Low-frequency variability of the Indian Monsoon-ENSO relation and the Tropical Atlantic: The 'weakening' of the '80s and '90s.

by

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Abstract

The Indian Monsoon-El Nino Southern Oscillation (ENSO) relationship, according to which a drier than normal monsoon season precedes peak El Nino conditions, weakened significantly during the last two decades of the 20th century. In this work an ensemble of integrations of an Atmospherical General Circulation Model (AGCM) coupled to an ocean model in the Indian basin and forced with observed sea surface temperatures (SSTs) elsewhere is used to investigate the causes of such a weakening.

The observed interdecadal variability of the ENSO-Monsoon relationship during the period 1950-1999 is realistically simulated by the model and a dominant portion of the variability is associated to changes in the tropical Atlantic SSTs in boreal summer.

In correspondence to ENSO, the tropical Atlantic SSTs display negative anomalies south of the Equator in the last quarter of the 20th century and weakly positive anomalies in the previous period. Those anomalies in turn produce heating anomalies which excite a Rossby wave response in the Indian Ocean in both the model and in reanalysis data, impacting the time-mean monsoon circulation.

The proposed mechanism of remote response of the Indian rainfall to tropical Atlantic sea surface temperatures is further tested forcing the AGCM coupled to the ocean model in the Indian basin with climatological SSTs in the Atlantic Ocean and observed anomalies elsewhere. In this second ensemble the ENSO-Monsoon relation is characterized by a stable and strong anticorrelation through the whole second half of the XX century.
1 Introduction

In the late 1800s Sir Gilbert Walker, visiting India to investigate the causes of a series of severe droughts, discovered the Southern Oscillation (SO) and its relationship with the Indian monsoon rainfall (IMR). The fundamental physical mechanism behind the tendency for a drier than usual Indian monsoon preceding peak El Nino conditions (and vice versa for La Nina events) resides in the atmospheric teleconnection between the Pacific and Indian basin (Klein et al., 1999; Krishnamurthy and Kirtman, 2003). As El Nino develops in boreal spring, subsidence over South Asia increases due to an eastward shift of the Walker circulation over the centre of the Pacific; this suppresses convection over South Asia and results in a weaker summer monsoon (Walker, 1924; Rasmusson and Carpenter, 1983; Webster and Yang, 1992; Ju and Slingo, 1995 between others) while positive SST anomalies are still growing in the central and eastern Pacific. Since the late 1970s this relationship has weakened substantially (Krishna-Kumar et al., 1999; Torrence and Webster, 1999). As a notable example, the 1997/1998 El Nino event, despite its intensity, produced only marginal rainfall anomalies over India.

The drop in correlation between ENSO and the IMR has been tentatively attributed to a broad range of phenomena, ranging from changes in the atmospheric fields due to global warming (Krishna-Kumar et al., 1999), to natural low-frequency atmospheric variability (Krishnamurty and Goswami, 2000); from changes in the atmospheric circulation in the North Pacific (Kinter et al., 2002) or in the North Atlantic and consequently in the Eurasian snow cover (Chang et al., 2001), to variability of the Pacific Decadal Oscillation (Krishnan and Sugi, 2003); from changes in the atmospheric and oceanic teleconnection patterns between the Pacific and Indian basins associated to ENSO characteristics pre- and post- the climatic shift in 1976 (Annamalai and Liu, 2005; Annamalai et al., 2005), to stochastic noise (Gershunov et al., 2001), and, finally, to the co-occurrence of ENSO and the so-called Indian Ocean Zonal Mode (IOZM) (Saji et al., 1999; Webster et al., 1999). The existence of the IOZM as an independent coupled mode of variability of the Indian Ocean, its relation with ENSO and its impact
on the IMR have been subject of an intense debate in the last five years (Iizuka et al., 2000; Allan et al., 2001; Baquero-Bernal et al., 2002; Saji and Yamagata, 2003; Annamalai et al., 2003; Fisher et al., 2005) and consensus has not been reached so far.

Likely most of (if not all) the proposed mechanisms above are simultaneously at work and a "planetary-scale prospective" has to be taken when investigating both ENSO and monsoon dynamics (Chen, 2003). However, it is useful trying to identify what physical process(es) contributed most to the recent changes in the ENSO-monsoon relationship, as a clear picture has not emerged so far. A successful attempt would be beneficial to seasonal forecasting studies and to the analysis of future climate scenarios projections. Unfortunately it is very difficult to distinguish between the various theories using only observational data-sets: The signatures of all the proposed hypotheses are most likely mixed together and the statistical significance of the analysis is questionable due to the limited time coverage of reliable reanalysis products. An alternative way to address the problem is to construct a series of ad-hoc modelling experiments to test the various hypotheses. Atmospheric General Circulation Models (AGCMs), however, may not be adequate to assess the variability of the ENSO-monsoon relationship due to their inability to simulate the observed anticorrelation, as shown in two separate model intercomparison studies (Sperber and Palmer, 1996; Wang et al., 2004). Coupled air-sea feedbacks are important over the Indian Ocean, where SST anomalies are mostly originated from surface heat fluxes and cannot be simply assumed as a driver of the atmospheric variability (Lau and Nath, 2000, 2003; Wu and Kirtman, 2004, 2005; Wang et al., 2005; Bracco et al. 2006, referred to as B2006 in the following). Multi-decadal integrations with fully coupled GCMs are not adequate either, because the timing of particular events, such as the 'climate shift' in the mid 70's, cannot be simulated (e.g. Kinter 2002).

In this paper we investigate the physical mechanisms responsible for the observed changes in the Indian monsoon-ENSO relationship using two different modelling strategies. In a first set of experiments an AGCM is forced by observed SST anomalies outside the Indian Ocean region and it is coupled to an
ocean model in the Indian basin, insuring that coupled air-sea feedbacks are properly simulated. We will show that this configuration can reproduce the observed out-of-phase relation between ENSO and the monsoon rainfall variability and its decadal modulation during the second half of the 20th century. In those integrations the Indian Ocean SST variability forced remotely by ENSO is predominantly induced by heat fluxes and the observed ENSO-IOZM correlation is not reproduced, due to the absence of an oceanic bridge between the Indian and the Pacific, as already noticed in Bracco et al. (2005) and B2006.

A simple regression analysis performed on the model output and on reanalysis data identifies changes in the ENSO forced SST variability in the tropical Atlantic as one of the major contributor to the decadal variability of the Indian monsoon-ENSO relationship. To test this hypothesis we perform a second set of integrations where climatological SSTs are imposed in the Atlantic Ocean region and observed SST anomalies force the AGCM elsewhere except over the Indian Ocean, where again the ocean model is coupled to the AGCM. The IMR inter-decadal variability is not recovered in this configuration, further indicating that changes in the Atlantic play a fundamental role in modulating the rainfall response over India. The physical mechanism on which the teleconnection between the Atlantic and the Indian region is based upon is also highlighted.

A brief description of the model and of the ensemble performed is given in Section 2. Results are presented in Section 3. Discussion and conclusions follow in Section 4.

2 Model and Configurations

The model used in this study consists of the ICTP AGCM (Molteni, 2003), nicknamed SPEEDY for Simplified Parameterizations, primitive-Equation Dynamics, in its 8-layer configuration and T30 horizontal resolution coupled in the Indian basin to MICOM version 2.9 (Bleck et al., 1992) in a regional domain with 20 vertical levels and 1°x1° horizontal resolution.

Examples of applications of the AGCM component can be found in e.g. Bracco et al. 2004, Kucharski et al. 2006a and 2006b.
In the first set of experiments (ENS1) the ocean model domain extends from 35°S to 30°N and from the coast of Africa to about 140°E. A 3°-wide zone south of 32°S and east of 137°E is used to blend ocean model SSTs to observed values and subsurface quantities to Levitus monthly climatology. A portion of the Western Pacific is included to insure that the dynamics in the Indian basin is the least affected by the blending boundary conditions. Outside the Indian Ocean the AGCM is forced with the reanalysis SST data-set provided by the Hadley Center (HadISST, Rayner et al., 2003) from 1950 to 1999.

In the second ensemble (ENS2) the Atlantic Ocean SSTs are set to climatological values; otherwise the experiment is identical to ENS1.

Each set of integrations consist of a 10-member ensemble. Different members are created by randomly perturbing the initial conditions and by performing a 20-year spin-up integration.

3 Results

3.1 The IMR-ENSO relation and its interdecadal variability

As a measure of the interannual variability of the Indian Summer monsoon, we define an index that describes the IMR as the average June-to-September (JJAS) rain anomalies over land in the region 70°E to 95°E and 10°N to 30°N, covering most of the Indian Peninsula. In Fig. 1a the IMR index calculated for the ENS1 ensemble mean is compared with the one obtained using precipitation data of the Climate Research Unit (CRU in the following, Doherty et al. 1999). Overall, the observed interannual variability is very well reproduced by the model. The standard deviations are 0.61 mm/day for the CRU data and 0.53 mm/day for the ensemble mean in ENS1. The correlation coefficient (CC) between the two indices is 0.63, statistically significant at the 95% confidence level. Such a result is encouraging in light of the results presented by Krishna-Kumar et al. (2005). In fact, the authors have analyzed the predictability of the Indian summer monsoon

\[ \text{We verified the independence of our results on the inclusion of the Western Pacific portion with a three-member ensemble in which the ocean model domain is limited to 110E.} \]

\[ \text{We prefer to use the IMR index based on CRU data instead of the All Indian Rainfall Index (Parthasarathy et al., 1995) because similar diagnostics can be applied more easily to the model results and to the gridded data. The two indices are obviously highly correlated, with coefficient 0.85. We tested the independence of our results on the choice of the data-set.} \]
rainfall using large ensembles from ten different AGCMs for the period 1950-1999 and have shown that the ‘perfect model’ skill is bounded to have median correlation around 0.65, with 60% of the simulated variance originating from internal atmospheric dynamics.

We further test the ability of the model to reproduce the observed relation between the IMR and ENSO analyzing its evolution through the year. To this purpose we calculate the lead-lag correlations between the IMR and the NINO3 indices, with the last being defined as the average SST anomalies in the region 150°W to 90°W and 5°S to 5°N. Fig. 1b shows the lagged CC between the Indian monsoon rainfall over JJAS and the 4-months average NINO3 index shifted by n months relative to the IMR for CRU (open circles) and ENS1 ensemble mean (crosses). Thus in Fig. 1b the coefficient corresponding to -1 month lag refers to the correlation between JJAS IMR and the May-to-August NINO3 index, and so on. The seasonal modulation of the ENSO-Indian Monsoon relationship is extremely well captured by the ensemble mean. Both in the model and in the observations the maximum negative correlation (around -0.6) is reached for contemporaneous indices and for slightly positive lags (i.e for the IMR leading the NINO3 index by a season or so). Given the particular set-up of our integrations, with observed SSTs forcing the AGCM in the Pacific, this result supports the idea that ENSO remotely influences the dynamics of the Indian Summer Monsoon in its development stage. This happens through changes in the heating patterns over the West and Central Pacific in the spring and summer preceding the winter peak (see, e.g., Ju and Slingo, 1995 for a detailed description of such a mechanism).

To further elucidate the relationship between ENSO and the IMR variabilities in observations and in the model output, Fig. 2 shows the regression pattern of CRU (Fig. 2a) and ENS1 ensemble mean (Fig. 2b) onto the NINO3 index in JJAS for the 1950 to 1999 period. All regression maps correspond to one standard deviation of the regression index. As can be clearly identified from Figs. 2a and 2b, ENSO leads to a substantial reduction of IMR in observations and in ENS1 (-0.35 mm/day for CRU, and -0.42 mm/day for ENS1). The correspondence between model and observations is overall very good, with the exception
of the Bay of Bengal region, where a larger than observed increase in rainfall is simulated in response to El Nino events east of the Indian peninsula, over Thailand, Laos, Cambodia and Vietnam (and therefore outside the region of interest in this study). As found in B2006, where the model climatology and variability are analysed in detail, and in agreement with results by Wu and Kirtman (2004), a fully coupled model over the West Pacific is needed to properly simulate the latitudinal extension of convective precipitation around the South China basin.

Assuming that the model is realistically simulating the Indian monsoon precipitation and the relationship with ENSO, we perform a simple correlation analysis between the IMR and the NINO3 indices in JJAS over the period 1950 to 1999 and over the 25-year intervals 1950 to 1974 and 1975 to 1999 separately. We have chosen to consider two intervals of equal length instead of dividing our data according to the so-called 'climate shift' of 1976/1977, as such a shift cannot be considered a priori responsible for the weakening of the ENSO-monsoon teleconnection. Analogous results are obtained if 1976/1977 is used instead.

Table 1 summarizes our findings. Overall the observed anticorrelation of the IMR with the NINO3 index is well reproduced in the ensemble mean. The correlation coefficients are $CC = -0.59$ for CRU data and $CC = -0.63$ for the ENS1 ensemble average. More importantly, the weakening in the Indian monsoon-ENSO relationship observed in the last quarter of the 20th century is well captured by the model. Indeed, when the 1975-1999 period is compared to previous 25 years the anticorrelation reduces of 0.24 for CRU data and of 0.28 in ENS1. The change in the modeled ensemble mean correlation is statistically significant at the 90% confidence level. If individual ensemble members are used, the average drop in anticorrelation is from -0.49 (1950-1974 period) to -0.33 (1975-1999 period), and the relative change remains about the same as for the ensemble mean. According to a t-test the change between the two periods is statistically significant at the 95% confidence level. 4

An even stronger reduction (0.34) is found if the All-Indian-Rainfall dataset is used instead of CRU.

To assess the statistical significance of the different means of the two ensembles of CC, we apply the Fischer's Z-transformation to the ensemble of CC for each period, then we perform a t-test to assess that the distributions have different means.
tion in anticorrelation between 0.1 and 0.5. One outlier displays an increase in anticorrelation of -0.24 during the last quarter of the XX century. As quantified by Krishna-Kumar et al. (2005), internal atmospheric dynamics plays an important role in this region and variability between the individual members has to be expected when considering changes in correlation over 25-year periods.

To further test the independence of our results on the specific periods considered, we assess the changes in correlation between the IMR and the NINO3 indices for the twenty-year intervals 1980/1999 and 1950/1969. The results are summarised at the bottom of Table 1. Overall, the results are analogous to what previously described, although in both CRU and for ENS1 the CC changes are somewhat weaker.

The observed and simulated changes in the precipitation patterns associated with ENSO are shown in Fig. 3. Regression maps versus the NINO3 index are calculated for each of the 25-year intervals and the difference is plotted for the CRU data (Fig. 3a) and the ensemble mean of ENS1 (Fig. 3b), with superimposed the modeled 925 hPa wind anomalies. Over India the observed and modeled patterns are very similar, although some details in the small scale structures are absent in the model runs, where topographic features are not well resolved due to the coarse model horizontal resolution. Over India the precipitation associated with positive ENSO events increases significantly in the latter period, both in the observations and ENS1. The average increase of the IMR in the NINO3 regressions is 0.09 mm/day in CRU data and 0.13 mm/day in ENS1 per standard deviation of the NINO3 index. The difference in the surface wind regression between the two periods indicates that an anomalous cyclonic flow over India has strengthened the monsoon circulation in concomitance with positive ENSO events.

Before examining in more details the physical mechanism that is responsible for the weakening of the ENSO-IMR anticorrelation, it is worth commenting on other regions where the modelled patterns are different from the observed ones. Over the east countries of South Asia and China the model is not reproducing the observed rainfall signal. As already mentioned, a coupled model extending further into the Pacific is indispensable to proper simulate precipitation patterns.
over South China Sea. Furthermore, as discussed in Bracco et al. (2005) and B2006, the absence of an oceanic teleconnection between the Indian and the Pacific basin precludes the development of the IOZM-ENSO relation as observed. The model integrations all display a mode of variability internal to the Indian Ocean with a dynamics comparable to the one of the observed IOZM (see e.g. Murtugudde et al., 2000; Rao et al., 2002; Gualdi et al., 2003), but independent of ENSO. The correlation between the ensemble mean IOZM, defined as in Saji et al. (1999), and the NINO3 indices is indeed 0.01 in JJAS and is not significant in all the ten members. It has been shown in Bracco et al. (2005) that whenever the OGCM domain includes the Tropical Pacific the observed relation is properly simulated and that the 'oceanic bridge' between the two basins is an essential component of the IOZM-ENSO connection, as also found in Annamalai et al. (2005). The simulation of the observed IOZM-ENSO relationship is fundamental to the correct representation of the rainfall variability associated with ENSO over Indonesia and China (B2006), but from the analysis presented here it is not necessary to reproduce a large extent of the variability over India.

The independency of the internal and forced SST variability in the Indian basin in ENS1 integrations allows to conclude that the modeled Indian monsoon rainfall and wind anomalies are remotely forced and associated with SST-anomalies outside the Indian Ocean.

As a first step to identify the SST anomalies responsible for forcing the AGCM response, we calculate the regression coefficients of the observed SSTs versus the NINO3 index using the HadISST data-set and including the Indian Ocean basin. Fig. 4a and Fig. 4b show the SST regression patterns in the periods 1950-1974 and 1975-1999 in JJAS, respectively. Fig. 4c shows the difference of the SST regressions between the periods 1975-1999 and 1950-1974.

For the regression difference (Fig. 4c) a global patterns appears, but the dominant signal in the tropical band is found in the Atlantic region, where the cold anomalies reach values of 0.4 K between 20°S and 0°S. The reason for this difference is mainly the cooling in the tropical Atlantic region in the later 1975-1999 period as can be seen in Fig. 4b, with a small contribution from a warming
of the tropical Atlantic in the first period 1950-1974 (Fig. 4a).

No significant changes are found in the immediate proximity of the Indian peninsula, further supporting the hypothesis that the internal dynamics of the basin is not primarily responsible for the observed weakening of the ENSO-IMR relation.

### 3.2 The role of Tropical Atlantic SSTs

The ENSO teleconnection to the tropical Atlantic has been subject of several studies and represents a substantial part of the variability in this region (Zebiak, 1993; Curtis and Hastenrath, 1995; Latif and Barnett, 1995; Enfield and Mayer, 1997; Giannini et al. 2001, Huang et al. 2002; Huang et al. 2004, between others). Chiang et al. (2000) and more recently Münnich and Neelin (2005) analyzed interdecadal changes in the patterns of such teleconnection in late spring and summer and the implications for the South America rainfall variability. Both works noticed that during the '80s and '90s the link between the Pacific and Atlantic ITCZs has been strong, while it was non-existent in the '50s and '60s, possibly resulting from the significant increase in the ENSO variance during the last quarter of the 20th century. Chang et al. (2006), however, supported by the numerical results of Huang (2004), emphasized that fluctuations in the equatorial Atlantic depend not only on the conditions of the tropical Pacific but also on the state of upper ocean in the Atlantic itself.

The cooling of the tropical Atlantic is responsible for a northward shift of the ITCZ over Africa (Latif and Barnett, 1995; Giannini et al., 2001). Results in Fig. 3b suggest that it may also influence the monsoon circulation over India. To further investigate the role of the Atlantic SSTs, in the ENS2 ensemble SST anomalies in the Atlantic Ocean region are set to zero and climatological values are used to force the AGCM in a longitude band from about 70 W to 30 E. Otherwise, ENS2 is identical to ENS1. The resulting CC between the observed and ensemble mean IMR drops from 0.63 to 0.54 for the 1950 to 1999 period. Considering individual ensemble members the reduction in the average CC is from 0.43 in ENS1 to 0.35 in ENS2 and this change is statistically significant at
the 95 % level. This suggests that important information for the IMR variability resides in the Atlantic SST anomalies.

Furthermore, the correlation coefficient between the ENS2 ensemble mean IMR and the NINO3 index is strong (-0.73) and does not vary significantly during the interval considered (0.013 as compared to 0.28 in case of ENS1; see Table 1). For the mean of individual ensemble members, the change of CC during the last two quarters of the XX century is even slightly negative (CC changes in individual ENS2 members range from -0.32 to 0.26) and so is the ensemble mean precipitation change over the IMR region (-0.04 mm/day). On the other hand, the differences between ENS2 and ENS1 are statistically significant at the 90 % confidence level.

Fig. 3c shows the differences in the regressions for rain and 925 hPa wind of ENS2 in analogy to those shown in Fig. 3b for ENS1. Clearly, the patterns in the two ensembles differ significantly. In the absence of the Atlantic Ocean SST anomalies the regression pattern changes sign over India, providing still slightly positive values over west India, but stronger negative values over east India and Bangladesh. This extends over the Indian Ocean with a strong dipolar structure in the precipitation field and a corresponding wind response. A somewhat similar feature occupies the eastern Indian Ocean in ENS1 as well (Fig. 3b), but is much weaker. From this simple analysis appears that Atlantic SST anomalies acts to counteract the ENSO-forced response in the Indian Ocean during the last 25 years of the XX century.

To isolate the effect of the Atlantic SST anomalies, in Fig. 3d we show the difference in the patterns of Fig. 3b and Fig. 3c (ENS1-ENS2). Not surprisingly, over Africa the pattern of Fig. 3d is very similar to that of Fig. 3b. However, the impact of the Atlantic is not limited to Africa and penetrates into the Indian Ocean, where it induces a strong westerly flow, and over the Indian Peninsula, where a cyclonic flow enhances the monsoonal circulation, reducing the anticorrelation between ENSO and IMR indices. Indeed, the average precipitation change between the two ensembles in the IMR region is 0.17.

To further assess the hypothesis that Atlantic SSTs influence the Indian Mon-
soon variability, we calculate the regression coefficients of rainfall onto an index defined as the *negative* of the area average SST anomalies in the south tropical Atlantic in the region comprised between 30°W to 20°E and 20°S to 0°S (referred to as the tropical Atlantic index in the following). In the CRU data (Fig. 5a) and ENS1 (Fig. 5b), the aforementioned northward shift of the ITCZ can be identified (for ENS1 only contours that are statistically significant at a 95% confidence level according to a 2-tailed t-test are plotted). In ENS1 the shift of the ITCZ extends into the Indian Ocean and further into the Indian peninsula and causes a cyclonic surface wind anomaly similar to that of Fig. 3b. The average IMR anomaly associated to the tropical Atlantic index is 0.15 mm/day for CRU data and 0.18 mm/day for the ENS1 ensemble mean.

Additionally, the regression of SSTs onto this index (Fig. 5d) displays a pattern similar to the one in Fig. 4c. Overall, the regression pattern of Fig. 5b is very similar to Fig. 3d, where the Atlantic effect is isolated by considering the difference of ENS1 and ENS2.

Finally, the role of the Atlantic SST variability can be isolated by calculating the regression of the ensemble mean difference (ENS1-ENS2) of rain and 925 hPa wind onto the tropical Atlantic index (Fig. 5c; again only differences that are statistically significant at the 95% confidence level are displayed). The resulting pattern is very similar to that of ENS1, confirming that is induced by the Atlantic SST variability.  

The correlation of the tropical Atlantic index with the NINO3 index is small and not significant if the interval 1950-1999 is considered (CC = 0.08). However, it varies from CC = -0.22 in the period 1950 to 1974 to CC = 0.41 in the period 1975 to 1999. Such a drastic change of 0.63 is statistically significant at the 95% confidence level.

Fig. 5a-c could suggest that tropical Atlantic SST anomalies influence substantially the IMR on an interannual basis. However, the amplitudes of the regressions seen in Fig. 5 are small with respect to the absolute NINO3 regres-

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5This may seem to be a trivial result, but it is not because SST anomalies outside the Atlantic region that covary with the tropical Atlantic index may influence the IMR. Fig. 5c shows that this is not the case.
sions (Fig. 2). Indeed, the correlation coefficient of the tropical Atlantic index with the IMR are only 0.24 and 0.25 for CRU data and the ENS1 ensemble mean, respectively. Therefore, the overall influence of tropical Atlantic SST anomalies on Indian Monsoon is weaker than that of ENSO, but the interdecadal variability of the SSTs over the tropical Atlantic plays a key role in modulating the strength of the ENSO-IMR relation.

3.3 Physical mechanism for Tropical Atlantic-IMR teleconnection

The physical mechanism by which the tropical Atlantic influences the Indian Monsoon Rainfall can be highlighted by considering the distribution of the regression coefficients of the ensemble average 200 hPa eddy streamfunction ($\psi$) of ENS1-ENS2 (therefore isolating the effect of the Atlantic Ocean SSTs) onto the tropical Atlantic index in JJAS (Fig. 6a). In the equatorial region the propagation of Rossby waves to the north is substantially suppressed and the Rossby wave response pattern is trapped in the tropics. The structure of the response in Fig. 6a is in the form of a quadruple pattern centered over equatorial Africa and is very similar to the one found by Ting and Yu (1998) in a simple baroclinic model for the June-to-August season (see, e.g., their Fig. 13. In their study the most efficient region for exciting this kind of response was found to be the central Pacific). Over the northern part of the Indian Ocean and the Arabian Sea an anticyclone dominates the regression pattern at 200 hPa. This feature has a baroclinic vertical structure with a cyclone at lower levels slightly west of India, evident in the 925 hPa wind regression of Fig. 5c. Such a cyclone enhances the monsoon circulation over the Indian peninsula. This is consistent with the fact that the $\psi$ regression in Fig. 6a projects positively onto the ENS1 mean streamfunction field shown in Fig. 6b for the period 1950 to 1999, which is similar to the observed time-mean eddy $\psi$ (e.g. Chen, 2003).

Figs. 6c and 6d augment this interpretation by showing that the 200 hPa velocity potential response (Fig. 6c, where a maximum suggests upper-level convergence, and a minimum indicates divergence) also projects onto the time-mean velocity potential of ENS1 (Fig. 6d) The cooling of the tropical Atlantic Ocean
causes upper-level convergence that is compensated by upper-level divergence in the surroundings. Note that the Indian Monsoon region lies in the strongest gradient of the time-mean velocity potential (Fig. 6d), as suggested by Chen (2003).

4 Discussion and Conclusions

In this work we analyzed the interannual variability of the Indian Monsoon Rainfall, focusing on its relation with the El Nino Southern Oscillation both in the observations and in an atmospheric general circulation model regionally coupled to an ocean model over the Indian Ocean basin.

We have shown that the observed interannual IMR variability can be realistically reproduced in our CGCM, within the limitations of the model horizontal resolution. The model is able to simulate both the observed Indian monsoon-ENSO relation and the variability observed over the second half of the 20th century. The ad hoc modelling strategy allows us to investigate the forced component of the IMR variability. The only AGCM forcing included in the integrations are the SST anomalies outside the Indian Ocean region. Within the Indian basin the coupled ocean model reproduces an internal mode of variability with the characteristics of the Indian Ocean Zonal Mode, but independent of ENSO. In such a model, the SST variability forced by ENSO within the Indian basin is small and does not display significant interdecadal changes over the second half of the 20th century (B2006). SST anomalies outside the Indian Ocean domain must therefore be responsible for the interdecadal variability of the ENSO-Indian monsoon relationship in our simulations.

In those experiments, the most important contributor to the interdecadal variability of the ENSO-Indian Monsoon relationship appears to be the tropical Atlantic Ocean. The analysis of the HadISST data-set reveals that the ENSO-SST teleconnection pattern has changed over the last quarter of the 20th century compared to the previous 25 years, with maximum differences occurring in the tropical Atlantic region, as also noted in the recent work by Münnich and Neelin (2005). Those changes are responsible for weakening the ENSO-monsoon rela-
tionship in our simulations.

Negative SST-anomalies in the south equatorial tropical Atlantic are indeed co-occurring with ENSO in boreal summer in the last quarter of the 20th century relative to the previous 25 years, inducing heating anomalies and a Rossby wave response in the tropics. Over the Indian Ocean the Rossby wave projects onto the time-mean pattern of the circulation, thus enhancing the time-mean Indian summer monsoon. This interpretation is supported by a second ensemble, equivalent to the first one, but with climatological SST in the Atlantic Ocean. In this second ensemble the ENSO-IMR anticorrelation is stable and strong over the whole period considered and does not display any significant interdecadal variability.

Two caveats should be attached to the analysis presented. Firstly, the spread in the change of correlation between the NINO3 and the IMR indices among individual ensemble members is quite large (ranging from 0.1 to 0.5 for 9 members, with one outlier of -0.24), due to the important role that internal atmospheric variability plays in this region. Secondly, the pattern of rainfall changes in the ensemble average (and for the individual members) agrees with the CRU data over the central and western Indian peninsula, but does not reflect the observed changes over Bangladesh and the countries surrounding the South China Sea. In the regionally coupled simulations (ENS1) the observed relation between ENSO and the Indian Ocean Zonal Mode is not simulated, because of the absence of sub-surface connections between the Indian and Pacific Oceans (Bracco et al., 2005). The IOZM influences the monsoon rainfall over South and East Asia and enhances the meridional Indian monsoon circulation (Slingo and Annamalai, 2000; Annamalai and Liu, 2005, among others). In ENS1 we cannot expect to reproduce all the features of the observed patterns because we are missing one of the dynamical factors that can modify the impacts of ENSO on the monsoon precipitation.

In summary, we have shown that the interdecadal variability of the ENSO-Indian monsoon relationship is, to a significant extent, modulated via the tropical Atlantic route. In this work we cannot assess the causes for the observed changes in the ENSO-SST teleconnection in the Atlantic region. They may result from
natural, internal variability of the chaotic climate system, or may be due to changes in the ENSO properties following the climate shift that occurred in the mid 70’s. The analysis of the interdecadal variability in the tropical Atlantic in a coupled model will be subject of future study.

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Figure 1
a) Time series of observed (CRU; open circles) versus ENS1 (crosses) ensemble mean Indian JJAS rain anomalies (averaged over land points of the region 70°E to 95°E and 10°N to 30°N). b) Plot of the lagged correlation coefficient between the JJAS IMR and the (lagged) 4-months average NINO3 index for CRU (open circles) and ENS1 (crosses). Negative lags (in months) mean that NINO3 is leading IMR, positive lags mean IMR is leading the NINO3 index. The units are for mm/day for a).

Figure 2
Regressions onto the NINO3 index (defined as average SST anomalies in the region 150°W to 90°W and 5°S to 5°N) for the period 1950 to 1999. a) CRU rain, b) ENS1 ensemble mean rain and 925 hPa wind. Units are mm/day for rain in a) and b) and m/s for wind in b).

Figure 3
1975/1999-1950/1974 differences of regressions of rain and 925 hPa wind onto the NINO3 index. a) CRU rain, b) ENS1 ensemble mean rain and 925 hPa wind, c) ENS2 ensemble mean rain and 925 hPa wind, d) ENS1-ENS2 ensemble mean rain and 925 hPa wind. Units are mm/day for rain in a) and b), d) and m/s for wind in b), d), e).

Figure 4
Regressions and regression differences of SSTs onto the NINO3 index. a) 1950/1974, b) 1975/1999, c) 1975/1999-1950/1974 difference of regressions. Units are K.

Figure 5
Regressions of rain and SST onto a tropical Atlantic Index defined as negative average SST anomalies in the region 30°W to 20°E and 20°S to 0°S. a) CRU rain, b) ENS1 ensemble mean rain and 925 hPa wind, c) ENS1-ENS2 ensemble mean rain and 925 hPa wind and d) SSTs. Units are mm/day for rain in a), b) and c) and m/s for wind in b) and c), units for d) are K.
Figure 6

a) Regression of the 200 hPa eddy streamfunction of the ENS1-ENS2 ensemble mean onto a tropical Atlantic index defined as negative average SST anomalies in the region 30°W to 20°E and 20°S to 0°S. b) Time-mean 200 hPa eddy streamfunction of ENS2. c) Regression of the 200 hPa velocity potential of the ENS1-ENS2 ensemble mean onto a tropical Atlantic index, d) Time-mean 200 hPa velocity potential. Units are 10^6 m^2/s.

Table 1

Correlation Coefficients (CC) between IMR and the NINO3 index (defined as average SST anomalies in the region 210°E to 270°E and 5°S to 5°N) for CRU (top), ensemble mean of ENS1 (2nd row) and mean of individual ensemble members of ENS1 (3rd row), ensemble mean of ENS2 (4th row) and mean of individual members of ENS2 (5th row) for the periods 1950 to 1999, 1950 to 1974, and 1975 to 1999. The last column shows the CC difference of the later minus the earlier period for CRU and the respective experiments. Last 2 rows: same as above, but for sub-periods 1950 to 1969, and 1980 to 1999 for CRU and ENS1.
Figure 1: a) Time series of observed (CRU; open circles) versus ENS1 (crosses) ensemble mean Indian JJAS rain anomalies (averaged over land points of the region 70°E to 95°E and 10°N to 30°N). b) Plot of the lagged correlation coefficient between the JJAS IMR and the (lagged) 4-months average NINO3 index for CRU (open circles) and ENS1 (crosses). Negative lags (in months) mean that NINO3 is leading IMR, positive lags mean IMR is leading the NINO3 index. The units are for mm/day for a).
Figure 2: Regressions onto the NINO3 index (defined as average SST anomalies in the region 150°W to 90°W and 5°S to 5°N) for the period 1950 to 1999. a) CRU rain, b) ENS1 ensemble mean rain and 925 hPa wind. Units are mm/day for rain in a) and b) and m/s for wind in b).
Figure 3: 1975/1999-1950/1974 differences of regressions of rain and 925 hPa wind onto the NINO3 index. a) CRU rain, b) ENS1 ensemble mean rain and 925 hPa wind, c) ENS2 ensemble mean rain and 925 hPa wind, d) ENS1-ENS2 ensemble mean rain and 925 hPa wind. Units are mm/day for rain in a) and b), d) and m/s for wind in b), d), e).
Figure 4: Regressions and regression differences of SSTs onto the NINO3 index. a) 1950/1974, b) 1975/1999, c) 1975/1999-1950/1974 difference of regressions. Units are K.
Figure 5: Regressions of rain and SST onto a tropical Atlantic Index defined as negative average SST anomalies in the region 30°W to 20°E and 20°S to 0°S. a) CRU rain, b) ENS1 ensemble mean rain and 925 hPa wind, c) ENS1-ENS2 ensemble mean rain and 925 hPa wind and d) SSTs. Units are mm/day for rain in a), b) and c) and m/s for wind in b) and c), units for d) are K.
Figure 6: a) Regression of the 200 hPa eddy streamfunction of the ENS1-ENS2 ensemble mean onto a tropical Atlantic index defined as negative average SST anomalies in the region 30°W to 20°E and 20°S to 0°S. b) Time-mean 200 hPa eddy streamfunction of ENS2. c) Regression of the 200 hPa velocity potential of the ENS1-ENS2 ensemble mean onto a tropical Atlantic index, d) Time-mean 200 hPa velocity potential. Units are 10^6 m^2/s.
**Table 1:** Correlation Coefficients (CC) between IMR and the NINO3 index (defined as average SST anomalies in the region 210°E to 270°E and 5°S to 5°N) for CRU (top), ensemble mean of ENS1 (2nd row) and mean of individual ensemble members of ENS1 (3rd row), ensemble mean of ENS2 (4th row) and mean of individual members of ENS2 (5th row) for the periods 1950 to 1999, 1950 to 1974, and 1975 to 1999. The last column shows the CC difference of the later minus the earlier period for CRU and the respective experiments. Last 2 rows: same as above, but for sub-periods 1950 to 1969, and 1980 to 1999 for CRU and ENS1.

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<tr>
<td>CC(CRU,NINO3)</td>
<td>-0.59</td>
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<td>-0.47</td>
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<td>CC(CRU,NINO3)</td>
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