On the Impacts of the Indian Summer Monsoon on ENSO in A Coupled GCM

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Abstract

This study explores the impacts of the Indian summer monsoon (ISM) on the El Niño-Southern Oscillation (ENSO) evolution by utilizing two coupled general circulation model simulations. The first simulation uses the Center for Ocean-Land-Atmosphere Studies (COLA) anomaly coupled model and the second simulation employs a new coupling strategy, interactive ensembles, that is designed to increase the signal to noise ratio. The effects of monsoon variability related and unrelated to ENSO are separated through composites in terms of both Niño-3.4 sea surface temperature (SST) and Indian summer monsoon rainfall (IMR) anomalies. It is found that ENSO-related monsoon variability has significant impacts on warm events. In the interactive ensemble simulation, a weak (strong) monsoon enhances (weakens) an ongoing warm event. The monsoon impacts are manifested in the surface zonal wind stress anomalies in the western-central equatorial Pacific. In the anomaly coupled simulation, the monsoon-ENSO relationship is difficult to detect. The ongoing cold events are only weakly affected by monsoon variability. Monsoon variability that is unrelated to ENSO also induces noticeable SST anomalies in the equatorial central Pacific in the following winter. In the interactive ensemble model, a weak (strong) monsoon induces noticeable warm (cold) SST anomalies.

The long-term change of the ISM-ENSO relationship is also analyzed. Fluctuations of the IMR-Niño-3.4 SST correlation are larger in the anomaly coupled model than in the interactive ensemble model. No apparent relation is found between the long-term change of the correlation and that of the monsoon and ENSO anomalies or variances. In the interactive ensemble simulation, the IMR and concurrent Niño-3.4 SST variances show in-phase long-term changes. In the anomaly coupled simulation, the long-term changes of the IMR and summer Niño-3.4 SST variances have an out-of-phase relation.

1. Introduction

El Niño-Southern Oscillation (ENSO) is one of the strongest signals in short-term climate variability. The Indian monsoon represents the typical seasonal signal and the anomalous monsoon heating has a large impact on local and remote circulations. These two climate systems are closely linked to each other although the processes for this connection are not well understood.

In this relationship, the Indian summer monsoon (ISM) has been historically viewed as a passive component and a lot of efforts have been made to investigate the processes by which ENSO affects the ISM variability (Webster and Yang 1992; Kawamura 1998; Nigam 1994; Ju and Slingo 1995; Soman and Slingo 1997; Yasunari and Seki 1992; Webster et al. 1998; Wang et al. 2002). On the other hand, there are indications that the ISM anomalies can impact subsequent ENSO evolution (Barnett 1984; Meehl 1987; Yasunari 1990; Webster and Yang 1992; Nigam 1994; Wainer and Webster 1996; Chung and Nigam 1999; Kirtman and Shukla 2000). Lag-lead correlation analysis shows that the Indian summer monsoon rainfall (IMR) has the largest negative correlation with the eastern Pacific sea surface temperature (SST) anomalies in the following winter (Yasunari 1990; Kirtman and Shukla 2000). Experiments with prescribed idealized monsoon heating in simple models suggest that a variable monsoon may contribute to the irregularity of ENSO (Wainer and Webster 1996). Chung and Nigam (1999) added the Asian summer monsoon heating anomalies into the Zebiak-Cane Pacific anomaly model (Zebiak and Cane 1987) and found that interactive Asian sector heating anomalies can increase the ENSO occurrence, strengthen some weak El Niños, and lead to a wider period distribution of ENSO variability. Based on results of

simulations with the Zebiak-Cane coupled model, Kirtman and Shukla (2000) found that a strong (weak) monsoon reduces (enhances) the amplitude of an ongoing warm event or enhances (reduces) the amplitude of an ongoing cold event. They also found that a variable monsoon serves as a trigger for ENSO. Thus, to enhance our understanding of ENSO variability and predictability, it is necessary to clarify the impacts of monsoon variability on the circulation in the Pacific and the associated SST anomalies.

A likely scenario for the impacts of ISM on ENSO is speculated as follows (e.g., Kirtman and Shukla 2000). The monsoon heating anomalies shift the Walker circulation and lead to anomalies in the Pacific trade winds. These wind stress anomalies change the surface heat flux, ocean current, and thermocline depth which then induces SST anomalies in the eastern-central equatorial Pacific. Though the Pacific wind stress anomalies induced by monsoon anomalies alone are small (Nigam 1994), they could be amplified through air-sea interaction in the tropical Pacific Ocean (Kirtman and Shukla 2000).

In reality, ENSO and the monsoon are integral parts of the coupled system and interact with each other. The monsoon anomalies include both ENSO-related and unrelated variations. Similarly, the Pacific SST anomalies likely have monsoon signals. This leads to difficulty in separating the wind anomalies induced by either ENSO or monsoon alone. Many previous studies examine ENSO or monsoon related anomalies based on composites with respect to a single index representing either ENSO or monsoon anomalies (e.g, Webster and Yang 1992). Because of the tendency for a warm (cold) event to concur with a dry (wet) monsoon (Meehl 1987; Webster and Yang 1992), the obtained anomalous circulation features display a high degree of similarity for these two

types of composites. Thus the composites cannot separate the anomalies induced by the monsoon from those induced by ENSO. Meehl (1987) pointed out that the composites of strong/weak monsoon years could be merely a reflection of signals in relation to cold/warm events. His comparison of SST and station sea level pressure composites for strong years which are cold events and for those which are not cold events demonstrated the similarity of spatial patterns, but the composites for the former years display the largest signal.

The monsoon anomalies concurrent with warm or cold events are likely ENSO-related. In this case, the monsoon feedback could be easily amplified through ongoing air-sea interaction in the tropical Pacific Ocean. Notable monsoon anomalies can appear in normal tropical Pacific conditions, and thus are ENSO-unrelated. In this case, the monsoon feedback might be less obvious, but may serve as a trigger for ENSO. Chung and Nigam (1999) considered the impact of ENSO-related monsoon heating anomalies on ENSO evolution through parameterization in terms of the concurrent ENSO SST anomalies. Wainer and Webster (1996) prescribed monsoon heating in a simple coupled model regardless of the current state in the Pacific Ocean. Such heating may be considered as ENSO-unrelated. Kirtman and Shukla (2000) parameterized or prescribed the monsoon induced Pacific trade wind anomalies in the coupled model. The anomaly wind stress pattern is derived based on an integration of an atmospheric general circulation model (AGCM) with climatological SST, and thus is necessarily independent of ENSO. The strength of the monsoon forcing is either prescribed (thus ENSO-unrelated) or parameterized in terms of contemporaneous SST anomalies in the eastern Pacific (thus ENSO-related). Since the monsoon induced wind stress anomalies are

applied in the summer when warm or cold events are developing in the tropical Pacific, in a coupled system they are likely subjected to feedbacks from ENSO and thus are not independent of ENSO.

In this study, we attempt to separate the Pacific wind anomalies induced by monsoon variability which are related to ENSO from those which are unrelated to ENSO. In order to do this, we make use of conditional composites. We consider strong/weak monsoon composites with or without anomalous conditions in the Pacific Ocean or warm/cold composite with normal or anomalous ISM. We also make comparisons of the two outputs of a coupled general circulation model (GCM), one anomaly coupled and the other interactive ensemble. Since the interactive ensemble model reduces the noise in the coupled system, such comparisons provide useful information regarding the effects of noise (internal variability) on the monsoon-ENSO relationship.

Another interesting topic is the long-term change of the ISM-ENSO relationship and its link with the long-term change of anomalies and variances in monsoon and Pacific regions. Observationally, the ISM-ENSO relationship showed an obvious weakening in recent two decades (e.g., Krishna Kumar et al. 1999). There have been attempts to understand the plausible reasons for this change (Krishna Kumar et al. 1999; Chang et al. 2001; Krishnamurthy and Goswami 2000; Kinter et al. 2002). Torrence and Webster (1999) showed that the ISM and ENSO have undergone interdecadal and decadal changes in variance and coherency over the last 125 years. Krishna Kumar et al. (1999) showed the out-of-phase relation in the long-term change of the IMR and Niño-3 SST until the late 1970s. Krishnamurthy and Goswami (2000) further noticed the in-phase change of the IMR and Niño-3 SST variances. However, the interdecadal change of the correlation

has no obvious relation with that of anomalies or variances. The weakening ISM-ENSO relationship was attributed to the southeastward shift of the Walker circulation in association with ENSO and the increased surface temperature over Eurasia in winter and spring due to global warming (Krishna Kumar et al. 1999) or the strengthening and poleward shift of the jet stream over the North Atlantic (Chang et al. 2001). Kinter et al. (2002) speculated that the recent change in the monsoon-ENSO connection is related to the atmospheric circulation over the entire North Pacific Ocean. Gershunov et al. (2001) found that the historic change of the relationship between ENSO and all Indian rainfall (AIR) on decadal time scales is significantly less variable than should be expected from sampling variability alone, suggesting that the decadal modulation of the correlation could be solely due to stochastic processes. We take advantage of outputs from two long period simulations of a coupled GCM to analyze the long-term change of the ISM-ENSO relationship. Because the two outputs have different degrees of noise, a comparison of correlation changes between the two simulations may give some idea about the roles that noise may play.

We describe in section 2 the data, models, and methods used in the present study. The mean and variability in the coupled model is described and compared to observations in section 3. Section 4 addresses the impacts of monsoon variability on the ENSO evolution based on composites derived from model outputs. The long-term change of the monsoon-ENSO relationship is discussed in section 5. Section 6 provides conclusions.

2. The data, models, and methods

a. The data

Observational data used in the present study include monthly mean Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) for the period of 1979-1999 (Xie and Arkin 1997). The reanalysis of the National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR) (Kalnay et al. 1996) provides monthly mean winds at 850 hPa and surface skin temperatures for the period of 1948-2001.

b. The models

We use outputs from two simulations of a coupled GCM: one anomaly coupled and the other interactive ensemble. The atmospheric model is the Center for Ocean-Land-Atmosphere Studies atmospheric GCM (COLA AGCM) which is a global spectral model with triangular truncation at wave number 42 and 18 unevenly spaced σ -coordinate vertical levels. Details of the COLA AGCM are documented in Kinter et al. (1997). For model changes, refer to Schneider (2001). The ocean general circulation model (OGCM) is version 3 of the Geophysical Fluid Dynamics laboratory (GFDL) modular ocean model (MOM3) (Pacanowski and Griffies 1998). Details of the anomaly coupling strategy are described in Kirtman et al. (1997) and Kirtman et al. (2002). The interactive ensemble coupling strategy was developed by Kirtman and Shukla (2002). Here, we only provide a simple description of the two coupling techniques.

In the anomaly coupled model, the atmosphere and ocean models are coupled through anomalies of heat, momentum, and freshwater fluxes. The anomalies are computed relative to their own model climatology. The ocean model climatology is

determined from an uncoupled extended integration with observed momentum flux and surface relaxation of temperature and salinity to observations. The atmosphere model climatology is determined from a multidecadal simulation with specified observed SST. The models are coupled once a day exchanging daily mean fluxes and SST.

In the interactive ensemble model, the same SST produced by MOM3 OGCM forces 6 realizations of COLA AGCM running in parallel. The ensemble mean fluxes (wind stress, heat flux, freshwater flux) from the 6 realizations of COLA AGCM are used to drive the MOM3 OGCM. Distinct from the traditional ensemble averaging of multiple realizations obtained *a posteriori*, the interactive ensemble technique allows the ensemble means of AGCMs to interact continuously with the OGCM as the integration evolves. The anomaly coupling strategy is also used in the interactive ensemble model. Kirtman and Shukla (2002) demonstrated that the interactive ensemble technique dramatically improves the simulation of the global teleconnection in association with ENSO, and the monsoon-ENSO relationship.

The coupled GCM was integrated for over 250 years. In this study, we use 265 years output from the anomaly coupled simulation and 322 years output from the interactive ensemble simulation. In the analysis of long-term changes of the monsoon-ENSO relationship (section 5), we extend the output from the interactive ensemble simulation to 406 years.

c. The methods

Most of previous studies make composites based on a single index, either ENSO or a monsoon index. Because of the high tendency for a warm (cold) event to occur contemporaneously with a dry (wet) monsoon, these types of composites can not separate

anomalies related to ENSO alone from those related to the monsoon alone. Because of this, we make a conditional classification based on both ENSO and monsoon anomalies. Here, we use the SST anomaly averaged over the Niño-3.4 region (170°W-120°W, 5°S-5°N) to represent ENSO. The June-September (JJAS) rainfall anomaly averaged in the region of 60°-100°E, 5°-25°N is used to denote the model ISM rainfall (or simply IMR) variation. Based on Niño-3.4 SST anomaly versus its standard deviation, we determine a specific year as a warm, normal, or cold year. Similarly, based on IMR anomaly and its standard deviation, we determine a specific year as a wet, normal, or dry year. We use the 0.43 standard deviation as the criterion to ensure that the three categories have nearly equal number of years assuming a normal distribution with a zero mean for both Niño-3.4 SST and IMR anomalies. The classification is sketched in Table 1.

According to the classification, the composite anomalies for the normal-wet (dry) years are induced by monsoon heating under the normal condition in the eastern-central equatorial Pacific, and thus are unrelated to ENSO. The warm (cold)-normal composites are unrelated to the monsoon. The warm (cold)-dry (wet) composites include the influence of both ENSO and the monsoon. The differences between warm (cold)-normal composites and warm (cold)-dry (wet) composites are attributed to the impacts of anomalous monsoon under the warm (cold) condition in the eastern-central equatorial Pacific, and thus represent ENSO-related monsoon anomalies.

The above classifications and composites are done for both coupled simulations. We made composites based on either JJAS or December-February (DJF) Niño-3.4 SST anomalies. In this study, we only report the results of composites based on JJAS Niño-3.4 SST anomalies.

3. The model climatology and variability

a. The climatology

Figure 1 shows the JJAS and DJF rainfall climatology derived from the CMAP observations and the anomaly coupled model. Since the interactive ensemble model uses the anomaly coupling strategy, the climatology is the same as the anomaly coupled model. In JJAS, the model simulates the eastern Pacific inter-tropical convergence zone (ITCZ) well. The model reproduces two large rainfall regions in the Arabian Sea and Bay of Bengal with a relative dry region in between, but these two regions are displaced away from the coast compared to CMAP. A systematic bias for JJAS rainfall is seen in the western and central North tropical Pacific where the model rain band is displaced northward of that observed. The model is dry over central and northern India and southeast China. The model South Pacific convergence zone (SPCZ) is eastward of that observed. The equatorial Indian Ocean rain band is south and west of that observed. Excessive rainfall appears over the western Atlantic Ocean.

In DJF, the eastern Indian Ocean-western Pacific is too dry along the equator, but too wet to the north and south. The SPCZ extends too far eastward. Excessive rainfall occurs in the eastern Pacific ITCZ.

The model mean SST distribution is very similar to the observed, indicating that the anomaly coupling is largely successful in preventing excessive climate drift. However, the SST in the warm pools (western Pacific, Indian Ocean, western Atlantic Ocean) is lower than observed by 0.5-1.0°C both in JJAS and DJF (Fig. 2). In DJF, the eastern equatorial Pacific cold tongue in the model is too strong and extends too far westward.

Figure 3 displays the JJAS and DJF winds at 850 hPa from the NCEP-NCAR reanalysis and the coupled model. In JJAS, the western Pacific subtropical high lies too far north. Consistent with this error, the westerlies over the Philippine Sea extend too far to the northeast. These features are related to the northward shift of the rain band in the western and central North Pacific. In the western-central South Pacific the trade wind shows an obvious southward displacement around the dateline so that the easterlies are located far too south in the western South Pacific compared to observations. Thus, the equatorial easterlies are too weak in the central and western Pacific. This is related to the discrepancy seen in the location of SPCZ.

In DJF, the westerlies in the eastern Indian Ocean through the western South Pacific are too strong and extend too far eastward, which corresponds to weaknesses in the model rainfall. Excessive convergence appears in the eastern Pacific ITCZ, which is consistent with the extremely strong rainfall simulated there. The central equatorial Pacific easterlies are again too weak.

b. The variability

Here, we mainly discuss the JJAS rainfall and DJF SST variability because JJAS is the Indian monsoon season and DJF is the mature phase of ENSO. We first describe the variability in the anomaly coupled model versus that in the observations. Then, we compare the variability between the anomaly coupled and interactive ensemble simulations. This gives some idea of how much noise is removed by the interactive ensemble coupling method.

Figure 4 shows the standard deviation of JJAS rainfall from CMAP, anomaly coupled model, and interactive ensemble model. The model rainfall variability displays a

low band along the equator in the eastern Indian Ocean-western Pacific. In the western-central North Pacific, the rainfall variability extends far too north to the subtropics. The variability in the western (eastern) South Indian Ocean is too large (weak). The variability center in the Arabian Sea and the Bay of Bengal extends too far westward, as does the large variability region in East Asia. Excessive rainfall variability appears in the western Atlantic Ocean. These discrepancies are related to those of mean rainfall seen in Fig. 1. In the western-central equatorial Pacific, the model rainfall variability center is weaker but extends far too east compared to observation.

Comparison of the standard deviation between the anomaly coupled model (Fig. 4b) and the interactive ensemble model (Fig. 4c) indicates how much of the variability is due to noise and how much is due to signal. It is expected that the interactive ensemble model would have considerably less variability than the anomaly coupled model. For convenience of comparison, we show in Fig. 4d the ratio of the JJAS rainfall standard deviation between the interactive ensemble and anomaly coupled model. In the central equatorial Pacific and tropical South Indian Ocean, the ratio is large with a magnitude over $2/3$. Over the tropical South Atlantic Ocean the ratio reaches 0.6. In the Bay of Bengal, the ratio is around $1/2$. The ratio in the Arabian Sea is somewhat larger than in the Bay of Bengal. In the mid-latitude North Pacific, the ratio is about 0.4-0.5. The spatial change of the ratio indicates that the signal is stronger in the central equatorial Pacific and tropical South Indian Ocean, but is weaker in the mid-latitude North Pacific and the Indian monsoon region (Bay of Bengal and Arabian Sea).

The standard deviation for DJF SST is shown in Fig. 5. Compared to observations, the equatorial Pacific SST variability in the model is shifted westward and

is weaker (about 2/3 of observed value) and the high variability band is also too narrow. The model SST variability in the equatorial western North Pacific is larger than the observed. In the mid-latitude North Pacific the model SST variability is close to observations. In the subtropical South Pacific the variability is generally too low.

The spatial distribution of DJF SST variability is very similar between the anomaly coupled model (Fig. 5b) and the interactive ensemble model (Fig. 5c). In comparison, the SST variability in the equatorial central Pacific is close for the two model simulations (Fig. 5d). In the southeastern tropical Indian Ocean, the ratio is large (0.7-0.8). The ratio is also large in the equatorial western North Pacific. In the North Indian Ocean the ratio is less than 0.6. In the mid-latitude North Pacific the ratio is large. The distributions of JJAS SST variability and ratio are similar to those in DJF except that the standard deviation is generally smaller and the ratio in the equatorial central Pacific is only around 0.7 (not shown). The spatial variation of the ratio indicates a robust signal in the equatorial central Pacific but mostly noise in the North Indian Ocean. The large ratio in the southeastern tropical Indian Ocean and mid-latitude North Pacific may suggest a contribution of air-sea interactions in those regions.

4. The impacts of monsoon on ENSO

In this section, we discuss the impacts of monsoon variability on ENSO evolution based on the composites described in section 2. Because of the long model outputs, there are a large number of cases for each type of composite, which makes composites robust. The numbers of years for different composites are shown in Table 2 for the interactive ensemble simulation (a) and the anomaly coupled simulation (b).

According to Table 2, in the interactive ensemble simulation there are many more warm-dry and cold-wet years than warm-wet and cold-dry years because of the strong negative correlation between IMR and Niño3.4 SST. In the anomaly coupled simulation there are more warm-dry years than warm-wet years, but the number of cold-wet years are close to the number of cold-dry years. As such, the ISM-ENSO relationship is very weak in the anomaly coupled model. This improvement of the monsoon-ENSO relationship in the interactive ensemble model was pointed out by Kirtman and Shukla (2002).

In the following, we first report the results from the interactive ensemble simulation in which the signal is more robust, followed by those from the anomaly coupled simulation. Both ENSO-related and ENSO-unrelated monsoon variability are considered. The separation of these two types of variability is based on the methods described in section 2. Our discussion will focus on changes in SST anomalies in the mature phase of ENSO, i.e., DJF.

a. The interactive ensemble simulation

We discuss the effects of ENSO-related monsoon variability first. This is demonstrated by the difference between the warm (cold)-normal composites and the

warm (cold)–dry (wet) composites. Figure 6a shows the composite SST anomalies along the equator (2°S–2°N) for warm-normal years in the interactive ensemble simulation. Figures b–d (e–g) show the differences of composite SST, heat content (hc), and zonal wind stress (taux) anomalies along the equator between warm-normal years and warm-dry (wet) years for the interactive ensemble simulation. Here, the ocean heat content (hc) is defined as, $hc = \int_0^H T dz / H$, where T is the ocean temperature and H is the depth of the upper ocean. $H = 250$ m is used in our calculation.

It is obvious that the warm events become warmer due to a weak monsoon (Fig. 6b), and cooler due to a strong monsoon (Fig. 6e). The monsoon induced SST anomalies in DJF are as large as 0.7°C for a weak monsoon (Fig. 6b). The SST anomalies are reduced by as much as 0.5°C for a strong monsoon (Fig. 6e). Consistent with these SST anomaly changes are large heat content anomaly differences propagating from central to eastern equatorial Pacific (Figs. 6c, f). In weak monsoon years, in the western equatorial Pacific westerly zonal wind stress anomalies are enhanced in late summer and the westerly differences intensify quickly around September and move eastward thereafter (Fig. 6d). These enhanced wind anomalies are associated with positive heat content anomaly differences (Fig. 6c). In contrast, in strong monsoon years, westerly anomalies are weakened after July and the weakened wind region moves eastward after October (Fig. 6g), followed by eastward propagating negative heat content anomaly differences (Fig. 6f). According to two-tailed Student t-test, the differences in the fall and winter after a weak monsoon are significant at 95% confidence level, suggesting the robustness of these changes. The differences in the winter after a strong monsoon are not significant, which indicates the large variability from case to case.

The cold events become weaker due to a weak monsoon (Fig. 7b), but stronger due to a strong monsoon (Fig. 7e). The monsoon induced SST anomalies, however, are much weaker compared to those in the warm events. The monsoon-related wind stress anomalies display a different evolution from those in warm events. The induced weakening of easterly anomalies in weak monsoon years is confined to the west (Fig. 7d). In strong monsoon years a noticeable enhancement of easterly anomalies occurs in central equatorial Pacific in summer and extends westward thereafter (Fig. 7g). These wind stress anomaly changes do not seem able to induce noticeable heat content anomaly changes (Figs. 7c, f) so that the SST anomalies only show small changes (Figs. 7b, e).

The impacts of a strong or weak monsoon on the Pacific SST anomalies when the contemporaneous Pacific SST anomalies are normal are demonstrated in Figure 8 which shows the composite anomalies along the equator in normal-dry years (a-c) and in normal-wet years (d-f) for the interactive ensemble model. Apparently, the monsoon induced SST, heat content, and wind stress anomalies are remarkable in DJF. A weak monsoon induces noticeable positive SST anomalies near the dateline (Fig. 8a). This effect is related to westerly anomalies developing in the western equatorial Pacific after July (Fig. 8c). The wind stress anomalies are weak during the monsoon season, but intensify quickly during the fall when local coupling takes over. The eastward extension of those westerly anomalies is limited likely due to the presence of persistent easterly wind anomalies in the central equatorial Pacific. As such, large positive heat content anomalies are mainly confined to the central equatorial Pacific (Fig. 8b). The negative SST anomalies induced by a strong monsoon are larger and have a wider zonal extension (Fig. 8d). Consistent large heat content anomalies propagate eastward (Fig. 8e). These

features relate to easterly anomalies extending eastward from the western to central equatorial Pacific (Fig. 8f).

b. The anomaly coupled simulation

Figure 9 shows the changes along the equator due to ENSO-related monsoon variability for the anomaly coupled simulation. The warm events are modulated by the weak monsoon in a similar way as those in the interactive ensemble simulation. The SST anomaly changes induced by a weak monsoon are as high as 0.7°C in January of the following year but in a small region (Fig. 9b). Noticeable eastward propagating heat content anomaly differences can be seen as well (Fig. 9c). The westerly wind stress anomalies are enhanced in August, experience a weakening in September, and then intensify quickly after October (Fig. 9d). The region of enhanced westerly anomalies also extends eastward. Compared to the interactive ensemble composites, the monsoon induced changes are smaller and less significant.

The effects of a strong monsoon differ from those in the interactive ensemble model and are generally small and insignificant. The SST anomaly differences in January-February after a strong monsoon are positive near 150°W but negative around 170°E (Fig. 9e). It seems that the SST anomalies are enhanced and shift eastward due to a strong monsoon. Heat content anomaly changes display a consistent distribution (Fig. 9f). The wind stress anomaly differences show more frequent fluctuations, especially, a switch from negative in August-September to large positive in October-November (Fig. 9g). A consistent transition is observed in the heat content anomaly differences in the central-eastern equatorial Pacific (Fig. 9f).

For both weak and strong monsoon years, the cold events display weak SST anomaly changes (Figs. 10b, e), which is similar to the interactive ensemble model. The heat content anomaly changes are weak as well (Figs. 10c, f). The wind stress anomaly changes display smaller scale features and more frequent switches of sign (Figs. 10d, g) compared to the interactive ensemble model.

Without ongoing warm or cold events in the Pacific, a weak monsoon induces negative SST anomalies in DJF (Fig. 11a), opposite to those in the interactive ensemble simulation. Negative heat content anomalies are maintained from October to the following March (Fig. 11b). Though there are westerly anomalies during August through December in the western equatorial Pacific, the effects of those westerly anomalies seem to be overcome by easterly anomalies in the central equatorial Pacific (Fig. 11c). In strong monsoon years, negative SST anomalies (Fig. 11d) develop in association with eastward moving easterly wind stress and negative heat content anomalies (Figs. 11e, f). Compared to the interactive ensemble case, these anomalies are weaker.

The interactive ensemble strategy provides a mechanism for isolating how the monsoon variability dynamically impacts ENSO. This interaction is difficult to detect in the anomaly coupled model because of excessive noise. However, it may be that the interaction is too strong in the interactive ensemble model compared to observations. Nevertheless, it does suggest how the monsoon could intermittently impact ENSO in the real world. Moreover, the monsoon induced large changes of SST anomalies in the ensuing winter explain the observed 6-month lead-lag monsoon-ENSO relationship.

5. The long-term change of the monsoon-ENSO relationship

The strongest negative correlation between the IMR and the eastern Pacific SST occurs when the SST lags by about 6 months (Yasunari 1990; Goswami et al. 1999; Kirtman and Shukla 2000). On the other hand, the influence of ENSO on the ISM is through simultaneous SST anomalies. In the following, we show the correlation of IMR with both contemporaneous and the following winter Niño-3.4 SST.

a. The long-term change of the correlation

Figure 12a (13a) shows the sliding correlation between the IMR and Niño-3.4 SST anomalies with a window of 31 years for the anomaly coupled (interactive ensemble coupled) simulation. There are apparent long-term changes in the correlation of IMR and Niño-3.4 SST either simultaneously or in the following winter. For the interactive ensemble model, the correlation with JJAS SST anomalies is generally weaker than with DJF SST anomalies (Fig. 13a). In comparison, the correlation change over time is larger in the anomaly coupled model than in the interactive ensemble model. For the correlation with DJF SST anomalies, the standard deviation is 0.22 in the anomaly coupled model and 0.14 in the interactive ensemble model. The larger standard deviation in the anomaly coupled model is expected since the noise is larger.

For the anomaly coupled model, though the correlation for the whole period is weak, high correlation appears in some epochs (Fig. 12a). For example, the negative correlation reaches -0.4 around year 2080 and -0.5 around year 2210. This suggests that for a coupled model study the plausible interdecadal change of the monsoon-ENSO relationship needs to be considered.

b. The long-term change of anomaly and variance

The interdecadal anomalies of IMR and Niño-3 SST show out-of-phase changes in observations (Krishnamurthy and Goswami 2000) but in-phase changes in a greenhouse warming scenario (Ashrit et al. 2001). The observed out-of-phase relationship is similar to that on interannual time scales. In the greenhouse warming scenario, the increase of mean monsoon rainfall is attributed to the enhanced land-sea thermal contrast because of a larger warming over the land than over the ocean. In the anomaly coupled model, there is no obvious relationship between the long-term changes of IMR and Niño-3.4 SST anomalies (Fig. 12b). In the interactive ensemble model, in-phase changes of IMR and JJAS Niño-3.4 SST anomalies can be seen in some periods, but not in other periods (Fig. 13b). The in-phase changes can not be attributed simply to the change of the land-sea thermal contrast because there are no large differences between the surface temperature changes over Eurasia and over the Indian Ocean in the model (not shown).

In observations, IMR and concurrent Niño-3 SST variances display in-phase interdecadal changes (Krishnamurthy and Goswami 2000; Torrence and Webster 1999). In the model, this relation is present in the interactive ensemble simulation (Fig. 13c). In the anomaly coupled simulation, IMR and JJAS Niño-3.4 SST standard deviations display an out-of-phase relation in the model period (Fig. 12c).

The warmer mean state in the eastern Pacific in recent decades is accompanied by increased ENSO activity (Krishnamurthy and Goswami 2000). In a global warming scenario, ENSO activity is enhanced (Ashrit et al. 2001). More variability with higher SST is evident in the interactive ensemble simulation (blue curves in Figs. 13b-c), but not in the anomaly coupled simulation (Figs. 12b-c).

c. Discussion of causes for the long-term change of ISM-ENSO relationship

Another question is the relation between long-term changes of the ISM-ENSO correlation and those of anomalies and variances. There is no consistent relation between the correlation change and the anomaly or variance change for both the anomaly coupled and interactive ensemble simulations. This is consistent with observations (Krishna Kumar et al. 1999; Krishnamurthy and Goswami 2000) and what occurs in a model greenhouse warming scenario (Ashrit et al. 2001).

Previous studies indicate that the breakdown of the ISM-ENSO relationship could be related to changes in ENSO effects and the land surface temperature effects, both of which are linked to mean state changes. We select two 30-year periods for each simulation, one with strong negative ISM-ENSO correlation and the other with weak or positive correlation, and compare the mean surface temperature differences in January-April (JFMA) and the mean surface wind difference in JJAS (Figures not shown).

For the anomaly coupled model, the surface temperature difference between years 2081-2110 (large positive ISM-ENSO correlation) and years 2121-2150 (strong negative ISM-ENSO correlation) is weak in the tropics and shows a dipole pattern over Eurasia with positive in the south and negative in the north. Thus, it seems that the correlation change in the anomaly coupled model has no clear relation with the Eurasian landmass warming.

For the interactive ensemble model, the surface temperature difference between years 2131-2160 (strong negative ISM-ENSO correlation) and years 2326-2355 (weak ISM-ENSO correlation) is negative in the eastern Pacific corresponding to a weakening of the correlation, which is opposite to observations. There is cooling over the Eurasian

landmass, which is only slightly larger than the tropical Indian Ocean. It seems that the correlation weakening in the interactive ensemble model cannot be ascribed to the change in the land-sea thermal contrast.

The interactive ensemble model shows a strong cooling and associated large cyclonic change over the North Pacific from years 2131-2160 to years 2326-2355, similar to the observed change around 1976, which has been proposed as a reason for the recent change in the monsoon-ENSO connection (Kinter et al. 2002). However, the southerly change over East Asia in the interactive ensemble simulation differs from the northerly change in Kinter et al. (2002). In the anomaly coupled simulation, the surface wind change from years 2081-2110 to years 2121-2150 displays a large anticyclone over the North Pacific corresponding to the intensification of the monsoon-ENSO correlation, while the flow change over east Asia is northerly.

Gershunov et al. (2001) proposed that sampling variability alone could induce changes in the IMR-ENSO correlation larger than that seen in observations. For the interactive ensemble model, the correlation for the whole period is about -0.48 with a standard deviation of 0.14. This value is lower than the 95% percentile of the bootstrapped standard deviation (see Table 1 of Gershunov et al. (2001)). For the anomaly coupled model, the correlation for the whole period is around -0.07 with a standard deviation of 0.22 which is still below the 95% percentile. This suggests the role of the noise in the long-term change of the ISM-ENSO relationship.

6. Conclusions

The COLA AGCM and MOM3 OGCM have been coupled through anomaly coupling and interactive ensemble coupling strategies. The coupled model has been integrated for over 250 years. The interactive ensemble coupling significantly reduces the noise in the Indian monsoon region. In the equatorial central Pacific and tropical South Indian Ocean, the apparent signal/noise ratio is large, thus the air-sea interaction plays a dominant role in determining the rainfall variability in those regions.

Utilizing long period outputs from the two coupled GCM simulations, we analyzed the impacts of monsoon variability on ENSO evolution. Through composites based on a novel classification which considers both the ISM and ENSO, we separated the monsoon induced anomalies which are related and unrelated to ENSO. We demonstrated that a strong or weak monsoon has significant effects on an ongoing warm event through surface zonal wind stress anomalies in the equatorial western-central Pacific. An ongoing warm event becomes stronger (weaker) in a weak (strong) monsoon year. Though the monsoon induced wind anomalies are smaller than those directly associated with warm events, they can amplify and propagate eastward through ongoing air-sea interaction in the Pacific Ocean. The cold events experience much weaker impacts from monsoon anomalies. The interactive ensemble composites have a more robust signal of monsoon impacts than the anomaly coupled composites do.

A strong or weak monsoon in normal ENSO years apparently can induce noticeable SST anomalies in the following winter. For strong monsoon years, the ensuing negative SST anomalies are larger in the interactive ensemble model than in the anomaly coupled model. For weak monsoon years, the ensuing SST anomalies are positive in the

interactive ensemble model, but negative in the anomaly coupled. This inconsistency is indicative of the model sensitivity to the variability of wind stress anomalies in the western-central equatorial Pacific.

The anomaly coupled simulation displays larger fluctuations in the correlation between IMR and Niño-3.4 SST anomalies than the interactive ensemble simulation does. No apparent relationship is found between the long-term change of correlation and the long-term changes of anomalies or variances in the monsoon and tropical Pacific regions. The change of the ISM-ENSO relationship does not seem related to model mean state changes. It is plausible that noise plays an important role in the long-term change of the ISM-ENSO relationship.

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Table 1. Classifications based on Niño-3.4 SST and IMR anomalies. In the table, IMRA and SSTA denote IMR and Niño-3.4 SST anomalies, respectively. σ_{IMR} and σ_{SST} refer to standard deviation of IMR and Niño-3.4 SST anomalies, respectively.

		Niño-3.4 SST		
		Cold ($\text{SSTA} < -0.43 \sigma_{\text{SST}}$)	Normal ($ \text{SSTA} < 0.43 \sigma_{\text{SST}}$)	Warm ($\text{SSTA} > 0.43 \sigma_{\text{SST}}$)
IMR	Wet ($\text{IMRA} > 0.43 \sigma_{\text{IMR}}$)	Cold-Wet	Normal-Wet	Warm-Wet
	Normal ($ \text{IMRA} < 0.43 \sigma_{\text{IMR}}$)	Cold-Normal	Normal-Normal	Warm-Normal
	Dry ($\text{IMRA} < -0.43 \sigma_{\text{IMR}}$)	Cold-Dry	Normal-Dry	Warm-dry

Table 2 Number of years for different composites

(a) Interactive ensemble simulation

	Cold	Normal	Warm
Wet	40	41	28
Normal	32	41	32
Dry	23	36	49

(b) Anomaly coupled simulation

	Cold	Normal	Warm
Wet	24	35	23
Normal	34	29	38
Dry	27	20	35

Figure 1. Mean JJAS and DJF rainfall derived from the CMAP observation and COLA anomaly coupled model.

Figure 2. Mean JJAS and DJF surface skin temperature (Tsfc) derived from the NCEP-NCAR reanalysis and COLA anomaly coupled model.

Figure 3. Mean JJAS and DJF 850 hPa winds derived from the NCEP-NCAR reanalysis and COLA anomaly coupled model.

Figure 4. Standard deviation of JJAS rainfall derived from the CMAP observation (a), COLA anomaly coupled model (b), and COLA interactive ensemble coupled model (c), and the ratio of the standard deviation between interactive ensemble and anomaly coupled model (d). The contour interval is 0.4 mm/day in (a)-(c) and 0.1 in (d). Shaded regions indicate standard deviation over 0.4 mm/day in (a)-(c) and ratio over 0.5 in (d).

Figure 5. Standard deviation of DJF Tsfc derived from the NCEP-NCAR reanalysis (a), COLA anomaly coupled model (b), and COLA interactive ensemble coupled model (c), and the ratio of the standard deviation between interactive ensemble and anomaly coupled model (d). The contour interval is 0.2°C in (a)-(c) and 0.1 in (d). Shaded regions indicate standard deviation over 0.2°C in (a)-(c) and ratio over 0.5 in (d).

Figure 6. Composite sea surface temperature anomalies (SSTA) in warm-normal years (a), and differences of composite SSTA (b), heat content anomalies (hcA) (c), and zonal wind stress anomalies (tauxA) (d) between warm-normal and warm-dry years along the equator (2°S - 2°N average) in the interactive ensemble coupled model. The differences are obtained by subtracting warm-normal composites from warm-dry composites. (e)-(g) are similar to (b)-(d) except for differences between warm-normal and warm-wet years. The contour interval is 0.2°C in (a)-(c) and (e)-(f), and 0.04 dyn/cm^2 in (d) and (g). Shading denotes SSTA over 0.1°C in (a), and significance levels as scaled by the color bar according to two-tailed Student t-test in (b)-(g). The time coordinate starts at January and ends at June in the following year. The monsoon season (JJAS) is indicated by the dashed lines. Refer to the text for the composite methods.

Figure 7. Similar to Fig. 6 but for cold years.

Figure 8. Composite SSTA (a), hcA (b), and tauA (c) in normal-dry years along the equator (2°S - 2°N average) in the interactive ensemble coupled model. (d)-(f) are similar to (a)-(c) except for normal-wet years. The contour interval is 0.2°C in (a), (b), (d), and (e), and 0.04 dyn/cm^2 in (c) and (f). Shading denotes SSTA over 0.1°C in (a) and (d), hcA over 0.1°C in (b) and (e), and tauA over 0.02 dyn/cm^2 in (c) and (f). The time coordinate is the same as in Fig. 6.

Figure 9. Similar to Fig. 6 but for the anomaly coupled model.

Figure 10. Similar to Fig. 7 but for the anomaly coupled model.

Figure 11. Similar to Fig. 8 but for the anomaly coupled model.

Figure 12. (a) Sliding correlation between the IMR and JJAS (red) and DJF (blue) Niño-3.4 SST anomalies; (b) Sliding mean anomaly of JJAS IMR ($\times 0.5$) (black), JJAS Niño-3.4 SST (red), and DJF Niño-3.4 SST (blue); (c) Sliding standard deviation of JJAS IMR ($\times 0.5$) (black), JJAS Niño-3.4 SST (red), and DJF Niño-3.4 SST (blue) anomalies. The sliding window width is 31 years. The calculation is based on the output from the anomaly coupled model. The unit is mm/day for IMR and $^{\circ}\text{C}$ for SST anomalies. The x-coordinate is the time (year).

Figure 13. Similar to Fig. 12 but for the interactive ensemble coupled model.