A Gill–Matsuno-type mechanism explains the tropical Atlantic influence on African and Indian monsoon rainfall

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ABSTRACT: Recent studies using coupled atmosphere–ocean models have shown that the tropical Atlantic has a significant impact on the Indian monsoon. In this article, the observational basis for this teleconnection is examined and the physical mechanism responsible for bridging sea-surface temperatures (SSTs) in the Atlantic and precipitation over India is investigated with idealized atmospheric general circulation model (AGCM) experiments in which constant SST anomalies are prescribed and ‘switched on’ in the tropical Atlantic region. A simple Gill–Matsuno-type quadrupole response is proposed to explain the teleconnection between the tropical Atlantic and the Indian basin, with an enforcement of the eastward response likely due to nonlinear interactions with the mean monsoon circulation. The simplicity of this mechanism suggests the reproducibility of this result with a broad range of AGCMs. Copyright © 2009 Royal Meteorological Society

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1. Introduction

Sea-surface temperature (SST) anomalies in the tropical Atlantic affect interannual rainfall variability over Africa. As shown by, for example, Trzaska et al. (1996), Janicot et al. (1998), Fontaine et al. (1998), Vizy and Cook (2001), Giannini et al. (2003) and Reason and Rouault (2006), there is a direct correlation between ocean temperatures and precipitation along the Gulf of Guinea coast. More recently, Kucharski et al. (2007, 2008a) (referred to as K2007/8 in the following) have found that the influence of tropical Atlantic SSTs extends to precipitation over India on interannual time-scales, contributing to the decadal modulation of the El Niño Southern Oscillation (ENSO)–Indian monsoon relationship, according to which a weaker than normal monsoon precedes peak El Niño conditions and vice versa during summers preceding La Niña events. K2007/8 have shown that the atmospheric general circulation model (AGCM) in this study, coupled to an ocean model in the Indian basin and forced by imposed SSTs elsewhere, can simulate a dominant portion of the Indian monsoon interannual variability, with a 0.62 correlation between modelled rainfall and Climate Research Unit (CRU; New et al., 1999) data over India for the period 1950–1999 only if observed tropical Atlantic SST anomalies are included.

Before 1975, the equatorial Atlantic SSTs were slightly warmer than climatological values, corresponding to positive ENSO events, and contributed to reinforcing the El Niño–monsoon anticorrelation. After 1975, on the other hand, a substantial cooling in the tropical Atlantic Ocean intensified the monsoonal circulation and thus weakened the ENSO–Indian monsoon relationship. If climatological SSTs are imposed over the tropical Atlantic, the simulated El Niño-Indian monsoon anticorrelation remains constant over the whole 50 year period (see Table 1 in Kucharski et al., 2007) and the correlation between observed and modelled rainfall over India drops to 0.42.

Here we investigate the physical mechanism by which south tropical Atlantic SSTs impact precipitation in adjacent areas with a focus on the implication for seasonal predictability in remote regions. While the atmospheric response to a heating anomaly in the tropical Pacific has been analyzed, e.g. in Jin and Hoskins (1995) (referred to as JH95 in the following), Lintner and Chiang (2007) and Seager et al. (2008) by performing targeted integrations, an analogous study in the Atlantic is currently lacking, with the exception of a multimodel analysis limited to the effect of North Atlantic SSTs on the surrounding regions (Rodwell et al., 2004).

We find that the teleconnection over Africa and the Indian Ocean can be understood, to a first approximation, as a simple Gill–Matsuno-type quadrupole response of the atmosphere to a heating anomaly in the equatorial Atlantic region (Matsuno, 1966; Gill, 1980). Equatorial Kelvin waves transport the signal from the heating region to the east, whereas equatorial Rossby waves...
are responsible for the westward propagation. Further supporting K2007/8, we also show that the tropical Atlantic influence on the Indian monsoon is robust and physically sound and, as such, potentially reproducible by AGCMs.

In Section 2 we present observational evidence for the tropical Atlantic–Indian monsoon teleconnection. The AGCM used in this study and the numerical experiments performed are described in Section 3. The analysis of the results follows in Section 4. Section 5 contains a summary and conclusions.

2. Observational evidence

In K2007/8, observational support for the tropical Atlantic–Indian monsoon teleconnection has been presented using CRU precipitation data and the All-Indian Rainfall Index (AIR; Parthasarathy et al., 1995). Furthermore, in an independent analysis using a central Indian rainfall index, Yadav (2008; see Figure 5(b) therein) has shown that the south tropical Atlantic is significantly correlated with the Indian monsoon in non-ENSO years. Here we extend the analysis to Climate Prediction Center Merged Analysis of Precipitation (CMAP; Xie and Arkin, 1997) and Global Precipitation Climatology Project (GPCP; Adler et al., 2003) precipitation datasets and to the National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) wind re-analysis (Kalnay et al., 2006) focusing on the July-to-September season (JAS).

Following Kucharski et al. (2008a), the Atlantic SST anomalies significantly correlated with the Indian monsoon rainfall are located between 30°W–10°E and 20°S–0°S, and are usually referred to as the southern tropical Atlantic (STA) pattern (Huang et al., 2004). The existence of a teleconnection between STA anomalies and ENSO has been proposed in several studies (Zebiak, 1993; Latif and Grötzner, 2004). Recent works, however, pointed out that ENSO influences the tropical Atlantic mainly north of the equator (Chang et al., 2006), while STA SST anomalies in the southern hemisphere result from air–sea interactions along the coast of Africa with intensification by local physical processes, including slow Rossby wave propagation, Ekman pumping and Bjerknes feedbacks (Hu and Huang, 2007). With a simple correlation analysis, the occurrence of a stable link between ENSO and STA anomalies cannot be established. The correlation coefficient (CC) between the STA index (area-averaged SST anomalies in the STA domain) and the NINO3.4 index (averaged SST anomalies in the region 190°E–240°E and 5°S–5°N) is non-significant over the period 1950–2002 (CC = −0.05), but its magnitude varies from 0.18 over the period 1950–1979 to −0.42 during the last twenty years of the record (1980–2002), which is 90% statistically significant (see figure 4 in Kucharski et al., 2007). The standard deviation of the STA index is about 0.3 K for both periods, compared with about 0.7 K for the NINO3.4 index. Figure 1 shows the regression of SSTs from the HadISST dataset (Rayner et al., 2003) on to the STA index for the period 1950–2002 in the tropical Atlantic region, multiplied by 2 (SSTs outside the tropical Atlantic are set to zero in this figure, see Section 3).

Given the strong influence exerted by the ENSO on the surrounding areas, and in particular over the Indian basin (e.g. Rasmussen and Carpenter, 1983; Webster and Yang, 1992; Ju and Slingo, 1995), a first step to isolate the ENSO-independent contribution of the tropical Atlantic on the Indian monsoon variability consists of assuming that the ENSO has a linear impact on STA SSTs. Under this assumption, we can remove the component linearly related to the ENSO from the STA index by calculating the residual time series:

$$STA_{RES}(t) = STA(t) - STA_{ENSO}(t),$$

where

$$STA_{ENSO}(t) = b \cdot NINO34(t).$$

$b$ is calculated with a least-squares linear regression between the STA and Nino3.4 indexes. The standard deviation of $STA_{RES}$ is only marginally reduced with respect to STA for the period 1980–2002 (0.28 instead of 0.3 K), and the same (0.3 K) over the 1950–2002 interval. The SST regression on to the $STA_{RES}$ index is almost identical to one on to the full STA index, supporting the conclusions of Chang et al. (2006). Focusing on the impact of tropical Atlantic SSTs on precipitation, Figure 2(a) shows the regression of CRU rainfall on to the $STA_{RES}$ index for the period 1950–2002, while Figure 2(b) and (c) display the regression of CMAP and GPCP rainfall for the period 1980–2002, respectively (CMAP and GPCP are available only after 1979). The regressions are per standard deviation of the $STA_{RES}$ index. Regions with anomalies that are 90% statistically significant according to a t-test are shaded. The rainfall responses in Figure 2 share many common features: increased rainfall over the Gulf of Guinea coast; reduced rainfall to the north in the Sahelian region and further to the east, particularly in the CRU data; reduced rainfall over west India. For rainfall indexes defined as area averages in the Indian region (70°E–85°E and 10°N–30°N), the standard deviations for the CRU, CMAP and GPCP indexes are 1.0, 1.0 and 0.9 mm day$^{-1}$, whereas the averages of the regressions in Figure 2 over the same regions are −0.38, −0.38 and −0.29 mm day$^{-1}$. Therefore, the contribution of the $STA_{RES}$ teleconnection to rainfall in the Indian region is in the order of 0.4
GILL RESPONSE TO TROPICAL ATLANTIC SST FORCING

Figure 2. Regressions of (a) CRU (1950–2002), (b) CMAP (1980–2002) and (c) GPCP (1980–2002) rainfall on to the JAS STA\textsubscript{RES} index, as defined in the text. Shading indicates regions of 90\% statistically significant anomalies (positive: orange/light grey, negative: blue/dark grey). The units are per standard deviation of the STA\textsubscript{RES} index. Contour intervals are 0.3 mm day\(^{-1}\). This figure is available in colour online at www.interscience.wiley.com/journal/qj

The existence of a teleconnection between the STA region and the Indian monsoon is further demonstrated by an analogous analysis performed on the dynamical Indian monsoon index (IMI). The IMI is defined as the meridional 850 hPa zonal wind difference averaged over the domains 40\(^\circ\)E–80\(^\circ\)E, 5\(^\circ\)N–15\(^\circ\)N minus 60\(^\circ\)E–90\(^\circ\)E, 20\(^\circ\)N–30\(^\circ\)N, and provides a measure of the Indian monsoon strength (Wang et al., 2001). Figure 3(a) shows the correlation coefficients (CCs) of the IMI derived from the NCEP/NCAR reanalysis with the SSTs from the HadISST dataset for the period 1950–2002. Shading indicates a 95\% statistically significant correlation according to a \(t\)-test. Unsurprisingly, we find a strong anticorrelation between the Indian monsoon strength and the eastern Pacific SST anomalies, the interannual variability of which is dominated by the ENSO signal. To interpret the correlation seen in other regions, we apply the same technique described above and we define the residual IMI, \(\text{IMI}_{\text{RES}}\), as the IMI component that is not linearly related to ENSO according to

\[\text{IMI}_{\text{RES}}(t) = \text{IMI}(t) - \text{IMI}_{\text{ENSO}}(t),\]  

(3)

where

\[\text{IMI}_{\text{ENSO}}(t) = b \times \text{NINO34}(t).\]  

(4)

Again, \(b\) is calculated with a least-squares linear regression. Figure 3(b) shows the CCs of \(\text{IMI}_{\text{RES}}\) with SSTs. Clearly, SST anomalies in the Gulf of Guinea display the largest correlations, while all other regions are characterized by barely statistically significant CCs. Overall, these results further support the findings of K2007/8.

Even though we removed the component of the STA index linearly related to the ENSO in the above precipitation regression maps, we cannot rule out the suggestion that SST anomalies in other regions influence the outcome. The only way to quantify the STA influence on the Indian monsoon is to perform targeted, idealized experiments that isolate the contribution of the STA anomalies.

3. Model and experimental design

The model adopted in this study is the International Centre for Theoretical Physics (ICTP) AGCM (Molteni, 2003). It is based on a hydrostatic spectral dynamical core (Held and Suarez, 1994), and uses the vorticity-divergence form described by Bourke (1974). The parametrized processes include short- and long-wave radiation, large-scale condensation, convection, surface fluxes of momentum, heat and moisture, and vertical diffusion. Convection is represented by a mass-flux scheme that is activated where conditional instability is present, and boundary-layer fluxes are obtained by
stability-dependent bulk formulae. Land and ice temperature anomalies are determined by a simple one-layer thermodynamic model. In this study the AGCM is configured with eight vertical (sigma) levels and with a spectral truncation at total wavenumber 30. Applications of the ICTP AGCM can be found in Bracco et al. (2004) and Kucharski et al. (2006a, 2006b).

The model climatology for boreal summer, the season in which we concentrate our analysis, is compared with observations in Figure 4. Figure 4(a) and (b) show the CMAP and GPCP JAS rainfall and the 925 hPa wind climatology from the NCEP/NCAR reanalysis for the period 1980–2002, Figure 4(c) displays the CRU JAS rainfall and NCEP/NCAR 925 hPa winds over the period 1950–2002, and Figure 4(d) shows the corresponding climatological fields for the model from 1950–2002. Overall the ICTP AGCM captures the main features of precipitation and low-level winds for both the African and Asian monsoon systems, although the rainfall over Africa extends too far northward compared with observations. The Indian monsoon precipitation, on the other hand, is concentrated over the ocean and is weaker than that observed over land. This is due to the inability of most (if not all) AGCMs to represent correctly the relationship between surface fluxes and SSTs over the Indian basin (Wang et al., 2005; Wu and Kirtman, 2005; Bracco et al., 2007). A dramatically improved simulation of the Indian monsoon can be obtained by regionally coupling the ICTP AGCM to an ocean model over the Indian basin (Bracco et al., 2007; K2007/8). In this work, however, we are interested in keeping the experimental set-up as simple as possible, to allow for easy reproducibility of the results, and we do not adopt the coupled configuration.

We conduct five sets of experiments. In the control run, the ICTP AGCM is forced by climatological SSTs. In the two further sensitivity experiments, EXP1 and EXP2, an SST anomaly is superimposed on the climatological values in the tropical Atlantic, in the region between 30°S and 30°N. Following K2007/8, this anomaly has been derived by computing a regression of the SST onto the JAS STA index for the period 1950–2002, and is confined almost entirely to the southern hemisphere (see Figure 1 for twice the anomaly). In Section 2 and in K2007/8 we have shown that SSTs in this region influence the Indian summer monsoon, with warmer (cooler) than normal SSTs favouring a decrease (increase) in precipitation over India. The derived anomaly is added to the climatological SSTs in EXP1 to provide a warm anomaly, while it is subtracted in EXP2 to cool the tropical Atlantic. Applying anomalies of equal magnitude and opposite sign allows

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**Figure 4.** Precipitation and 925 hPa wind climatology for JAS from (a) CMAP and NCEP/NCAR (1980–2002) re-analysis, (b) GPCP and NCEP/NCAR (1980–2002) re-analysis, (c) CRU and NCEP/NCAR (1950–2002) re-analysis and (d) the model (1950–2002). Units are mm day$^{-1}$ for precipitation and m s$^{-1}$ for wind. This figure is available in colour online at www.interscience.wiley.com/journal/qj
us to assess the existence of significant nonlinearities in the response to the tropical Atlantic SSTs. In fact, we find that in the region extending from Central America to South Asia, the responses to positive and negative SST anomalies are fairly similar in both amplitude and pattern, but of opposite sign, suggesting that nonlinearities are negligible, at least in our runs. In the following the response in any chosen field is defined as the difference between EXP1 and EXP2.

A second set of sensitivity experiments has been performed to assess the possibility that the reduced rainfall and heating induced by the tropical Atlantic teleconnection in the Indian basin feed back to the response seen in EXP1 and EXP2. We prescribe in the northern Indian Ocean (60°E–95°E and 5°N–25°N) a SST anomaly scaled to produce a rainfall (heating) perturbation comparable in strength to that induced in the Indian region by the remote response to the tropical Atlantic. In EXP3 the anomaly is subtracted from the climatological SSTs to provide a cold pattern, while in the partner experiment EXP4 it is added to warm the northern Indian Ocean. The resulting EXP3–EXP4 SST anomaly is shown in Figure 5 and has a magnitude of about 0.25 K.

For all sensitivity experiments the model is run for 100 JAS seasons. The atmosphere at each season is initialized using the state generated from a continuous 100 year run of the ICTP AGCM forced by the observed SST. As a note of caution, the reader should consider that there may be possible imbalances between the initial conditions of the modelled atmosphere derived from the run forced by observed SSTs and the idealized cases described above. The assumption we made is that such imbalances do cancel out when considering the difference between EXP1 and EXP2, or between EXP3 and EXP4.

4. Results

4.1. Time-mean responses to south tropical Atlantic anomalies

Figure 6 shows the time-mean responses of (a) precipitation and 925 hPa wind, (b) surface pressure, (c) 925 hPa streamfunction and (d) 200 hPa eddy streamfunction to the tropical Atlantic SST anomalies shown in Figure 1. The responses are calculated as the means from days 20–90, with the first 19 days discarded as transient. Shaded regions are 95% statistically significant according to a t-test under the assumption that the 100 realizations are independent.

Over Africa the precipitation signal consists of an increase near the Coast of Guinea, in the area within 10E–10W and 5N–10N, and a weak decrease further to the east. This is in agreement with Figure 2 and previous results by Vizy and Cook (2001) and Giannini et al. (2003). Vizy and Cook (2001) analyzed AGCM experiments performed on an idealized planet including only the land mask of the African continent with flat topography and forced by prescribed SST anomalies in the Gulf of Guinea. According to this work, precipitation near the Guinean Coast increases following a Gill-type response to the positive SST anomalies. Those numerical simulations differ from ours because the prescribed anomalies extend over a smaller area over the tropical Atlantic. More importantly, the aqua-planet set-up outside the African Continent inhibits interactions with the Indian monsoon system, which is the focus of the present work. The influence of south tropical Atlantic SSTs on West African rainfall has been confirmed by the realistic AGCM experiments of Giannini and co-authors.

Our runs further support the interpretation of the rainfall pattern over Africa as part of a Gill–Matsuno-type response: a warming over the tropical Atlantic induces a pressure gradient extending from the north-east to the south-west over Africa (Figure 6(b)). The surface winds are driven from the higher pressure over north-east Africa to the lower pressure over south-west Africa and the Atlantic Ocean. This induces an overall low-level convergence and increased rainfall over western equatorial Africa, and low-level divergence to the east of the gradient associated with reduced rainfall. The opposite happens for a cold tropical Atlantic anomaly.

Further to the east, the patterns depicted in Figure 6 are consistent with a weakening (strengthening) of the Indian monsoon in case of a warm (cold) tropical Atlantic SST anomaly, as derived from the observations in Figure 2. Focusing on the Indian monsoon region, the outcome of our idealized experiments agrees with that of Kucharski et al. (2007; see their figure 6) and Kucharski et al. (2008a; see their figure 3). In summary, the pattern of the upper-level response in streamfunction of Figure 6(d) is a quadrupole with an upper-level trough (ridge) over India and the Arabian Sea related to a positive (negative) anomaly in SST over the Atlantic. The vertical structure of the response is baroclinic with a positive (negative) surface pressure and low-level streamfunction anomaly over India (Figure 6(b) and (c)) that favours low-level divergent (convergent) flow, causing a decrease (increase) in monsoon rainfall.

When comparing Figure 6(a) in this article with figure 3(b) of Kucharski et al. (2008a), a small but noticeable difference is that here the precipitation response in the Indian basin appears to be slightly shifted to the east, over the ocean. Furthermore, the increase of rainfall over Bangladesh corresponding to warm south tropical Atlantic SSTs seen in Figure 2 of this article and in figure 3(b) of Kucharski et al. (2008a) is not reproduced in the precipitation response seen in our Figure 6(a). It is likely that the lack of coupling in the Indian Ocean in the present idealized experiments may account for such a
Figure 6. Response to the SST anomaly in Figure 1 (EXP1–EXP2) of (a) precipitation and 925 hPa winds, (b) surface pressure, (c) 925 hPa stream function and (d) 200 hPa eddy streamfunction, averaged from days 20–90. Shading indicates regions of 95% statistically significant anomalies (positive: orange/light grey, negative: blue/dark grey). Contour intervals are 0.6 mm day$^{-1}$ for precipitation in (a), 0.2 hPa in (b), $2 \times 10^6$ m$^2$s$^{-1}$ in (c) and $3 \times 10^6$ m$^2$s$^{-1}$ in (d). The unit of the wind response in (a) is m s$^{-1}$. This figure is available in colour online at www.interscience.wiley.com/journal/qj

The precipitation signal extends also to the west, over Central America and North Brazil, where positive STA anomalies cause a significant increase in boreal summer rainfall. The response in the Central America/Caribbean region is consistent with the observed late summer/early autumn correlations of rainfall with tropical Atlantic SSTs (Taylor et al., 2002).

We can interpret the precipitation response to the STA SST anomaly as a proxy for the heating one, and compare it with the work of JH95. Obviously, this response is more complicated than in baroclinic models where the heating can be directly prescribed, as in JH95. Nevertheless, the upper-level streamfunction response depicted in Figure 6(d) is broadly consistent with the one shown in JH95 for a heating anomaly in the Pacific (their figure 2(c)). As in the idealized case of a Gill–Matsuno-type response in the absence of a mean flow, the waves are equatorially trapped and there is no Rossby-wave propagation into the extratropics. This is confirmed by Figure 7(a) and (b), where the 200 hPa zonal winds in JAS are displayed for the NCEP/NCAR re-analysis and the model, respectively. Indeed, in the regions just north of the Equator in the Atlantic basin and over Africa and the Indian Ocean, the zonal winds mostly have a negative sign, favouring a purely equatorially trapped Gill–Matsuno response. In comparison with JH95 results, however, our experiments display a slightly stronger eastward response in the upper level streamfunction field.

Figure 8(a) and (b) show the vertically integrated moisture flux and moisture flux convergence (see e.g. Hagos and Cook (2008) for a discussion of the relevance of the moisture flux). The local redistribution of moisture is strongly confined in two regions, one between northeast Sahelian Africa and the Gulf of Guinea Coast, the other between the Indian region and the equatorial Indian Ocean, with a small net exchange of moisture between the Indian peninsula and the Gulf of Guinea Coast. This may be interpreted as part of the total climate adjustment process to the heating in the south tropical Atlantic region, and does not imply that the west African rainfall and Indian rainfall are inherently connected.
The mechanism we advocate for the rainfall reduction over the Indian region is dynamical. The northeastern part of the baroclinic quadrupole response favours descending motion over the Indian region; at the same time subsidence and precipitation anomalies change in concert as part of the total atmospheric adjustment to the tropical Atlantic heating, and the net result is a moisture flux divergence and suppressed precipitation. The reader should bear in mind that there is a possible thermodynamic interpretation for the reduction (increase) of precipitation in regions remote from the warm (cold) SST anomaly. Lintner and Chiang (2007) have recently shown in idealized numerical experiments with an ENSO-type SST anomaly that an important contribution to the reduction of rainfall in remote tropical regions is the warming of the total atmospheric column away from the heating itself, which induces a stabilization of the atmosphere. Figure 8(c) shows the response of the vertically averaged atmospheric temperature profile. As suggested in Lintner and Chang (2007), the atmospheric column warms to the east of the warm SST anomaly over eastern Africa and in the equatorial Indian Ocean, but not over the Indian subcontinent, where a cooling occurs instead. Thus in our simulations the stabilization of the atmosphere may be relevant to reducing rainfall in the eastern African region, but dynamical processes dominate over thermodynamic ones in the Indian peninsula. The cooling is likely due to the reduced atmospheric moisture caused by a sinking motion that minimizes the amount of heat trapped in the atmosphere. Interestingly, the vertically averaged temperature response shows a rather symmetric structure north and south of the equator in the Indian Ocean region, with warming at about 20°N–30°N and 30°S–20°S. The 200 hPa eddy streamfunction response in Figure 6 shows a similar symmetry. This is indeed suggestive of forced sinking motion in both regions, and is confirmed by the analysis of the 500 hPa omega field (not shown). We also verified that the cooling in both regions is due to reduced downward long-wave radiation. The main difference between the northern and the southern regions is that the southern one is dry and without precipitation in the climatological mean (see Figure 4), and thus displays a very limited precipitation response.

4.2. The role of heating over the Indian region

The time-mean responses to SST anomalies in the south tropical Atlantic show considerable rotational structures to the east of the anomalies, particularly near the Indian region (see Figure 6). Although JH95 also found some rotational structure to the east of their heating anomaly, this effect seems to be more pronounced in our simulations. Furthermore, although the upper-level structure of the eddy streamfunction is fairly (anti-)symmetric north and south of the equator, the surface pressure response clearly is not.

To investigate a potential cause for the differences between our experiments and those of JH95, we performed EXP3 and EXP4. In those runs a SST anomaly is imposed over the Indian Ocean and its intensity (−0.25 K, see Figure 5) is scaled such that the resulting precipitation anomalies in the Indian region are similar to the ones found in the responses of EXP1 and EXP2 (see Figure 6). Those simulations allow us to assess whether and how reduced rainfall and heating in the Indian region influence the structure of the response, possibly enhancing it and causing the pronounced rotational structure. Those are the only goals of these simulations, because the existence of an isolated SST anomaly in the northern Indian Ocean cannot be justified in any way, owing to the strong teleconnections of the SSTs in the northern Indian Ocean with the equatorial Indian and eastern Pacific Ocean SSTs (Annamalai and Murtugudde, 2004). Furthermore the response to such an SST anomaly may be overestimated; many studies have indeed shown that SST forcing in the Indian Ocean region induces spurious responses in AGCM simulations (Krishna Kumar et al.,...
Figure 9. Response to the SST anomaly in Figure 5 (EXP3–EXP4) of (a) precipitation and 925 hPa winds, (b) surface pressure, (c) 925 hPa stream function and (d) 200 hPa eddy streamfunction, averaged from days 20–90. Shading indicates regions of 95% statistically significant anomalies (positive: orange/light grey, negative: blue/dark grey). Contour intervals are 0.6 mm day$^{-1}$ for precipitation in (a), 0.2 hPa in (b), $0.2 \times 10^6$ m$^2$ s$^{-1}$ in (c) and $0.3 \times 10^6$ m$^2$ s$^{-1}$ in (d). The unit of the wind response in (a) is m s$^{-1}$. This figure is available in colour online at www.interscience.wiley.com/journal/qj

2005; Bracco et al., 2007; Kucharski et al., 2008b) and may lead to the breakdown of the correct representation of the ENSO–Indian monsoon relationship in the models.

Figure 9 shows the time-mean responses for EXP3 and EXP4, analogous to those of Figure 6. In the Indian region the resulting patterns are strikingly similar to those of Figure 6. In both cases a high surface pressure is under development and the structure of the response is baroclinic, with a reduction in the streamfunction at 200 hPa. Interestingly, although the rainfall anomaly is of similar magnitude in the Indian region to that in EXP1 and EXP2, the pressure response is only about half of the response found in EXP1 and EXP2. Overall, EXP3 and EXP4 support the hypothesis that the high pressure over India is dynamically induced and is subsequently enhanced by the development of negative heating anomalies in that region. Furthermore, the negative heating over India appears to weaken the Kelvin-wave response further to the east, limiting its propagation into the Pacific basin, as indicated by the upper-level streamfunction response to the east of the heating anomaly shown in Figure 9, which has the opposite sign to that of Figure 6.

4.3. Transient response

We now describe the time evolution of the atmospheric response to the SST anomaly in the tropical Atlantic during the first eight days or transient period, to illustrate how the signal is propagated by Kelvin and Rossby waves. One shortcoming of our runs is that SSTs change in a discontinuous way (the full SST anomaly is imposed at the start of the integrations on 1 July) and the initial imbalance in the atmospheric model state may potentially bias the results. Nonetheless, we think we can gather valuable information from this exercise by considering the differences between EXP1 and EXP2. Figure 10 shows the 200 hPa eddy streamfunction response averaged over (a) the first two days, (b) days 3–4, (c) days 5–6 and (d) days 7–8. By day 8 the response has reached into the western Pacific to the east and the eastern Pacific to the west. The classical quadrupole structure becomes evident at days 3–4, in agreement with more idealized results (e.g. figure 2(c) of JH95).

The time evolution of the 200 hPa velocity potential response (Figure 11), shown again for (a) the first 2 days, (b) days 3–4, (c) days 5–6 and (d) days 7–8, hints at how the Walker circulation connecting the African and Asian monsoons is modified by the presence of a SST anomaly in the Atlantic. The upper-level divergence (i.e. velocity potential minimum) over the tropical Atlantic and Africa is compensated for mainly by upper-level convergence (i.e. velocity potential maximum), initially over the Indian Ocean/western Pacific region and later extending into the central Pacific. This indicates that the relative position of the response to the initial SST anomaly in the tropical Atlantic is modified by the location and strength of the Walker circulation associated with the main monsoon systems, as proposed by JH95.

A good description of the propagation of the equatorially trapped responses to the east and to the west is provided by the longitude–time plot of the zonal wind responses averaged from 5°S–5°N (Figure 12). The resulting picture again matches very closely figure 4 in JH95. The warm (cold) SST anomaly in the equatorial Atlantic causes upper-level divergence (convergence) and leads to upper-level westerly (easterly) wind perturbations that propagate as Kelvin waves to the east of the anomaly.
with a speed of about 20° per day, in agreement with analogous idealized experiments by Lintner and Chiang (2007). Such a speed is about half of the propagation velocity in the Gill–Matsuno solution (about 45 m s$^{-1}$, corresponding to about 40° per day). Such a reduction is due to the inclusion of moist dynamics, which modifies the Kelvin wave propagation with respect to the dry Gill–Matsuno model (Neelin and Su, 2005). Mekonnen et al. (2008), in a recent article, showed the existence of such Kelvin waves in the African region in observations and found that their activity is positively correlated with tropical eastern Pacific and Atlantic, including Gulf of Guinea, SSTs (their figure 9). The upper-level divergence (convergence) also leads to upper-level easterlies (westerlies) to the west that propagate as equatorial Rossby waves with a speed of about 6° per day.

5. Discussion and conclusions

In this work we investigated the physical mechanism by which tropical Atlantic SSTs teleconnect to the Indian Ocean basin in boreal summer. To first order, the response
to a heating anomaly in the tropical Atlantic is a classic Gill–Matsumo-type quadrupole. Kelvin waves transport the signal from the tropical Atlantic to the Indian Ocean and equatorial Rossby waves to Central America and the eastern Pacific. The surface pressure signature over India is a high (low) anomaly developing in relation to warm (cold) tropical Atlantic SSTs, thus leading to low-level divergence (convergence) and reduced (increased) rainfall over India. As already shown in K2007/8, the response projects largely on to the time-mean circulation, and a tropical Atlantic warming (cooling) weakens (strengthens) the time-mean Walker circulation. As a consequence, local feedbacks with the time-mean Walker circulation may further contribute to strengthening the teleconnection between the SSTs in the tropical Atlantic and the atmospheric circulation over the Indian basin. This could explain the amplitude of the eastward response found in our experiments, which is slightly stronger than predicted by the Gill–Matsumo mechanism. Indeed, a sensitivity experiment with a cold SST anomaly in the northern tropical Atlantic, resulting in a high (low) anomaly developing in relation to warm (cold) SSTs, thus leading to low-level heating anomalies in the tropical Atlantic SST gradient during the anomalous evolution in the Indian region increases the surface pressure locally, enhancing the Gill–Matsumo response to a warm anomaly in the south tropical Atlantic.

The observation that in summer heating anomalies in the tropical Atlantic/African region modulate the Indian monsoon is suggestive of a possible link between the African and Indian monsoon systems. However, to take advantage of this relation seasonal prediction systems will need skill at forecasting the evolution of SSTs in the tropical Atlantic. Unfortunately, this is one region where current coupled models suffer from significant biases (Stockdale et al., 2006). This study indicates that the enhancement of coupled model skills in the tropical Atlantic not only is imperative for the prediction of west African monsoon rainfall, but will improve the Indian monsoon seasonal forecasting.

The authors hope that the simplicity of the mechanism proposed to explain the link between tropical Atlantic SST and precipitation anomalies over Africa and India will encourage further studies using a broad range of AGCMs.

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