anomalies.
Figure 6.31. MSSA of the PO-CSST run: EIMR index computed from the JJAS seasonal means of the RC1, RC8, sum of first 10 RCs and the residuals (top to bottom, all in red) compared with the total seasonal anomalies of the EIMR index (black). Units are in mm day$^{-1}$. 
Figure 6.32. MSSA of the PO-CSST run: Lead/lag regression between the monthly means of S-PC1 of the RC1 and the monthly anomalies of the precipitation for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) precipitation anomalies.
Figure 6.33. MSSA of the PO-CSST run: Lead/lag regression between the monthly means of S-PC1 of the RC1 and the monthly anomalies of the SST (shading) and winds at 850hPa (vectors) for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) SST or wind anomalies.
Figure 6.34. MSSA of the PO-CSST run: Lead/lag regression between the monthly means of S-PC1 of the RC8 and the monthly anomalies of the precipitation for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) precipitation anomalies.
Figure 6.35. MSSA of the PO-CSST run: Lead/lag regression between the monthly means of S-PC1 of the RC8 and the monthly anomalies of the SST (shading) and winds at 850hPa (vectors) for the JJAS season. Lag 0 denotes simultaneous regression in the JJAS season. Lag +1(-1) denotes the June S-PC1 regressed to July (May) SST or wind anomalies.
Chapter 7 Summary and Conclusions

The objective of this study was to examine the role of the Indian and Pacific oceans in the monsoon intraseasonal and interannual variability. The intraseasonal and interannual variability are obtained by isolating oscillatory and seasonally persistent signals from the daily data using the MSSA technique. The methodology adopted for this study consisted of a set of regionally coupled simulations using the NCEP’s CFS CGCM. The simulations are IO-VSST (Indian Ocean forced by daily varying SST), IO-CSST (Indian Ocean forced by climatological daily SST), IOWP-VSST (Indian Ocean and the western Pacific forced by daily varying SST) and PO-CSST (Pacific Ocean forced by climatological daily SST). A fully coupled simulation of the CFS was considered as the control. The specific questions addressed in this study were:

1) What is the role of air-sea interaction in the Indian and west Pacific oceans in the northeastward and northwestward propagating summer intraseasonal modes?

2) What is the role of the air-sea interaction in the Indian and western Pacific in the seasonally persistent signals of the monsoon i.e., the ENSO and the dipole modes? Is there a seasonally persistent signal in the absence of the ENSO?

The first part of the study examined the intraseasonal oscillations and seasonally persistent signals in the fully coupled control simulation. It is found that the model has three intraseasonal oscillations with periods 106, 57 and 30 days respectively. The 106-
day mode has eastward propagation along the equator from 60°E to 160°E and northward propagation in the Indian monsoon region (60°E-100°E) from equator to 20°N. The spatial structure of this mode consists of positive (negative) anomalies extending from Arabian Sea to the western Pacific and negative (positive) anomalies over the equatorial Indian Ocean. These features are similar to the northeastward propagating mode in the observations except that the period of this mode is more than double its observational counterpart. The 57-day mode has a more local structure with maximum variance in the Arabian Sea and Bay of Bengal. While northward propagation is present in the Indian and west Pacific, the 57-day mode does not have eastward propagation. The 30-day mode has spatial structure and propagation features similar to the 28-day mode found in the observations. It is also worth mentioning that the combined variance of these modes is only about 7% while the corresponding number in observations is 11%, despite the large variance found in the daily data in the model. This indicates that the all of the daily variability do not contribute to coherent intraseasonal signals but rather to the high frequency fluctuations without any definite pattern in time and space.

A seasonally persistent signal with strong correlations with the eastern Pacific SST was resolved as the first eigenmode and was identified as the ENSO mode of the model. The spatial structure of this mode has maximum variance over the eastern equatorial Indian Ocean and little variance over the Indian monsoon region (70°E-100°E, 0°-35°N). This must be due to some deficiency in the model since the observed ENSO mode has variance across the entire south Asian monsoon region. It is also found that there is no low frequency climate signal that dictates the seasonal rainfall over the EIMR
region. As a consequence of this, the model fails to reproduce the observed negative correlation between the EIMR index and eastern Pacific SST anomalies. It was also found that an east-west dipole pattern in precipitation and SST in the Indian Ocean dominate the spatial structure of the ENSO mode.

The first goal of the study was to examine the role of SST variability in the Indian and Pacific oceans in the 106, 57 and 30-day modes found in the control run. It is found that the 106-day mode exists with weaker amplitude and less variance in the simulation in which the air-sea interaction over the Indian Ocean is suppressed (IO-VSST run). However, this mode was not obtained when the Indian Ocean SST was reduced to climatological values. It is argued that the 106-day mode in the IO-VSST run is a forced response in the atmosphere to the weak but coherent intraseasonal variability in the prescribed SST in that run. These results also indicate that daily SST variability above the climatological values is necessary for this intraseasonal mode. It was also found that the 106-day mode exists in the Pacific climatological simulation (PO-CSST) with comparable magnitude and variance as that of the control, which indicates that the ENSO variability does not influence this mode. The 30-day, northwestward propagating mode is obtained in the simulations where the Indian Ocean was prescribed with daily mean or climatological SST (IO-VSST and IO-CSST runs). But, this mode was not obtained when the western Pacific SST was prescribed (IOWP-VSST and PO-CSST runs). Based on these results, it is argued that the SST variability and air-sea interaction over the west Pacific are critical for the 30-day mode.
The second objective of this study was to examine the role of the air-sea interaction in the Indian Ocean in the interannual variability and specifically the ENSO mode. In general, the interannual variability of rainfall is increased over the regions where the SST is prescribed, which is consistent with the findings of Wu and Kirtman (2005). It can be argued that the increase in variability is due to the lack of negative feedback between the precipitation and SST, which operates in the nature and in the coupled model. For example, a positive rainfall anomaly and associated enhanced convection can induce a reduction in the incoming shortwave radiation and consequently cooling in the ocean surface, which would in turn reduce the rainfall anomaly. This coupled feedback cannot operate in an SST-prescribed simulation and therefore the rainfall anomalies would grow in amplitude until the underlying SST forcing changes.

The lead/lag correlation between the EIMR index and monthly NINO3 shows that the negative relationship between the eastern Pacific SST and monsoon rainfall is regained in the uncoupled runs (IO-VSST, IO-CSST and IOWP-VSST). This result, although at first appears to be in disagreement with that of Wu and Kirtman (2004b), can be argued to be consistent with their central idea, that the correct air-sea coupling over the Indian Ocean is critical for the ENSO-monsoon relationship. It is demonstrated that the ENSO-monsoon correlation is improved in the uncoupled runs, because the spatial structure of the ENSO mode is more comparable to observations and extends to the EIMR region and to the western parts of the Indian Ocean. The east-west dipole in SST and precipitation found in the ENSO mode of the control run disappeared in the uncoupled simulations. It is argued that this dipole is formed as a result of strong air-sea
interactions in the Indian Ocean. The western Indian Ocean in this model appears to be too sensitive to the atmospheric fluxes and therefore cools down quickly in response to the ENSO-induced circulation anomalies, creating an east-west dipole pattern. It is found that such a strong dipole can prevent the ENSO signals from reaching the EIMR region. On the other hand, the ENSO-induced rainfall anomalies over the eastern Indian Ocean are weak and disappear quickly in the absence of air-sea interaction over that region. These results point out the importance of correct air-sea interaction over the Indian Ocean for the simulation of the ENSO-monsoon teleconnection.

In the absence of an ENSO signal in the Pacific, the monsoon variability is mainly due to a persistent mode, which has maximum variance along the southwestern coast of India and the Bay of Bengal. This mode has good correlations with the SST in the Indian Ocean during the monsoon season although our analyses do not reveal an SST forcing operating in this mode. Instead, the SST anomalies are generated as a result of atmospheric forcing due to precipitation and wind anomalies. These anomalies do not persist beyond the monsoon season. A dipole in SST and precipitation is found even in the absence of the ENSO, although its contribution to the rainfall in the EIMR region is negligible. Evidence of strong air-sea interaction is seen in this mode as in the dipole pattern associated with the ENSO mode.

It is recalled that the central hypothesis that motivated the present study is that the interannual variability of monsoon can be considered as a linear combination of a two persistent modes, which are related to the ENSO and the Indian Ocean dipole variability respectively (KS08). Observational analyses have shown that these two modes can either
constructively interfere as in 1987 producing the above-normal rainfall or destructively interfere as in 1997. Our results show that a destructive interference between the ENSO and the dipole modes does not occur in the coupled model used in this study. Instead, the dipole and the ENSO modes always co-occur as in the case of 1987.

Finally, it is pointed out that the results presented here are limited to the model used in this study. Similar analyses using different CGCMs are necessary to establish these results. It is also noted that the present study faced limitations due to the biases in the model, mainly the presence of the 106-day intraseasonal mode and the lack of a dipole mode that is separate from the ENSO variability. The reasons for the existence of the 106-day mode are not clear from the current analysis. Further analyses focusing on the time scale selection of the intraseasonal mode as well as the precipitation-SST phase lag may provide insight into this problem. Similarly, as mentioned in the previous paragraph, the ENSO and the dipole modes always co-occurred in this coupled model and therefore an investigation as to what happens in years like 1997 could not be accomplished. It is also acknowledged that while our results provide ample evidence of the air-sea interaction in generating the dipole in the ENSO mode, further analysis of the ocean subsurface variables can strengthen our argument. In future, the ocean thermocline properties in the western Indian Ocean will be examined in detail. Further analyses are also necessary to explain the existence of the dipole variability in precipitation and the SST in the absence of the ENSO. The role of the western Pacific air-sea interaction is another point which could not be addressed based on the experiments performed in this study.
REFERENCES


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