Possible Impact of the Indian Ocean SST on the Northern Hemisphere Circulation during El Niño*

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ABSTRACT

Two atmospheric general circulation models (AGCMs), differing in numerics and physical parameterizations, are employed to test the hypothesis that El Niño–induced sea surface temperature (SST) anomalies in the tropical Indian Ocean impact considerably the Northern Hemisphere extratropical circulation anomalies during boreal winter [January–March +1 (JFM +1)] of El Niño years. The hypothesis grew out of recent findings that ocean dynamics influence SST variations over the southwest Indian Ocean (SWIO), and these in turn impact local precipitation. A set of ensemble simulations with the AGCMs was carried out to assess the combined and individual effects of tropical Pacific and Indian Ocean SST anomalies on the extratropical circulation. To elucidate the dynamics responsible for the teleconnection, solutions were sought from a linear version of one of the AGCMs.

Both AGCMs demonstrate that the observed precipitation anomalies over the SWIO are determined by local SST anomalies. Analysis of the circulation response shows that over the Pacific–North American (PNA) region, the 500-hPa height anomalies, forced by Indian Ocean SST anomalies, oppose and destructively interfere with those forced by tropical Pacific SST anomalies. The model results validated with reanalysis data show that compared to the runs where only the tropical Pacific SST anomalies are specified, the root-mean-square error of the height anomalies over the PNA region is significantly reduced in runs in which the SST anomalies in the Indian Ocean are prescribed in addition to those in the tropical Pacific.

Among the ensemble members, both precipitation anomalies over the SWIO and the 500-hPa height over the PNA region show high potential predictability. The solutions from the linear model indicate that the Rossby wave packets involved in setting up the teleconnection between the SWIO and the PNA region have a propagation path that is quite different from the classical El Niño–PNA linkage.

The results of idealized experiments indicate that the Northern Hemisphere extratropical response to Indian Ocean SST anomalies is significant and the effect of this response needs to be considered in understanding the PNA pattern during El Niño years. The results presented herein suggest that the tropical Indian Ocean plays an active role in climate variability and that accurate observation of SST there is of urgent need.

1. Introduction

For short-term global climate prediction, the sea surface temperature (SST) anomalies associated with the

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terns noted by Horel and Wallace (1981). Ignoring many complexities, the conceptual framework of how El Niño influences the PNA pattern can be summarized as follows: positive SST anomalies in the equatorial central and eastern regions of the Pacific Ocean favor enhanced deep convection there; this leads to an increase in the release of latent heating throughout the troposphere and in the divergent flow at upper levels; subsequently, the upper-level divergence forces planetary Rossby waves that project onto the PNA pattern (Hoskins and Karoly 1981; Webster 1981; Simmons 1982; Branstator 1985). One limitation of this conceptual scenario is the requirement that Rossby wave forcing has to be in mean westerly flow while the observed forcing during El Niño years is in mean easterlies; this and other related issues are addressed in subsequent studies and the details can be found in the review article by Trenberth et al. (1998).

Many atmospheric general circulation models (AGCMs) forced by ENSO-related SST anomalies have simulated the PNA pattern with moderate success (Shukla and Wallace 1983; Blackmon et al. 1983; Geisler et al. 1985; Lau 1981; 1985; Lau and Nath 1994; Hoerling et al. 1997; Hoerling and Kumar 1997; Kumar and Hoerling 1997, among others). The observed relationships, theoretical frameworks, successful simulations by AGCMs, and, more importantly, our ability to predict tropical SST 6–12 months ahead (Latif et al. 1994) led to a great interest in the possibility of producing skillful seasonal forecasts over North America (e.g., Kumar et al. 1996; Hoerling et al. 1997; Hoerling and Kumar 1997; Shukla 1998; Barnston et al. 1999; Shukla et al. 2000; Hoerling and Kumar 2002). On the other hand, large natural variability exists over the PNA region (Lau 1981). Therefore, ensemble simulations with AGCMs are required to extract the SST-forced response. In the present study, we are interested in quantifying the forced extratropical circulation response to tropical Indian and Pacific Ocean SST anomalies during El Niño winter.

One of the outstanding issues in the tropical–extratropical linkage is our understanding of whether the SST anomalies beyond the equatorial central and eastern Pacific feed back onto the ENSO-induced atmospheric response (Alexander et al. 2002; Barsugli and Sardeshmukh 2002; Hoerling and Kumar 2002). Many observational studies show that ENSO, primarily through the atmospheric bridge, induces substantial SST variations in other tropical oceans (e.g., Pan and Oort 1983; Latif et al. 1994; Lau 1997; Klein et al. 1999). Alexander et al. (2002) provide a comprehensive list of studies that related observed SST anomalies in the tropical Pacific with anomalies in other ocean basins. The tropical Indian Ocean, the focus here, witnesses a basin-wide warming in boreal winter that peaks in the following spring after the mature phase of El Niño (Fig. 1a). Are these warm anomalies a passive response to El Niño or do they influence the PNA pattern? We are interested in answering this question by performing systematic experiments with two AGCMs.

Many past studies, using either linear models or AGCMs, have addressed this problem. The possible role of diabatic heating or SST anomalies from different ocean basins on the PNA pattern has been diagnosed using an “influence function.” The influence function (g) can be interpreted as an inverse Green’s function. That is, rather than showing how the response at all locations is affected by a specified source of forcing, g shows how the forcing from all locations affects a specified response. Based on this diagnostic, idealized forcing experiments with linear models suggested that Indian Ocean diabatic heating anomalies force a North Pacific height response opposite to that forced by tropical Pacific diabatic heating anomalies (Simmons et al. 1983; Branstator 1985; Ting and Sardeshmukh 1993; Jin and Hoskins 1995; Newman and Sardeshmukh 1998). Alexander et al. (2002) estimated the sensitivity of a negative height anomaly over the central North Pacific in winter to anomalous SST 60 days earlier. According to their estimates, g showed that SST anomalies in the tropical Indian Ocean force a North Pacific height response opposite to that forced by eastern-central Pacific SST anomalies. Using a National Centers for Environmental Prediction (NCEP) AGCM, Barsugli and Sardeshmukh (2002) pointed out that there is a nodal line in the sensitivity at about 100°E, with warm SST anomalies in the Indian Ocean resulting in a negative value of the PNA index. Earlier AGCM studies of Kumar and Hoerling (1998b), Farrarra et al. (2000), and Spencer et al. (2004) also pointed out the possible role of Indian Ocean SST anomalies on the PNA pattern.

Before claims can be made about the active role of Indian Ocean SST anomalies on the extratropical circulation variability, it is necessary to understand two aspects: (i) the processes responsible for the SST variations, and (ii) the local SST–rainfall relationship. Observational (Klein et al. 1999) and ocean modeling (Murtugudde and Busalacchi 1999) studies find that surface heat flux is the major mechanism for El Niño–induced warming in the northern Indian Ocean. To substantiate this further, Alexander et al. (2002) forced an AGCM with prescribed ENSO-related SST anomalies over the tropical Pacific while outside the forcing region the AGCM was coupled to a mixed-layer model. The predicted SST from these ensemble solutions captures the variability over the northern Indian Ocean.
realistically but that over the southwest Indian Ocean (SWIO; 15°S–0°, 50°–80°E) is modest at best. In a similar modeling effort, Lau et al. (2005) also note a reduction in the simulated SST variability over the SWIO. In these studies, two possible limitations are that the mixed layer model used is far too simple, and the role of the ocean dynamics is not considered. In summary, the atmospheric bridge mechanism does not explain all of the observed SST fluctuations over the SWIO, where the anomalies typically reach a local maximum (Fig. 1a).

The primary motivation for the present study stems from a recent recognition that thermocline variations over the SWIO influence SST anomalies there (Xie et al. 2002; Huang and Kinter 2002). The fact that subsurface (thermocline) variability exerts a strong influence on SST over the SWIO is revealed by the high simultaneous local correlation between them (Fig. 2a). Xie et al. (2002) demonstrated that ENSO is the dominant forcing for the SWIO thermocline variability. They noted that when an El Niño event occurs, anomalous surface easterlies appear in the equatorial Indian Ocean, forcing a westward-propagating downwelling Rossby wave in the southern Indian Ocean.

Do local SST anomalies determine precipitation anomalies over the SWIO, or in other words, do they result in a predictable atmospheric signal? Since large SST anomalies in the Indian Ocean often co-occur with those in the Pacific, Annamalai et al. (2005a) tested the sensitivity of precipitation variations over the SWIO (Fig. 1b) to SST anomalies from different ocean basins from a suite of ensemble AGCM experiments. They concluded that precipitation anomalies over the SWIO are largely tied to local SST anomalies. Further observational evidence is presented here. Figure 2b plots the simultaneous correlation between SST and precipitation for January–March. Significant positive correlations are noticeable over the SWIO and along the Somali coast in the western Indian Ocean. One inference is that interannual SST variations (Fig. 1a) exert a strong thermodynamic impact on the local precipitation variations (Fig. 1b). The key point here is that the interrelationship among the observed quantities (thermocline → SST → precipitation) bears coherency over the SWIO (Fig. 2), the region of relevance for the present research. In contrast, negative correlations over the eastern-equatorial Indian Ocean suggest the role of remote forcing. In other words, warm SST anomalies there (Fig. 1a) are a result of reduced precipitation (Fig. 1b). A note of caution should be offered here: the regions of maximum positive correlation between SST and precipitation (Fig. 2b), and between SST and thermocline anomalies (Fig. 2a), are not collocated, reflecting the nonlinearity in the SST–convection relationship over the Indian Ocean. It should be pointed out here that the seasonal mean (January–March, JFM) SST over the SWIO is >28°C, the threshold required for

Fig. 1. Seasonal average (January–March) composite anomalies of (a) SST (°C) and (b) precipitation (mm day⁻¹) from CMAP, and (c) 500-hPa height anomalies. In all panels positive (negative) values are shaded progressively (dashed contours). The contour interval (CI) is (a) 0.5°C, (b) 0.5 mm day⁻¹, and (c) 15 m. The 0 contour is not shown. Eight El Niño years chosen are (1951/52, 1957/58, 1963/64, 1969/70, 1972/73, 1982/83, 1986/87, and 1997/98); for observed precipitation (b), only the last 3 yr are used. In (c) the three major centers associated with the PNA pattern are shown with boldface signs.
the occurrence of deep convection (Annamalai and Murtugudde 2004). During El Niño years, observed warm SST variations over the SWIO are in the range of 0.5°–0.8°C (Fig. 1a), and due to the nonlinearity in the Claussis–Clapyeron equation, these small SST variations can result in large precipitation anomalies. In fact, Graham and Barnett (1987) showed that tropical convection increases rapidly in the range of about 1°C around 28°C.

In summary, diagnostic studies by linear dynamical models in the 1980s and 1990s, and more recently sensitivity experiments with AGCMs, highlight the possible impact of tropical Indian Ocean precipitation on the North Pacific and North American climate during boreal winter of El Niño years. On noting the low correlation between basin-wide indices of Indian Ocean SST and precipitation, some AGCM studies (e.g., Kumar and Hoerling 1998b) have disallowed the Indian Ocean as a source of predictability. But a careful examination on regional scales indicates evidence for a significant SST–precipitation association in the SWIO (Fig. 2a). Having recognized that SWIO SST variations are forced by ocean dynamics (Fig. 2a), we have studied the effect of Indian Ocean SST anomalies on the PNA pattern more carefully.

In the present study, we systematically evaluate the effect of Indian Ocean SST anomalies on the extratropical circulation, with a special emphasis on the North Pacific height anomaly. Our focus is on the 3 months (January–March; JFM+1) of the year following the mature phase of El Niño when the extratropical response to tropical SST changes is large and highly predictable (Kumar and Hoerling 1998a; Peng and Kumar 2005). However, the response to SST changes upon the model atmosphere appears to depend on the details of the model numerics and physics, as well as the configuration of the climatological background flow (Ting and Sardeshmukh 1993; Lau 1997). To accomplish robust results, we perform identical experiments with two different AGCMs that differ considerably in numerical formulation and physical parameterizations. To examine the potential predictability, ensemble simulations are conducted. Finally, to understand the dynamics involved and the sensitivity of the results to the AGCMs’ basic state, experiments are performed with a linear version of one of the AGCMs.

In general, our results from the AGCMs confirm the recent findings from other studies. The salient new results of the present study are (i) local SST anomalies control precipitation variations over the SWIO and (ii) the root-mean-square error in the 500-hPa height response over the PNA region between the reanalysis and AGCM solutions reduces by about 15%–42% when the Indian Ocean effect is included with the Pacific. Linear model solutions reveal that Rossby waves excited by Indian Ocean heating do not follow the great circle, but are refracted toward the effective beta maximum associated with the Asian jet curvature. Further sensitivity experiments indicate that the zonally varying ambient flow is instrumental for the substantial PNA response to Indian Ocean heating. Based on these model solutions, we conclude that the effect of Indian Ocean SST anomalies on the extratropical circulation anomalies is real and significant and, therefore, needs to be considered for a realistic simulation of the PNA pattern during the winter of El Niño years.

The paper is organized as follows. Section 2 presents...
the models used along with details about the experimental designs. AGCMs results on the tropical–extratropical linkages are presented in section 3. The solutions from the linear model and the processes involved in the teleconnection are discussed in section 4. The summary and implications of the results are discussed in section 5.

2. Data, models, and experimental designs

a. Data

The atmospheric data used in our study are taken from the NCEP–National Center for Atmospheric Research (NCAR) reanalyses products for the period 1950–2003 (Kalnay et al. 1996). The atmospheric variables are available at standard pressure levels with a horizontal resolution of 2.5°. The SST for the analysis period is taken from Reynolds and Smith (1994). In addition, we utilize the monthly mean rainfall data from the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP; Xie and Arkin 1996) for the period 1979–2003. The thermocline depth (h), represented by the depth of the 20°C isotherm, is taken from the Simple Ocean Data Assimilation (SODA; Carton et al. 2000). The usefulness of SODA for studying the Indian Ocean coupled dynamics has been demonstrated in Xie et al. (2002).

b. AGCMs

For identical forcings, the reproducibility of the results from different AGCMs enhances the confidence in the major conclusions arrived at. This is important since systematic biases in AGCMs can distort the underlying sensitivity to forcing, and therefore parallel analyses using ensemble simulations of different AGCMs are clearly needed (Kumar and Hoerling 1998a).

1) ECHAM5

We use ECHAM5, the latest Hamburg version of the European Centre for Medium-Range Weather Forecasts (ECMWF) model. It is a global spectral model, which we ran at T42 resolution and with 19 sigma levels in the vertical. The nonlinear terms and the parameterized physical processes are calculated on a 128 × 64 Gaussian grid, with a horizontal resolution of about 2.8° × 2.8°. As in the earlier version of the model (ECHAM4; Roeckner et al. 1996), the convection scheme is based on the mass-flux concept of Tiedtke (1989); the surface fluxes of momentum, heat, and water vapor are based on the Monin–Obukhov similarity theory; and the radiation scheme is derived from Morcrette et al. (1998). Major changes in the model include implicit coupling of the atmosphere to the land surface (Schulz et al. 2001), advective transport (Lin and Rood 1996), a prognostic–statistical scheme for cloud cover (Tompkins 2002), and a rapid radiative-transfer model for longwave radiation (Mlawer et al. 1997). Model details may be found in Roeckner et al. (2003). This AGCM has been shown to realistically simulate the response over the tropical Indo-Pacific oceans during the entire life cycle of El Niño (Annamalai and Liu 2005; Annamalai et al. 2005a,b).

2) CCSR

Another AGCM used here is the latest version of the model cooperatively developed at the Center for Climate System Research (CCSR) of the University of Tokyo and the National Institute for Environmental Studies. Hereafter, we simply refer to this model as CCSR. Like ECHAM5, it is a global spectral model with a horizontal resolution of T42 and with 20 sigma levels in the vertical. The physical parameterizations include a sophisticated radiative transfer scheme (Nakajima and Tanaka 1986), a simplified Arakawa–Schubert cumulus convection scheme (Arakawa and Schubert 1974), a prognostic cloud water scheme (Le Treut and Li 1991), a bulk surface fluxes scheme (Louis 1979), and a simple land surface model (Manabe et al. 1965). Detailed descriptions and general performance of the model can be found in Numaguti et al. (1995) and Sumi (2000). This AGCM has demonstrated skill in simulating the tropical response to prescribed tropical SST anomalies (Sumi 2000; Shen et al. 2001).

Palmer and Mansfield (1986), Ting and Sardeshmukh (1993), and several others have concluded that the extratropical response to tropical heating relies heavily on the AGCM’s basic state. The ECHAM5 and CCSR simulated climatologies, based on 7-yr means, are realistic in many aspects. Figure 3 compares the precipitation and 200-hPa zonal wind climatologies between the models and observations. It is worth pointing out that in both models, in the Northern Hemisphere, the location and intensity of the upper-level Asian jet are accurately simulated. Compared to observations, in both AGCMs, the westerlies protrude into the equatorial eastern Pacific, and penetrate northwestward over the Asian continent in CCSR. These differences in the structure, and the associated vorticity gradients, would influence the propagation of Rossby waves (Hoskins and Karoly 1981; Ting and Sardeshmukh 1993), an issue investigated in section 4b. Model–observation differences in precipitation include the following: (i) overestimation of the precipitation intensity along the South Pacific
convergence zone (SPCZ) and off of Madagascar in the southern Indian Ocean and (ii) underestimation of precipitation intensity over the equatorial central Indian Ocean. Despite these differences, in the sensitivity experiments described below, the precipitation patterns over the tropical Indo-Pacific region, and the Northern Hemisphere extratropical circulation, are reasonably simulated by both models. Our intention here is not to intercompare the model results but to achieve robust responses, at least qualitatively, to the prescribed identical forcing.

c. Experimental designs

We carry out a set of three ensemble experiments (Table 1). In the tropical Indo-Pacific (TIP) runs, seasonally varying SST anomalies are imposed in the tropical Indo-Pacific region from 30°S to 30°N and seasonally varying climatological SSTs are imposed elsewhere. The tropical Pacific Ocean (TPO) runs are similar to the TIP runs, except that SST anomalies are inserted only into the tropical Pacific. The tropical Indian Ocean (TIO) runs are like the TIP runs, but with SST anomalies imposed only in the tropical Indian Ocean. This set of experiments is expected to reveal the individual and combined effects of SST anomalies on the local and remote climate variabilities. For example, the difference between TIP and TPO solutions will suggest the effect of Indian Ocean SST anomalies when the Pacific is interactive, while the TIO and TPO solutions indicate the noninteractive effect. Finally, to elucidate the exclusive local effect, an additional 25-member ensemble run wherein the CCSR AGCM is forced only with SWIO SST anomalies is conducted (Table 1).

For each of the above forcing scenarios, both AGCMs are run for 7 months, from December through June. A composite El Niño anomaly is added to the climatological SST for each month from December of the year 0 through June of year +1. The composite El Niño is based on the average for 8 yr when December–February DJF (0/1) mean SST anomalies over the Niño-3.4 region and JFM (+1) mean SST anomalies over the SWIO (15°S–0°, 50°–80°E) exceed one standard deviation. The preference for choosing the JFM SST rather than the DJF SST average over the SWIO is due to the known delay in the thermocline response over the SWIO due to ENSO forcing (Xie et al. 2002, their Fig. 9b). Prior to making the composite, any SST variability longer than 8 yr is removed. To account for the atmospheric sensitivity to initial conditions, a 10-member ensemble approach, with changes only in the initial conditions varying from 1–10 December but pre-
serving the same SST forcing, are conducted. An initial examination of the solutions reveals that AGCM simulations have more intrinsic variability. Therefore, to enhance the signal to noise ratio, all of the sensitivity experiments with this AGCM are carried out with a 25-member ensemble, with initial conditions varying from 1–25 December. In all cases, initial conditions are taken from the control (CTL) run, with the monthly climatological SST as the boundary forcing. We skip the model output for the first month to minimize the influence of the initial conditions. The composites (both constructed from observations and ensembles of AGCM solutions) are subjected to t testing for statistical significance. Values significant at the 95% level or greater are plotted.

d. Simple linear atmospheric model

The model used here, which we refer to as the linear baroclinic model (LBM), is described in detail by Watanabe and Kimoto (2000) and Watanabe and Jin (2003). It is a global, time-dependent, primitive equation model, linearized about the observed climatology derived from the NCEP–NCAR reanalysis. It is a linear version of the CCSR AGCM having horizontal and vertical resolutions as described above. The model employs diffusion, Rayleigh friction, and Newtonian damping with a time scale of (1 day)$^{-1}$ for $\alpha = 0.9$ and $\sigma = 0.03$, while (30 day)$^{-1}$ is used elsewhere. The relatively weaker damping in the free atmosphere mimics nonlinearity in linear models (Ting and Yu 1998), while the top-level stronger damping aids in absorbing vertically propagating waves (Wu et al. 2000). This model has been successfully used to understand the linear dynamics due to tropical heating (e.g., Annamalai and Sperber 2005). Here, the prescribed forcing is anomalous diabatic heating proportional to the simulated precipitation anomalies by AGCMs. More details about the experimental designs are provided in section 4.

3. Results from the AGCMs

From all the model solutions, we present 3-month (JFM) averages of precipitation and 500-hPa geopotential height anomalies (ensemble mean minus model climatology). Wherever appropriate, the model results are compared with observations–reanalysis data, but a limitation with the reanalysis composite is that it is a mixture of an SST forced signal and a contribution from the internal dynamics. In the analysis of the 500-hPa height anomalies, we focus on the models’ ability to capture the main centers of action over the PNA region [depicted as plus (+) and minus (−) signs in Fig. 1c], and the robustness between the two model solutions or between an individual model and the reanalysis fields. These are examined through estimates of standard statistics such as pattern correlation and root-mean-square error (RMSE) over the central North Pacific (25°–70°N, 160°E–120°W), and over the entire PNA region (20°–70°N, 160°E–40°W). In these calculations, to be consistent with the reanalysis the model results are gridded at $2.5^\circ \times 2.5^\circ$ resolution. The statistics are summarized in Table 2. The signatures over the Southern Hemisphere extratropics will be the focus of future research.

a. Response to tropical Pacific SST anomalies

For a prescribed SST forcing, the success of capturing a realistic extratropical teleconnection crucially depends on the AGCM’s ability to simulate the details in tropical rainfall. Figure 4 shows precipitation simulated by the TPO solutions from ECHAM5 (Fig. 4a) and CCSR (Fig. 4b). Relative to observations, in both models, the main dipole pattern in precipitation along the equatorial Pacific is obtained reasonably well, but the response is stronger in ECHAM5, while the ITCZ and SPCZ are weaker in CCSR. Compared to the observations (Fig. 1b), a key point to note is that both AGCMs fail to capture the positive precipitation anomalies over the SWIO.

The cumulative effect of the simulated precipitation on the circulation is global (Fig. 5). In both models, in the Northern Hemisphere, the stationary wave response depicts a wave train arching from the tropical Pacific across North America, having maximum amplitude over the PNA region. Compared to the reanalysis results (Fig. 1c), a closer examination at the solutions, however, indicates that in CCSR the centers of the Aleutian low and the positive height anomalies over northwest America are shifted eastward by about 25°. Similarly, in ECHAM5, the center of the positive height anomalies over northwest America are shifted westward by about 40°, but the spatial structure associated with the Aleutian low is well captured. Relative to the reanalysis data, the magnitude of the response in ECHAM5 is too strong, while that in CCSR is reasonable. For example, over the central North Pacific, the maximum absolute height value is about 160 m (75 m) in ECHAM5 (CCSR). The spatial shift and the differences in magnitude between the two model solutions account for the low pattern correlation and high RMSE over the North Pacific but the statistics show some improvement when this value is estimated over the larger PNA region (Table 2a). In summary, from the statistics between the individual model and the reanalysis (Table 2b), one can infer that the spatial pattern is better cap-
b. Response to tropical Indian Ocean SST anomalies

The precipitation response from the TIO solutions is presented in Fig. 6. In contrast to the TPO runs (Fig. 4), both AGCMs forced by Indian Ocean SST anomalies capture the observed (Fig. 1b) positive precipitation anomalies over the SWIO; however, there is disagreement in the latitudinal position of the local maximum. The message is that the TIO solutions confirm the inference made from observations (Fig. 2b) that SWIO SST anomalies determine the precipitation anomalies there.

Figure 7 plots the 500-hPa height response along with the 200-hPa streamfunction anomalies for ECHAM5. One notes that Indian Ocean forcing excites wave trains with amplitudes that are large over the North Pacific. The spatial pattern as inferred from the streamfunction anomalies (Fig. 7c) is suggestive of a diabatically forced wave train originating from the Indian Ocean. The wave activity arching northeastward from the forcing region appears to be modulated by the strength of the Asian jet. More discussion on the dynamics involved in setting up the teleconnection is provided in section 4.
One salient result from Figs. 7a and 7b lies in the vicinity of central North Pacific. Both AGCMs capture positive height anomalies there in contrast to negative anomalies simulated by TPO solutions (Fig. 5). To a good approximation, similar out-of-phase features between the TIO and TPO solutions are also noted over the other main centers of action over the PNA region. It is sufficient to say that the height anomalies forced by

Table 2a. Pattern correlation and RMSE for the 500-hPa height anomalies estimated between the ECHAM5 and CCSR model solutions.

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<th>Model runs</th>
<th>Pattern correlation</th>
<th>RMSE</th>
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<td></td>
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<tr>
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<td>TIP</td>
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Table 2b. Pattern correlation and RMSE for the 500-hPa height anomalies estimated between the model solutions and the reanalysis data.

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<th>Pattern correlation</th>
<th>RMSE</th>
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the Indian Ocean SST anomalies oppose and destructively interfere with those forced by the tropical Pacific SST anomalies over the PNA region (Fig. 5). Such a behavior may conceivably be of some importance in the real atmosphere. Quantitatively, the positive PNA index forced by the tropical Pacific SST anomalies is reduced by the negative PNA index forced by the Indian Ocean SST anomalies. For instance, one can deduce from Figs. 7a and 7b that the maximum value over the central North Pacific in ECHAM5 (CCSR) is approximately 70 m (50 m), which is about 45% (75%) of the absolute value due to the Pacific SST anomalies (Fig. 5). In the TIO runs, the pattern correlation between the two models (Table 2a) is lower than that noticed in the TPO runs but the amplitude between the models has declined substantially as revealed by the low RMSE values.

The robustness of the above result is examined in two different ways. First, we show a 500-hPa height anomalies difference plot (Fig. 8) obtained from the TIP – TPO solutions. In both AGCMs, we note that the sign of the main action centers over the PNA region are opposite to those in the TPO solutions (Fig. 5), a feature resulting from the additional rainfall anomalies over the tropical Indian Ocean. Compared to the exclusive effect of the Indian Ocean SST anomalies as deduced from the TIO solutions (Fig. 7), the amplitude of the height anomalies when the forced responses are
(a) 500hPa Height – ECHAM5

(b) 500hPa Height – CCSR

(c) 200hPa stream function – ECHAM5
interacting with each other is substantially reduced in CCSR while the positions of the centers are shifted westward by about 20° in ECHAM5.

Second, to clearly demonstrate that the extratropical response noted in Fig. 7 arises from SWIO SST anomalies, we conducted an additional experiment in which SST anomalies over the SWIO region only are prescribed in the CCSR model. The results from this solution are presented later (see Fig. 12). The amplitude and spatial distribution of the precipitation (forcing) and the 500-hPa height response from this run (Figs. 12a and 12c) bear close similarity to the results ob-

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**Fig. 7.** (a),(b) Same as in Fig. 5 but for the TIO solutions, with the same conventions as in Fig. 5. (c) The 200-hPa streamfunction anomalies from ECHAM5 from the TIO runs. Positive values are shaded progressively and negative values are shown with CI of $2.0 \times 10^{15} \text{ m}^2 \text{s}^{-2}$.  

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**Fig. 8.** Same as in Fig. 7 but for the TIP–TPO solutions. In (a),(b), the simulated major centers over the PNA region are shown with boldface signs.
tained from the TIO runs (Figs. 6b–7b). Barring some differences in the spatial orientation and magnitude of the response, the solutions from the TIO, TIP – TPO, and SWIO experiments converge to the point that, over the PNA region, the tropical Indian Ocean SST anomalies force a height response opposite to that forced by the tropical Pacific SST anomalies during boreal winter of El Niño years.

c. Response to tropical Indo-Pacific SST anomalies

Is the simulated height response over the PNA region more realistic when SST anomalies over the tropical Indo-Pacific basin are prescribed together? In Fig. 9 we show the 500-hPa height anomalies obtained from the TIP solutions, along with the composite pattern obtained from the reanalysis results (Fig. 9c). Regarding the spatial structure, there is little change between the two (TPO versus TIP) simulations. With respect to the amplitude, however, we note changes. In particular, over the central North Pacific the maximum height values in the TIP (TPO) runs are 100 (160) m and 40 (75) m in ECHAM5 and CCSR, respectively. These changes are manifested in the reduction of RMSE (Table 2). For example, between the two models, the RMSE over the North Pacific (PNA) in the TIP runs is decreased by about 32% (31%). Similarly, between the reanalysis and ECHAM5, there is a reduction of about 48% (42%) in RMSE over the North Pacific (PNA). Owing to a better simulation of the strength of the response, between CCSR and the reanalysis, the attenuation in RMSE is modest at 11% (15%) over the North Pacific (PNA). These results underscore our conjecture that for a realistic simulation of the PNA index attention needs to be paid to the inclusion of the contribution from the tropical Indian Ocean. Since the extratropical response in an AGCM is forced by the precipitation anomalies throughout the Tropics, it is not surprising to note that the strength of the height response over the PNA sector in the TIP solutions appears more realistic.

d. Potential predictability

In AGCMs the response to the boundary forcing (signal) is embedded with the internal atmospheric variability (noise). From the ensemble solutions we estimate reproducibility, a measure for detecting the model’s robustness to the imposed boundary forcing (e.g., Kumar and Hoerling 1998a). Here, the signal is the variance of the ensemble mean, and the noise is the sum of the deviation of the individual member variance from the ensemble mean variance. Reproducibility (R) is related to the F distribution for testing variances of two normally distributed populations. Here, we test the signal to noise ratio for model-simulated precipitation and 500-hPa height, and the more this ratio deviates from 1, the stronger the evidence for unequal population variances. In other words, if the noise is zero, then each realization is an exact replica and the reproducibility is infinity (Sperber and Palmer 1996).

In mathematical notation, variance of the ensemble mean (signal, $\sigma_{sig}^2$) is given by

$$\sigma_{sig}^2 = \left( \frac{1}{M \sum_{i=1}^{M} X_i} \right)^2,$$

Fig. 9. (a),(b) Same as in Fig. 7 but from the TIP solutions. (c) The composite constructed from the reanalysis results. In (a)–(c), the three major centers associated with the PNA pattern are shown with boldface signs.
where $X_i$ corresponds to the seasonal average (JFM) of the $i$th member of the ensemble of size $M$ and the noise ($\sigma^2_{\text{noise}}$) is given by

$$\sigma^2_{\text{noise}} = \frac{1}{M-1} \sum_{i=1}^{M} (X_i - \bar{X})^2. \quad (2)$$

The ratio ($\sigma^2_{\text{sig}}/\sigma^2_{\text{noise}}$) measures reproducibility ($R$).

Like in previous studies, the reproducibility ($R$) of the precipitation over the tropical Pacific is very high, as was previously discussed in Annamalai et al. (2005a). Here, we examine $R$ for the precipitation (Fig. 10) and 500-hPa height (Fig. 11) over the PNA region from the TIO simulations. For precipitation, the appealing result is that the signal exceeds the background noise by several factors over the SWIO, the main forcing region. As with the precipitation, in both AGCMs, the reproducibility of the height anomalies over the PNA region is high. These results are further supported by solutions in which SST anomalies over only the SWIO region are specified (Figs. 12b and 12d). The tropical precipitation is regarded as the principal factor in the chain of tropical–extratropical interactions (e.g., Barsugli and

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**Fig. 10.** The reproducibility ($R$) or potential predictability of precipitation in the TIO solutions. Here, $R$ is essentially the signal to noise ratio and is related to the $F$ distribution for testing variances of two normally distributed populations. The 95%, 99%, and 99.9% significant values are (a) 2.5, 3.0, and 6.0, respectively, and (b) 1.8, 2.1, and 3, respectively.
Sardeshmukh 2002). Therefore, the high reproducibility in the extratropical response can be attested to by the remarkable reproducibility of the forcing.

In all of the solutions reported above, there is considerable model-to-model variability, both in the tropical precipitation and in the extratropical response. For given identical SST forcings, compared to the CCSR solutions, the magnitude of the simulated precipitation anomalies over the entire tropical Indo-Pacific Ocean region is too strong in ECHAM5 (Figs. 4 and 6). Since the effective atmospheric forcing of teleconnection depends on the strength of the simulated precipitation anomalies, it is not surprising that the 500-hPa height anomalies in ECHAM5 are stronger by about 80% compared to the CCSR results (Figs. 5 and 7).

The simulated extratropical differences between the AGCMs can be attributed to (i) differences in the intensity of the precipitation, (ii) the detailed structure of the basic state through which poleward energy propagation from the tropical heat source occurs (Ting and Sardeshmukh 1993), (iii) the midlatitude flow’s sensitivity to the rainfall response and nonlinear interactions with transient eddies (Lau and Nath 1994), and (iv) the intrinsic dynamical behavior of the models’ extratropical flows (e.g., Palmer 1993). Since our focus is on assessing the qualitative robustness in the solutions, docu-
menting the reasons for the differences between the models is beyond the scope of the present research. Recent studies examining the extratropical response to tropical SST anomalies with multiple AGCMs have encountered such problems (Shukla et al. 2000; Hoerling and Kumar 2002).

In summary, the AGCM results presented so far suggest the following robust and important new features: (i) the local relationship between SST and precipitation over the SWIO, (ii) a substantial reduction in RMSE for the 500-hPa height response over the PNA region when the effect of Indian Ocean SST anomalies is considered along with those in the tropical Pacific, and (iii) the high reproducibility in the tropical forcing and Northern Hemisphere circulation response fields. Some minor points of interest from the TIO runs (Fig. 6) include positive rainfall anomalies along the East Asian monsoon front that are weak or absent in the TPO solutions (Fig. 4) and the reduction in precipitation over southern Africa, along the SPCZ, and over the tropical western Pacific. Watanabe and Jin (2003), Lau and Nath (2003), and Annamalai et al. (2005a) have discussed the linkage between the SWIO and the tropical western Pacific–East Asian winter monsoon front, while Goddard and Graham (1999) have highlighted the relationship between SWIO SST and southern African rainfall. The TIO solutions produce positive precipitation anomalies over the eastern equatorial Indian Ocean, where the observations are negative, implying the role of remote forcing from the tropical Pacific in determining the precipitation anomalies there (Fig. 4).

4. Results from the LBM

It is not the intention of this section to see how closely a linear model captures either the observed or AGCM results. On the other hand, based on the results reported in section 3, we seek LBM solutions to help answer the following questions: (i) Is the dynamics involved in the tropical–extratropical teleconnection different between the tropical Pacific and Indian Ocean heating? (ii) To what degree do the structure and intensity of the response over the PNA region to Indian Ocean heating depend on the basic flow? (iii) Does the strength of the midlatitude response depend on the latitudinal position of the heating over the tropical Indian Ocean?

The LBM is forced by the column-integrated heating profiles shown in Figs. 13a and 13c. To represent the
positive precipitation anomalies over the Indian Ocean (tropical Pacific Ocean), we impose diabatic heating anomalies centered at $7^\circ$S, $60^\circ$E ($0^\circ$, $160^\circ$W) with a heating rate of $2$ K day$^{-1}$ ($4$ K day$^{-1}$). The zonal and meridional extents of the heating anomalies over the Indian (Pacific) Ocean are $40^\circ$ and $20^\circ$ ($100^\circ$, $15^\circ$), respectively. To represent top-heavy heating, the maximum heating is located at $400$ hPa. The vertical heating profile used is similar to that of Reed and Recker (1971) and has been employed in many previous studies (e.g., Jin and Hoskins 1995). The horizontal shape of the heating is elliptical. The position and orientation of the imposed heating anomalies closely "mimic" the simulated positive precipitation anomalies in the ECHAM5 solutions (Figs. 4 and 6). To validate the CCSR solutions, similar experimental designs, but specifying the Indian Ocean heating at $18^\circ$S, $60^\circ$E, are performed (section 4b). To address the questions posed above, LBM solutions to a variety of basic flows problems are obtained. In all cases, the LBM is integrated for 30 days, and, with the dissipation terms adopted, the tropical response approaches a steady state after day 10 and the extratropical response is attained within 2 weeks, which is consistent with the results of Jin and Hoskins (1995). The response at day 20 is analyzed here.

### a. Response to realistic ambient flow

First, to better understand the similarities and differences in the dynamics involved in setting-up the teleconnection between Pacific and Indian Ocean heating, we present results from experiments carried out with zonally varying ambient flow from the reanalysis data. This approach is justified since both of the AGCMs, despite differences in their basic state (Fig. 3), yielded consistent results to Pacific and Indian Ocean SST anomalies, at least qualitatively.

#### 1) Extratropical response

Figure 13b shows the 500-hPa height response for the tropical Pacific heating. In agreement with many pre-
vious idealized simple modeling studies, our LBM solutions capture the classical PNA pattern. The results based on ECHAM5 basic flow are not discussed further because they are nearly the same as the presented results from the reanalysis basic state. The height response to Indian Ocean heating (Fig. 13d), in particular over the PNA region, has opposite polarity to that due to central Pacific heating, except over southeastern United States. The LBM solutions corroborate the results from AGCMs and confirm the results reported in many previous linear model studies (Simmons et al. 1983; Ting and Sardeshmukh 1993; Jin and Hoskins 1995). In an AGCM, the midlatitude response depends on direct tropical forcing and the interaction between quasi-stationary anomalies and storm tracks. In the LBM, however, the response is entirely due to direct tropical forcing. Therefore, compared to the AGCM solutions (Fig. 7), the amplitude of the 500-hPa height response over the PNA region forced by Indian Ocean heating (Fig. 13d) is much weaker. Because of the linearity, the absolute, but not relative, magnitude of the LBM response is the primary concern in this section.

The qualitative similarity between the AGCM (Fig. 7) and LBM (Fig. 13d) solutions is noteworthy, and suggests the importance of direct tropical forcing for the extratropical response.

2) Dynamics

The specified heating over the SWIO leads to local ascent and upper-tropospheric divergence, and subsequent generation and propagation of Rossby waves. Sardeshmukh and Hoskins (1988) emphasized that the advection of vorticity by the divergent wind from the source region leads to a Rossby wave source in westerly wind regions in which Rossby wave motion is effective. Ting and Yu (1998), however, found that the divergence term dominates the total Rossby wave source. Therefore, we examine the temporal evolution of the 200-hPa divergence anomalies from the Indian Ocean heating (Fig. 14).

The heating is switched on at $t = 0$; within a day (Fig. 14a), the heating leads to in situ upper-tropospheric divergence. The convergence anomalies on the poleward flanks of the divergence are suggestive of changes
in the local Hadley circulation. Similarly, the convergence to the east of the divergence implies modulations to the local Walker circulation. It should be mentioned here that the prescribed heating in the Indian Ocean lies over upper-level easterlies (Fig. 3a). The subtropical anomalous convergence associated with the descending branch of the local Hadley circulation results in an effective Rossby wave source due to the prevalence of mean westerlies there. A significant change occurs between day 1 and day 2. By day 2 (Fig. 14b), the anomalies strengthen in magnitude and expand spatially except in the southern descending branch of the local Hadley circulation. A new divergence anomaly center is noticeable around 28°N, 100°–130°E. This subtropical divergence is found to the northeast of the heating and is viewed as a sign of the Rossby wave propagation rather than as a part of the forcing. In other words, it just takes 2 days from the switch-on for the emanation and propagation of Rossby waves, leading to divergence anomalies elsewhere.

Further evidence for the intensification of the divergence–convergence anomalies and propagation of Rossby waves into and through middle latitudes in the first week is seen at $t = 7$ days (Fig. 14c). Within 2 weeks (Fig. 14d), the wave trains lead to a response at middle and high latitudes, as far away as Greenland in the Northern Hemisphere. Consistent with the results of Jin and Hoskins (1995) and Ting and Sardeshmukh (1993), our solutions suggest that the diabatically forced wave train is involved in setting up the teleconnection noted in Fig. 13d. One of the new results here is that the tropical heating anomalies produce substantial midlatitude divergence anomalies, which may result in the generation of Rossby wave sources–sinks far away from the source region. We examined the 200-hPa divergence anomalies from the TIO runs of the AGCMs (figure not shown). Barring differences in the magnitude and longitudinal orientation of the anomalies, the primary features obtained in the LBM response are evident in the AGCMs too, implying similar processes operating in both the LBM and AGCMs in setting up the teleconnection. Another new result from the LBM and discussed below is the difference in the propagation path of Rossby waves for the two heating patterns (Figs. 13a and 13c).

To better diagnose the wave propagation, we show the 200-hPa meridional component of the wind (Fig. 15). In agreement with the height responses, the 200-hPa meridional wind components due to the two heating patterns oppose each other over the PNA region. The central Pacific heating has large signatures in the deep Tropics of both hemispheres. The Indian Ocean heating, on the other hand, leads to a stronger response in the Northern Hemisphere extratropics.

It should be noted that the solution to the Indian Ocean heating is quite different from the PNA-like response to the Pacific heating in terms of the wave structure. For example, the wave packets shown in Fig. 15b do not follow the great circle, but rather extend eastward along the latitudinal circle. This feature is reminiscent of the so-called waveguide pattern resulting from trapping of Rossby waves by the strong jet curvature (Hsu and Lin 1992). It is known that if there is a maximum in the effective beta (thick solid lines in Fig. 15b) that is equivalent to the meridional gradient in the basic-state potential vorticity, then Rossby waves tend to be refracted toward that maximum (cf. Hoskins and Ambrizzi 1993). This is indeed the case for the winter Asian jet extending from southern Europe to the western North Pacific. Consequently, the steady response (Fig. 15b) tends to extend along the longitudinal direction. The energy propagation in the waveguide is rapid enough for the waves to retain appreciable amplitude away from the source region. This may possibly explain why the Indian Ocean SST anomalies have a large impact on the far downstream, that is, over the PNA region (section 3b). While the waveguide teleconnection patterns during boreal winter have been identified in association with the North Atlantic Oscillation (Branstator 2002; Watanabe 2004; Hoerling et al. 2004), the present study indicates that SWIO heating anomalies can be an alternate source for effectively exciting them over the PNA region, a feature not discussed previously. As in the case of the conventional PNA response (Held et al. 1989), the midlatitude storm tracks may also play a role in enhancing the North Pacific response to Indian Ocean SST anomalies, an issue beyond the scope of the present study.

b. Sensitivity to the basic flow

1) ZONAL-MEAN FLOW

The sensitivity of the above results to the zonally varying flow, in particular that waves excited by the SWIO heating can retain substantial amplitude far away from the source region, is tested by performing an additional run in which the same heating pattern is used to force the LBM with a zonal-mean state (Fig. 13c). The basic state includes both zonal-mean zonal and meridional winds. The 500-hPa height and 200-hPa meridional wind responses are shown in Figs. 16a and 16b. Compared to the solutions in Fig. 13d, the amplitude of the height anomalies over the PNA region is negligible (shading–contour interval in Fig. 16a is one-half of that
in Fig. 13d) and even the sign of the anomalies over northwest America is opposite. Similarly, the perturbation upper-level meridional wind does not reveal the Rossby waves trapped by the subtropical jet (Fig. 16b), implying that the regionally intensified Asian jet is responsible for the waveguide teleconnection by the SWIO heating as found in Fig. 15.

2) CCSR AGCM BASIC STATE

Relative to the reanalysis results, the basic state differs considerably in CCSR (Fig. 17a), and the latitude of the simulated Indian Ocean precipitation anomalies lies at about 18°S (Fig. 6b)—too far south compared to the observations (Fig. 1b). It is known that errors in the model climatology can affect the normal modes of the flow. Here, to understand the sensitivity of the extratropical response to the AGCM basic state and the forcing location, the Indian Ocean heating experiments are repeated by specifying the heating anomalies at 7°S, 60°E and 18°S, 60°E. In these setups, the governing equations are linearized around the zonally varying basic flow from the reanalysis data.

![Diagram](image)

**Fig. 15. Perturbation meridional wind (V) response at 200 hPa for (a) the central Pacific heating pattern shown in Fig. 13a and (b) the Indian Ocean heating pattern shown in Fig. 13b. Positive (negative) values are shaded progressively (shown as dashed contours). The CI is 1.0 m s⁻¹, and the 0 contour is not shown. Also shown in thick solid lines in (b) is the effective beta for the climatological zonal wind at 200 hPa. The contour interval for the effective beta is 2.0, and the 0 contour is suppressed. The governing equations are linearized around the zonally varying basic flow from the reanalysis data.**
compared with those from earlier experiments (Figs. 7b, 13d, and 15b).

The LBM solutions for forcing placed at 18°S, 60°E indicate that the spatial structure of the 500-hPa height response over the PNA region (Fig. 17b) remains similar to that obtained from the AGCM simulations (Fig. 7b). As regards the Rossby wave signature (Fig. 17c), pronounced wave activity refracting toward the effective beta maximum is clearly evident in the Northern Hemisphere. Owing to the stronger westerlies around 30°N in CCSR (Fig. 17a), the effective beta maximum is displaced 10° southward compared to the reanalysis results (Fig. 15b). Interestingly, the 500-hPa height response to heating placed at 7°S, 60°E (Fig. 17d) is nearly identical to that obtained by fixing the heating at 18°S, 60°E (Fig. 17b). In both solutions here, the signals over the PNA region extending to Greenland bear close similarity to that shown in Fig. 15b; however, there is a longitudinal shift in the position of the anomalies, which is not due to the location of the forc-
ing but rather to the differences in the basic flow. As expected, for the heating placed at 18°S, wave activity is noticeable over the Southern Hemisphere but confined to the longitudes of 0°–180°. In comparing the responses between the two hemispheres, due to the stronger westerly jets, the Rossby wave signature is predominant in the Northern Hemisphere extratropics.

In summary, our LBM solutions confirm the earlier results reported in the literature. Some new findings include (a) identification of the differences in the propagation path of the Rossby waves between Pacific and Indian Ocean heating that could be associated with the aspect ratio of the heating associated with the two regions (B. J. Hoskins 2006, personal communication), (b) the insensitivity of the midlatitude response to the forcing latitude in the tropical Indian Ocean, and (c) the role of zonally varying ambient flow in preserving the downstream response to Indian Ocean heating.

5. Summary and discussion

In this section, first we summarize and highlight the new results obtained in the present study. Then, we present possible evidences for the role of Indian Ocean SST anomalies on local and remote climate variabilities, and argue the need for better monitoring of the regional SST anomalies. We close the section by highlighting some areas of future study that could help to further substantiate the impact of Indian Ocean SST anomalies on the PNA response.

a. Summary

Ocean dynamics have been shown to affect SST variations over the SWIO, and these, in turn, impact local precipitation anomalies. These variations may also affect the Pacific–North American (PNA) circulation during boreal winter (JFM +1) of El Niño years. In
this study, we investigated this hypothesis in ensemble simulations with two AGCMs—ECHAM5 and CCSR—differing in numerics and physical parameterizations. The linear version of the CCSR AGCM is used to identify the basic processes.

When forced with El Niño–related SST anomalies, which are confined to the tropical Pacific, both AGCMs capture the well-known dipole precipitation anomalies along the equatorial Pacific (Fig. 4). Furthermore, both AGCMs show that the anomalous atmospheric heating caused by these SST anomalies force the familiar PNA pattern. A close examination with reanalysis data, however, reveals that the spatial pattern of the 500-hPa height anomalies is better captured in ECHAM5 while the amplitude is closer in CCSR (Table 2). However, both models fail to simulate the observed positive precipitation anomalies (Fig. 1b) over the SWIO.

When forced with only SST anomalies in the Indian Ocean, the models produce the observed precipitation anomalies over the SWIO (Fig. 6). These in turn excite wave trains of large amplitude over the northeastern Pacific (Fig. 7). The spatial pattern of the circulation response suggests a diabatically forced wave train that originates in the Indian Ocean (Fig. 7c). An important consequence for the PNA region is that the 500-hPa height anomalies forced by Indian Ocean SST anomalies oppose and destructively interfere with those forced by tropical Pacific SST anomalies (Fig. 5).

The positive PNA index forced by the tropical Pacific SST anomalies is reduced by the negative PNA index forced by Indian Ocean SST anomalies. The key result from our model solution is that when the effects of both the Indian Ocean and the Pacific SST anomalies are considered together, the RMSEs of the 500-hPa height responses over the North Pacific and the PNA regions are significantly (15%–42%) reduced. There are large quantitative differences in the intensity of the simulated response between the two AGCMs, but the qualitative robustness between them, however, is reassuring. Despite the differences, the success in reproducing the SWIO and 500-hPa height response over the PNA region across the ensemble members indicates some degree of robustness.

Solutions with the linear model show that Rossby waves emanate from the Indian Ocean into and through the middle and high latitudes (Figs. 12–15). We identify that the wave packets forced by Indian Ocean heating do not follow the great circle path, but extend along the waveguide and refract toward the effective beta associated with the zonally varying flow (Figs. 15 and 17). Sensitivity experiments reveal that a realistic ambient flow affects the downstream response to Indian Ocean heating (Fig. 16). We also note that tropical heating leads to the creation of midlatitude divergence anomalies. Additional experiments with the CCSR basic flow suggest that the midlatitude response is not sensitive to the forcing latitude in the Indian Ocean. Based on the results from the idealized experiments conducted here, we conclude that a description of the extratropical response to forcing by tropical SST anomalies must take into account the contribution from the tropical Indian Ocean.

b. Discussion

Considerable progress in seasonal forecasting has been made in the last decade following the realization that the tropical Pacific SST anomalies associated with ENSO are significantly correlated with atmospheric anomalies over the middle latitudes in spite of the presence of large natural variability. The research described here shows that El Niño–induced SST anomalies in the SWIO also affect the PNA response.

Thus, although the spatial expanses of the SST and heating anomalies in the tropical Indian Ocean are much smaller than those in the tropical Pacific, our model solutions show a substantial reduction in the RMSE in 500-hPa height over the PNA region when the SST anomalies from the Indian Ocean are specified along with those in the Pacific. This effect of the Indian Ocean SST anomalies needs to be further verified and quantified from other AGCMs. Observations show that the PNA response was rather weak during the winter of 1972/73, despite strong SST anomalies in the equatorial Pacific forcing (Shukla and Wallace 1983; Kumar et al. 1996). Based on the present results and those of others, we need to explore if the Indian Ocean SST anomalies counteracted the effect of the Pacific anomalies in 1972/73.

Hoerling et al. (2004), for example, noted that the PNA index and North Atlantic Oscillation index are sensitive to Indian Ocean SST anomalies. They even attributed the observed linear trend seen in the North Atlantic Oscillation to the trend in Indian Ocean SST anomalies during 1950–99. In an earlier study (Xie et al. 2002), we showed that SWIO SST anomalies influence the number of tropical cyclone days in the southern Indian Ocean. A subsequent study (Annamalai et al. 2005a) showed that these SST anomalies affect the intensity of tropical western Pacific convection and the East Asian winter monsoon, as well as the onset of the Indian summer monsoon. In the present study, we recognize its importance upon the PNA region. In summary, we argue that SWIO SST may offer a source of local and remote climate predictability, and therefore, we suggest that further studies of the dynamics of In-
dian Ocean, as well as accurate measurements of SST variations there, are needed.

The relationship between the Indian Ocean and the equatorial Pacific SST anomalies, however, is more complicated than implied above. Annamalai et al. (2005b) presented dynamical evidence that the Indian Ocean SST anomalies strengthen (weaken) the ongoing El Niño in the post- (pre-) 1982 period. Kumar and Hoerling (1998a) indicated that the seasonal predictability over the PNA region was lower during 1959–65 than during the post-1982 period. They attributed this difference to the intensity of El Niño events. The Indian Ocean, therefore, may affect the intensity of the El Niño more during some periods and less during others (Annamalai et al. 2005b). In other words, Indian Ocean SST in boreal fall may affect the intensity of El Niño events, which then feedback to strengthen or weaken the Indian Ocean SST anomalies in boreal winter. To understand this two-way interaction between the tropical Pacific and Indian Oceans and to accurately simulate the effects of tropical SST anomalies, fully coupled ocean–atmosphere models are needed. Furthermore, the present and other recent studies (e.g., the review in Annamalai and Murtugudde 2004) imply that monitoring Indian Ocean circulation is necessary in order to make better seasonal predictions during El Niño years not only of the regional atmospheric response, but also the midlatitude response.

In a future study, we plan to use more ensemble members and to include SST anomalies in the tropical Atlantic in order to determine the predictability of the Northern Hemisphere extratropical circulation during El Niño years. In particular, we intend to assess the contribution of the internal dynamics, as in the studies of Straus and Shukla (2002) and Peng and Kumar (2005), and focus on the clear distinctness in the spatial patterns due to SST-forced versus internal dynamics, as pointed out by Straus and Shukla (2002). We plan to address the possible reasons for the “large differences” in the intensity of the simulated response between the two AGCMs noted here, together with issues regarding the circulation response in the Southern Hemisphere to Indian Ocean heating.

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REFERENCES


Peng, P., and A. Kumar, 2005: A large ensemble analysis of the
influence of tropical SSTs on seasonal atmospheric variability. J. Climate, 18, 1068–1085.


