Atmospheric Response Patterns Associated with Tropical Forcing

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ABSTRACT

Atmospheric response patterns associated with tropical forcing are examined with general circulation models driven by global sea surface temperature (SST) variations during 1950–99. Specifically the sensitivity of mid-latitude responses to the magnitude and position of tropical SST anomalies is explored. This controversial problem, spanning more than a quarter century now, centers on whether response patterns over the Pacific–North American region are affected or changed by inter–El Niño variability in tropical forcing. Ensemble methods are used in this study to reliably identify the signals related to various tropical SST forcings, and the sensitivity is determined from analysis of four different climate models.

First, the fraction of Pacific–North American (PNA) wintertime 500-hPa height variability that is potentially predictable and is linked to interannual variations in the global SSTs is identified. This SST-forced component accounts for as much as 20%–30% of the total seasonal mean height variability over portions of the PNA region, and the most important boundary forcing originates from the tropical Pacific Ocean. The spatial expression of the teleconnections that are linked to this potentially predictable SST-forced fraction of height variability is next identified. The leading model pattern is similar to the classic observed teleconnection associated with the linear ENSO signal, and explains 80% of the SST-forced height variance over parts of the North Pacific and North America. Two additional wavelike patterns are identified that are also associated with tropical forcing. One is related to the pattern of tropical SST variations often seen during the transition of the tropical ocean that marks the interlude between ENSO extremes, and the pattern of forcing related to it is distinctly non-ENSO in character. The other is related to the nonlinear component of the atmospheric response to ENSO’s extreme opposite phases. Response patterns having annular-like structures over the Northern Hemisphere that are related to multidecadal variations in tropical Indo-Pacific and Atlantic SSTs are also highlighted.

Subtle modifications in upper-level responses to different tropical SST forcings are shown to yield disproportionate sensitivity in North American surface climate. Particularly pronounced is the reversal in sign of the precipitation anomalies over the region spanning the Canadian border to southern California in response to equatorial Pacific convection anomalies shifting from 170°E to 140°W. The behavior is reproduced in experiments using both realistic and idealized SST anomalies, and this behavior is found to emerge particularly when the far eastern equatorial Pacific Ocean is strongly warmed as occurred during the 1982/83 and 1997/98 El Niños.

Despite the existence of different response patterns to tropical SST forcings, it is shown that the seasonal hindcast skill of PNA 500-hPa heights for 1950–99 originates mainly from the single, leading teleconnection structure. The conclusion drawn from this result is that the atmospheric sensitivity to different tropical SST forcings, though real, is weak and easily masked by the year-to-year climate variations due to internal atmospheric processes.

1. Introduction

The existence of a characteristic atmospheric teleconnection forced by changes in tropical SSTs associated with El Niño–Southern Oscillation (ENSO) is well known. The observational evidence for such a pattern, first suggested by Bjerknes (1966) in his case study of the 1957/58 El Niño winter, rests on correlation analysis between tropical ENSO indices and geopotential height, such as conducted for 1951–78 in Horel and Wallace (1981). Figure 1 updates their analysis through 1999 for 500-hPa heights. The question remains open whether this linear signal constitutes the sole tropically forced teleconnection, or whether there exist additional response patterns to so-called different flavors of tropical SST forcing associated with individual ENSO events, or to some other non-ENSO SST forcing. Understanding the atmospheric response to tropical forcing beyond the linear ENSO signal is an outstanding problem in clarifying the sources and predictability of climate variations on seasonal to centennial timescales.

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Observations suggest that different response patterns occur during warm versus cold phases of ENSO, apparently related to the different patterns of tropical rainfall anomalies that distinguish ENSO’s extreme phases (Mon- troy et al. 1998; Hoerling et al. 1997, 2001c). However, there are simply too few observed cases from which to determine whether extratropical circulations are sensitive to the changes in tropical SSTs from one event to another. To be sure, observed seasonal anomalies during different El Niños are often distinct from each other, though contrary interpretations have been offered. Madden (1976) estimated the variability of observed monthly sea level pressure due to the internal atmospheric variations attributable to daily weather fluctuations. His comparison of this so-called natural variability to the total variability revealed only small differences in the midlatitudes implying that the aforementioned inter–El Niño variations in climate anomalies are consistent with “weather noise” and are not due to sensitivity to boundary forcing. This argument has been supported by results from general circulation models (GCMs; Geisler et al. 1985; Kumar and Hoerling 1997) that find only a weak extratropical sensitivity to changes in tropical Pacific SSTs from event to event. Yet, the opposite conclusion, drawn from other GCM runs, is that the inter–El Niño differences in extratropical seasonal anomalies can be explained by SST variations (Palmer and Owen 1986) supporting the contention that a substantial signal exists in the extratropics, but which varies from one ENSO event to another (Trenberth 1993).

Even less is known from observations about the impact of non-ENSO SST forcing. An outstanding question is whether the fraction of extratropical variability that can be explained by tropical SSTs as a whole differs appreciably from the ENSO fraction of Fig. 1. Case studies for individual winters suggest that tropical forcing from outside of the equatorial central Pacific can be important for explaining climate anomalies in the extratropics (e.g., Palmer and Owen 1986; Hoskins and Sardeshmukh 1987; Kumar et al. 2001). Recent GCM hindcasts forced by observed global SSTs (Shukla et al. 2000) show high skill scores for Pacific–North American 500-hPa heights during several non-ENSO winters. Though skill for individual seasons can be high due to chance alone, the fact is that the ensemble average of these experiments exhibit a significant response to the prescribed global SSTs indicative of a boundary-forced signal. Precisely which SSTs are responsible has yet to be determined.

Owing to the brief instrumental records, it has been difficult to draw conclusions on the existence of extratropical atmospheric signals beyond the ENSO pattern of Fig. 1 from empirical studies alone. GCM methods can in principle overcome these observational limitations, yet the few efforts to date on this specific problem have yielded conflicting results as mentioned above, perhaps due to the limited ensemble sizes used, the realism of the SSTs employed, or the realism of the GCMs themselves. A complementary approach has been to pursue this question with simpler dynamical models. When driven by specified structures of tropical diabatic heating, these confirm the fundamental tropical origin for the teleconnection in Fig. 1 (e.g., Webster 1981; Hoskins and Karoly 1981; Simmons 1982). They also suggest that, owing to sources of energy related to the detailed structure of the three-dimensional atmospheric circulation, the extratropical responses to tropical forcing should be geographically fixed (e.g., Simmons et al. 1983; Branstator 1990). Indeed, the localized east Pacific variance maximum associated with the linear ENSO teleconnection is consistent with the presence of a spatially fixed low-frequency perturbation energy source located downstream of the east Asian jet.

These results do not rule out the importance of differences among tropical forcing patterns, and indeed knowledge of that forcing in space and time is required for a complete explanation of the distribution of extratropical low-frequency variability (Borges and Sardeshmukh 1995; Newman et al. 1997). For seasonal timescales, both linear time-independent models (e.g., Ting and Sardeshmukh 1993) and nonlinear time-varying models (e.g., Ting and Yu 1998) predict a sensitivity in extratropical circulation patterns to the different longitudes of equatorial diabatic forcing. To be sure, the response patterns in such models do not shift as much as the forcing (consistent with the role of the climatological stationary waves), but the stationary wave sensitivity appears sufficiently strong to be of practical im-

Fig. 1. The linear correlation between observed winter seasonally averaged (DJF) 500-hPa heights and an SST index of ENSO for the period 1950–99. The index, shown in Fig. 2, is based on the PC time series of the leading EOF of tropical Pacific SST variability. Solid contours (dashed) denote positive (negative) correlations. Contour interval is 0.1. Local correlations are statistically different from zero at the 95% level for values larger than 0.3.
portance for predicting sensible weather elements such as North American rainfall and temperature. It remains to be determined if such behavior in the simple models exists in more realistic dynamical systems, in particular ones that include storm track feedbacks related to transient disturbances that are known to be important for the extratropical ENSO response (Kok and Opsteegh 1985; Held et al. 1989; Hoerling and Ting 1994).

Progress on this matter may also clarify the origin of atmospheric variations on longer than interannual timescales. It has been suggested, for example, that decadal–centennial climate variations have a tropical root in light of interhemispheric symmetries in paleoclimate change (e.g., Evans et al. 2001). For the period of the twentieth century alone, GCM simulations suggest that the Tropics have forced multidecadal changes in atmospheric circulation over the Pacific–North American (PNA) sector (Graham 1995; Kumar et al. 1994), as well over the North Atlantic–European sector (Hoerling et al. 2001a). One relevant forcing appears to be a tropical-wide sea surface warming since 1950 that has increased equatorial Indo-Pacific rainfall (e.g., Hoerling et al. 2001a). ENSO is also believed to be relevant on these long timescales since its associated SST variations possess power across a wide frequency range (e.g., Cane and Zebiak 1985). However, equatorial Pacific SST changes since 1950 project only modestly on the ENSO structure (e.g., Knutson and Manabe 1998), and it is again unclear whether the expected change in extratropical circulation should resemble the ENSO teleconnection pattern of Fig. 1.

Assessing the tropical forcing of global climate variations, regardless of their timescale, requires that the atmospheric sensitivity be better understood. We are especially interested to learn whether the fraction of seasonal variance explained by tropical SSTs may be greater than implied by the linear ENSO correlations of Fig. 1. The practical importance of this problem for extratropical seasonal weather forecasting is clear. Currently, ENSO is the key source of seasonal prediction skill over the United States, particularly during winter (e.g., Barnston 1994). If atmospheric response patterns other than those based on this linear ENSO signal exist as implied by some studies, then the challenge is to identify the responsible boundary conditions, and evaluate the prospects for improved seasonal predictions.

Following a description of the datasets in section 2, we review some principal features of the observed total winter season variance in tropical SST, tropical rainfall, and extratropical 500-hPa heights in section 3. We estimate the fraction of that variance that can be explained by an index of ENSO during 1950–99. Section 3 presents a parallel analysis of atmospheric GCM simulations for 1950–99, and an intercomparison with observations is performed in order to establish the realism of the model behavior. The sensitivity of atmospheric response patterns to different seasonal mean tropical forcings is diagnosed in section 4, an analysis that is only feasible from the GCM data and their large ensembles spanning the last half century. Identified also are the 500-hPa response patterns associated with specific ENSO events in order to clarify the importance of inter–El Niño SST variations. Section 5 provides a summary and concluding remarks.

2. Data and methodology

a. Observations

Global sea surface temperatures are analyzed for the period 1950–99. Since 1982, this data is based on in situ observations and satellite estimates, combined using the optimum interpolation method described in Reynolds and Smith (1994). For the period 1950–81, in situ observations are used together with an eigenvector reconstruction as described in Smith et al. (1996). The resulting monthly SST data are analyzed at 2° resolution, and are subjected to an empirical orthogonal function (EOF) analysis in order to derive an objective index of the state of the tropical Pacific Ocean. The EOF analysis is of the covariance matrix that consists of 600 monthly time samples of anomalous SSTs during the period. The analysis is conducted over the region 20°N–20°S, 120°E–60°W, and the first EOF (EOF1), shown in Fig. 2, explains 45% of the total SST variance over this domain. As described further in Hoerling et al. (2001b), this particular index is found to correlate highly with other widely used indices of ENSO.

Observed seasonally averaged anomalies for winter [December–January–February (DJF)] 500-hPa height, rainfall, and SST are used to calculate the total seasonal variance of each field. The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996) 500-hPa monthly heights for 1950–99 are avail-
available on a 2.5° grid. Satellite estimates of tropical rainfall are based on the Global Precipitation Climatology Project (GPCP; Huffman et al. 1997) dataset, and cover the period 1980–99.

The ENSO fraction of the total variance for each of the above fields is also estimated. First, the seasonal anomalies are regressed upon the EOF1 index given by the time series in Fig. 2. Because the atmospheric responses for warm and cold phases of ENSO are not equal and opposite, we calculate separate regressions for the positive and negative phases of this index following Hoerling et al. (2001c). The ENSO component of seasonal anomalies for each winter is then computed from the scalar product between the EOF1 index value and the appropriate regression map. The ENSO variance calculated from these anomalies includes both the linear and an estimate for the nonlinear component of the ENSO signal. It should be noted that owing to this nonlinear part, a covariance can exist between the ENSO and non-ENSO components of seasonal anomalies. As such, the total variance will not necessarily equate to the sum of variances due to the ENSO component and the residual component associated with all other processes.

b. Atmospheric GCM experiments forced by realistic SSTs

Monthly mean global SSTs for the 1950–99 period are imposed as evolving lower boundary forcing for a suite of atmospheric GCM simulations. Four different models, whose characteristics are summarized in Table 1, are employed, and 46 realizations of the last half of the twentieth century have been generated. A total of 12 simulations using NCAR’s Community Climate Model (CCM3; Kiehl et al. 1998), 12 using NCEP’s Medium-Range Forecast model (MRF9; Kumar et al. 1996), 10 with the European Centre-Hamburg model (ECHAM3; Roeckner et al. 1992), and 4 simulations using the climate model of the Geophysical Fluids Dynamics Laboratory (GFDL; Broccoli and Manabe 1992) have been forced with the monthly evolving global SST boundary conditions. Two other types of four-member GFDL runs are included, one in which only the tropical Pacific SSTs (30°N–30°S) vary while climatological SSTs are prescribed elsewhere, and a second also forced by tropical Pacific SSTs but with a 50-m slab mixed layer in the remaining oceans. The ensemble members of a particular model differ from each other only in the specification of the atmospheric initial conditions, and runs begin on January 1950 using initial atmospheric states selected from control integrations of the respective GCM.

All the models employed are global spectral, and use sigma coordinates for vertical discretization. Table 1 summarizes the spectral truncations of the GCMs, and the effective horizontal resolution for all the models is roughly 3° latitude/longitude. Different parameterizations of moist physics are used in the various GCMs, and the reader is referred to the cited references in Table 1 for further details of the dynamical formulations and parameterized physics employed.

Our purpose in the GCM analysis is to provide an assessment of the variability and response to tropical forcing based on a “grand GCM” average of the four individual models, rather than to highlight the performance of any single model. As such, the GCM results are based on an aggregation of the individual models, with the combination occuring on a common 128 × 64 Gaussian grid. Further details on the specific method of how the four model datasets are combined are given in sections 3b and 4a.

The reader is referred to Hoerling et al. (2001c) for an intercomparison of these four GCMs regarding each model’s sensitivity to ENSO’s extreme and opposite phases. In brief, each model generates almost the same spatial pattern of atmospheric response to an index of ENSO that correlates highly with the observed pattern of Fig. 1. There are nonetheless appreciable model differences in their ENSO sensitivity, and in particular, the amplitude of responses is model dependent. For example, ECHAM3 exhibits the largest extratropical response to ENSO whereas CCM3 exhibits the weakest signal. The sensitivity of the GFDL model is described in more detail in the review article of Alexander et al. (2002) found in this issue.

c. Atmospheric GCM experiments forced by idealized SSTs

For one of the GCMs (MRF9), additional sets of experiments are conducted using idealized tropical SST forcing. These are designed to study the sensitivity in atmospheric response patterns to various equatorial locations of SST forcing, sampling the range of known inter–El Niño SST variability in addition to forcing from the west Pacific and Indian Oceans. The model is run in perpetual January mode, as opposed to the seasonal cycle simulations described above, and the forcing consists of a monopole warm SST anomaly whose spatial pattern is similar to EOF1 (Fig. 2). It possesses a max-
imum amplitude of +2°C centered on the equator, with a zonal scale of 70° longitude and a meridional scale of 20° latitude. Sea surface temperatures over the remaining global oceans are assigned climatological January values. A 24-month integration is performed for each position of the SST anomaly, and 12 sets of experiments are conducted in which the SST pattern is displaced westward by roughly 15° longitude increments, beginning with the position that mimics the composite El Niño location of Fig. 2, and ending with the maximum SST anomaly located in the Indian Ocean.

3. Seasonal wintertime variances and the role of ENSO

a. Observations

It is evident from the EOF analysis of Fig. 2 that ENSO is the leading source of tropical Pacific interannual SST variability. This is further quantified in Fig. 3 in which the total variance of wintertime SSTs across the entire Tropics is compared to the ENSO component alone. The two variances (top and middle) are virtually indistinguishable east of the date line and within 10° of the equator. Also, half of the total SST variance is ENSO related in the equatorial central Indian Ocean (bottom), consistent with a tropical “atmospheric bridge” that was evident in Pan and Oort (1983) and is reviewed in Alexander et al. (2002). By contrast, ENSO accounts for much less than half of the total interannual SST variance in the equatorial western Pacific and Atlantic basins.

Whereas the equatorial eastern Pacific experiences the strongest interannual SST variance, the central Pacific witnesses the strongest variance in rainfall (Fig. 4, top). Here too, ENSO is the principal source of that variability (Fig. 4, middle), and likewise it exerts a strong control on wintertime rainfall in vicinity of the Maritime Continent and South America. ENSO explains little of the rainfall variance over that the central Indian Ocean however, despite explaining half of the SST variance over that region. As such, changes in atmospheric or oceanic circulation appear to be responsible for the Indian Ocean SST changes during ENSO, and these exert little feedback onto the local atmospheric convection.

The central question is the extent to which the variations in tropical forcing drive extratropical circulation, and in particular what is the relationship between such forcing and the seasonal variance of 500-hPa heights? Shown in Fig. 5 is the total variance of the Northern Hemisphere (NH) heights (left), the principal features of which are the oceanic maxima over the eastern North Pacific and North Atlantic Oceans. ENSO is an important contributor to the North Pacific maxima (middle) explaining over one-third of the total wintertime height variance in that area (right). The individual maxima in the ratio of variance trace out the teleconnection pattern associated with ENSO, and is similar to that shown in Fig. 1 using correlations.

ENSO also explains a significant fraction of local climate variability within the wintertime Southern Hemisphere (SH), as illustrated in Fig. 6 for the July–August–September (JAS) season when such teleconnections are strong (e.g., Karoly 1989). As for the NH,
the ENSO variance is largest over the central Pacific ocean at a longitude that is immediately poleward of the equatorial rainfall anomalies (not shown). There is little indication for additional poleward and downstream propagation of an ENSO signal, in contrast to the situation in the boreal winter NH.

b. GCM

Shown in Fig. 7 is total seasonal variance of GCM rainfall and the component associated with ENSO. Consistent with observations, the simulated interannual variations in wintertime rainfall are largest over the central equatorial Pacific, a region where the models’ ENSO signal dominates. An ENSO component is also evident in the far western Pacific and over South America that is associated with a Tropics-wide teleconnection pattern. The ENSO signal in the western Pacific is primarily occurring over oceanic areas, whereas the land masses including Indonesia, Borneo, and New Guinea experience little wintertime ENSO disruption in rainfall. A similar situation is evident in the observations, though the simulated ENSO signal over the oceanic regions of the western Pacific is weaker than observed. While this may reflect a GCM bias, a study of historical station data since 1900 reveals considerable decadal variability in ENSO-related rainfall signatures over the Tropics (Diaz et al. 2001), and as such a sampling bias may also exist in our satellite rainfall analysis of only the last 20-yr period.

The wintertime NH variance of 500-hPa heights (Fig. 8) is realistically simulated (cf. with Fig. 5), both with respect to the amplitude and spatial structure of the total variance. Importantly, the model confirms that the principal location of ENSO contribution to the extratropical seasonal height variance occurs over the PNA region. In this area, the observational estimate of the variance explained by ENSO was roughly 30%-40% within the PNA centers of action, and the analysis of the much larger volume of GCM data recovers a very similar fraction of variance.

Realistically simulated also is the wintertime SH total variance and its ENSO component (Fig. 9). The amplitude of the SH ENSO variance maximum over the South Pacific, while being only about half of its northern winter counterpart, nonetheless accounts for a similar fraction of the local total seasonal variance. Somewhat contrary to observations, the GCM results show a poleward and downstream propagation of the ENSO signal across the Southern Ocean. It is possible that such coherence may be a model bias, and not consistent with the atmospheric response pattern in nature. Likewise, in view of sparse observations over these reaches of the Southern Ocean, one is forced to also question the em-
4. The sensitivity of extratropical response patterns

a. Experiments with realistic SSTs

Using the grand ensemble GCM dataset, we compare in Fig. 10 the wintertime variance of 500-hPa heights forced by all SST variations to that forced by ENSO alone (for clarity, we repeat the middle panel of Fig. 8 in Fig. 10). This total boundary forced variance (top) is calculated from the variance of the grand ensemble GCM anomalies during 1950–99, a result often referred to as the external component of the seasonal variance (e.g., Kumar and Hoerling 1995; Rowell 1998). It should be compared to the total variance of Fig. 8 (left panel) which is the linear sum of the external component and the atmospheric internal variability. The bulk of NH wintertime external variance originates from forcing associated with ENSO. This is especially true over the PNA region where the external variance and the ENSO component are virtually identical. Only over the central North Atlantic and extreme western North America can one discern a local maximum in external variance that is unrelated to ENSO.

We have repeated the calculation in Fig. 10 (top) in which the GFDL simulations, four of which do not include any SST variations outside of the tropical Pacific, are not included in the grand ensemble. The resulting external variance is nearly identical.
An EOF analysis of the ensemble global 500-hPa heights is performed to determine the structures of atmospheric response that account for the simulated boundary forced variance. Figure 11 shows the first five EOFs for the DJF season, which together explain 87% of the global boundary-forced height variance, and contoured in each panel is the local correlation between the eigenvector’s principal component (PC) time series and that of the ensemble 500-hPa height. Shown on the right side are the PC time series, and a linear estimate of the tropical forcings that accompany each EOF. Regressions of the GCM ensemble precipitation anomalies onto that time series are shown in shading, whereas the overlaying contours are of the local correlation between the winter season SST anomalies and the PC time series.

The first structure function alone explains 62% of the global variance, and as much as 80% of the local variance over the North Pacific and North America. It is similar to the observed linear ENSO teleconnection of Fig. 1, and indeed its PC time series correlates at 0.9 with the EOF1 index time series of tropical Pacific SSTs. The third, and to a lesser extent the fifth, EOF in Fig.
11 also contain wavelike structures over the PNA sector, though with centers shifted relative to the leading EOF. Each has a plausible physical interpretation with regard to tropical forcing; the third EOF is related to non-ENSO SST forcing, while the fifth EOF is related to the modification of the linear ENSO signal by inter-ENSO SST variability. EOF3 explains up to 50% of the GCM's boundary forced variability over the central North Pacific and western North America, and has centers of action in quadrature with those of the linear ENSO teleconnection.

The pattern appears to originate from the tropical western Pacific, consistent with the rainfall regression that indicates enhanced convection over the locally warm SSTs between 130° and 170°E. Its origin is not ENSO related, however, given the fact that the major loadings of the PC time series of EOF3 are during non-ENSO years. Instead, the pattern of tropical forcing is one typically seen during the transition of the tropical ocean that precedes a mature ENSO event. In fact, the correlation pattern of SSTs in the Indo-western Pacific region projects on the so-called optimal initial structure for SST anomaly growth of Penland and Sardeshmukh (1995), a strong leading indicator for large SST anomalies in the Niño-3.4 region (5°N–5°S, 170°W–120°W) 6–9 months later.

The pattern of EOF5, by contrast, reflects the sensitivity to inter-ENSO tropical Pacific SST variability, and in particular the nonlinear component of the atmospheric response to ENSO’s extreme opposite phases. Many of the large departures of its PC time series occur during ENSO events (e.g., 1964/65, 1969/70, 1970/71, 1973/74, 1991/92, 1994/95, 1997/98). Note further that opposite phases of ENSO tend to project on the same phase of EOF5: for example, the extreme 1973/74 cold event and the extreme 1997/98 warm event each have –2 standardized departures of the PC. Suggested here is that EOF5 describes the nonlinear ENSO signal, in contrast to EOF1 which describes the linear ENSO signal. Indeed, the spatial pattern of EOF5 bears a strong resemblance to observational and GCM estimates of the nonlinear component of the atmospheric responses to ENSO’s extreme opposite phases (see Figs. 3d and 10d of Hoerling et al. 1997). We will subsequently show with composite analysis of the GCM data that different patterns of tropical interannual SST anomalies do indeed engage slightly different extratropical response patterns, confirming the impression from these higher-order EOFs.

Finally, the remaining EOFs of Fig. 11 are somewhat different in that their time series exhibit lower-frequency behavior than the predominately interannual fluctuations associated with EOFs 1, 3, and 5. The second EOF describes a tropical- and subtropical-wide increase in 500-hPa heights and a polar latitude decrease in height. This annular-like pattern is related to a warming of the Indo-Pacific warm pool together with a tropical Atlantic warming, features seen in the trend of SSTs in recent decades. Indeed, the PC time series suggests that this pattern has been emergent since about 1980. Note also that the largest loading occurred in 1998/99, a period studied by Kumar et al. (2001) and Hoerling et al. (2001b) who argued that the observed annular-like anomalies of the period 1998–2000 were forced by unusual SST warming over the tropical Indo-Pacific. The fourth EOF has its principal centers of action over the North Atlantic, the structure of which projects strongly on the North Atlantic Oscillation (NAO). Its time series shows a trend toward the positive NAO phase during the last half century, and appears related to a tropical forcing of North Atlantic climate change since 1950 as argued from additional suites of CCM3 simulations in Hoerling et al. (2001a).

Two independent assessments of the extratropical circulation sensitivity to forcing from the tropical Pacific are performed in order to verify the results of the EOF analysis. In one approach, 500-hPa height and tropical rainfall anomalies are constructed from the grand ensemble GCM data for particular years during 1950–99 that sample observed interannual SST variability in the tropical Pacific. Shown in Fig. 12 are the winter responses to three cold SST states; one associated with weak anomalies (1965,
left), a second with moderate anomalies (1955, middle), and a third with strong cold event conditions (1974, right). These correspond to $-0.6$, $-1.0$, and $-1.7$ standardized departures of the EOF1 index (see Fig. 2). In order to assess whether the strength of responses scale linearly with SST, the GCM anomalies shown in Fig. 12 are standardized by the amplitude of ENSO index. The spatial patterns of extratropical responses are very similar for the three cases, with anomaly correlations among them exceeding 0.85 when averaged over the PNA region (20°–60°N, 180°–60°W). Such lack of sensitivity in the 500-hPa heights is consistent with the lack of sensitivity in tropical rainfall responses for these cases (Fig. 12, bottom). The principal feature is the suppressed rainfall near the date line, with weaker positive rainfall anomalies in the sub-tropics.

While their spatial patterns are very similar, the amplitude of anomalies for the three cold events do not scale linearly in SST. For example, the root-mean square (rms) of the standardized 500-hPa height anomaly for the strong cold event is less than half that for the weak cold event (the actual rms height anomaly for the strong cold event is in fact slightly greater than for the weak event). This behavior is in accord with the different amplitudes of the standardized equatorial rainfall anomalies. Indicated is a saturation of the atmospheric response for progressively colder states of the equatorial central Pacific as previously noted in Kumar and Hoerling (1998) and Hoerling et al. (2001c).

It is for the various manifestations of equatorial Pacific warming that a sensitivity in the extratropical response patterns emerges. Figure 13 illustrates these for a weak warming (1980, left), a moderate warm event (1958, middle), and a strong warm event (1983, right) that correspond to $+0.6$, $+1.3$, and $+2.8$ standardized departures of the EOF1 index. Note that the amplitude of responses scales more linearly for increasingly warm SST forcing than it does for increasingly cold SST forcing. Importantly, the spatial pattern of 500-hPa height responses is very different for the strong warm event. As a measure of this difference, the pattern correlation of 500-hPa heights between the top right and left side panels in Fig. 13 is only 0.45 over the PNA region.

This sensitivity in response patterns involves more than a simple zonal shift of the wave pattern due to shifts in equatorial rainfall anomalies. Note in Fig. 13 that the center of the North Pacific anomalous low is actually anchored for the various SST warmings, and that the change in signal occurs primarily over the North American and North Atlantic regions. Only for the North American positive height anomaly is the notion of a phase shift reasonable, with the anomalous anticyclone moving from 125°W when enhanced equatorial rainfall occurs near 170°E, to 90°W when the enhanced equatorial rainfall occurs near 140°W. Even that shift is less than the zonal displacement of the tropical rainfall anomalies. Clearly these circulation sensitivities cannot be reconciled with the linear dynamics associated with the direct response to changes in tropical forcing on a zonally uniform base state. The stationarity of the Pacific response points to the existence of a geographically fixed energy source related to the deviations from zonal symmetry of that base state, and to feedbacks such as
Fig. 11. (left) The spatial structures and (right) PC time series of the first five EOFs of the ensemble GCM winter (DJF) season 500-hPa heights. The analysis is of the covariance matrix of 49 DJF samples of the simulated anomalous heights during 1950–99, and the analysis domain is global. The local correlation between the PC time series and the 500-hPa height are drawn every 0.15, and are shown for the polar cap to 20°N. The ordinate of the PC time series is of the standardized departure. Right side also shows the regressions of the GCM ensemble precipitation onto the PC time series (shading interval 0.2 mm day⁻¹ for a 1 standardized departure of PC index), and the correlation between the PC time series and SSTs drawn at 0.2 interval (beginning at 0.1). Local correlations are statistically different from zero at the 95% level for values larger than 0.3.
changes in storm tracks whose importance has been revealed in budget studies (e.g., Held et al. 1989).

The modest sensitivity in upper-air responses during different warm events yields a disproportionately large sensitivity in North American precipitation patterns. As shown in Fig. 14, both the Pacific West Coast and Atlantic East Coast experience a reversal in sign in their precipitation signal for this range in warm tropical Pacific SST forcing. When enhanced equatorial rainfall is confined to the date line (Fig. 13, left), the remote surface response includes widespread drying along the entire Pacific Coast and also over much of the northeast United States (Fig. 14, top). The areal coverage of this U.S. dry signal becomes progressively smaller as enhanced equatorial rainfall spreads into the eastern Pacific with increasingly stronger warm events. For the SST forcing associated with the strong 1982/83 event (Fig. 13, right), the U.S. precipitation response becomes wet almost everywhere (Fig. 14, bottom). This is undoubtedly related to the change in upper-tropospheric circulation responses, and in particular their attendant storm tracks that impinge anomalously onto the central Pacific coast when equatorial rainfall is abundant in the eastern Pacific. Also indicated is a sensitivity of storm tracks in the vicinity of the U.S. eastern seaboard, with a wet (dry) signal suggestive of disturbances moving anomalously northward (eastward) when equatorial convection is enhanced in the eastern (central) Pacific.

b. Experiments with idealized SSTs

As a second approach toward understanding the extratropical sensitivity to forcing from the tropical Pacific, we analyze GCM simulations using idealized warm SST anomalies as described in section 2b. As with the experiments using realistic SSTs, large ensembles are analyzed in order to reduce the bias in detecting the simulated response patterns. Shown in Fig. 15 are the atmospheric responses to warm SST forcing at three different positions across the equatorial Pacific; the eastern Pacific (right), central Pacific (middle), and western Pacific (left). The major area of enhanced rainfall (lower panels) follows the warm SST anomalies, and shifts roughly 70° in longitude as the positive SST anomaly is displaced across the Pacific basin. For some tropical regions, the rainfall anomalies show little sensitivity to the various positions of Pacific SST warming, examples of which are the recurrent drying over tropical eastern
Fig. 12. GCM ensemble winter (DJF) anomalies of (top) 500-hPa height, (middle) tropical rainfall, and (bottom) tropical SST for varying strengths of equatorial Pacific cold events during (left) 1965, (middle) 1955, and (right) 1974. The year refers to the Jan of the season. The height and rainfall anomalies have been scaled by the amplitude of the EOF1 index of ENSO for each case. Contours for height are drawn every 5 m, with positive (negative) values solid (dashed). Shading interval for rainfall is every 1 mm day$^{-1}$, with positive (negative) anomalies in blue (red). Shading interval for SST is every 0.5°C, with positive (negative) anomalies in red (blue).
FIG. 13. GCM ensemble winter (DJF) anomalies of (top) 500-hPa height, (middle) tropical rainfall, and (bottom) tropical SST for varying strengths of equatorial Pacific warm events during (left) 1980, (middle) 1958, and (right) 1983. The year refers to the Jan of the season. The height and rainfall anomalies have been scaled by the amplitude of the EOF1 index of ENSO for each case. Contours for height are drawn every 5 m, with positive (negative) values solid (dashed). Shading interval for rainfall is every 1 mm day$^{-1}$, with positive (negative) anomalies in blue (red). Shading interval for SST is every 0.5°C, with positive (negative) anomalies in red (blue).
South American and wet conditions over tropical western South America.

The sensitivity in extratropical 500-hPa heights (Fig. 15, top panels) is diminished in comparison to the considerable change in tropical Pacific forcing. The longitude of low pressure anomalies over the North Pacific is virtually fixed, as occurred also for the response patterns related to the different characters of observed El Niño events (see Fig. 13). The most significant change in response patterns is the position of the anomalous North American high that shifts 30° westward as enhanced equatorial convection shifts from the eastern to the western Pacific. Owing to this latter change, the correlation between the circulation anomalies associated
Fig. 15. Perpetual Jan MRF9 simulations of (top) 500-hPa height and (bottom) tropical rainfall anomalies for varying positions of equatorial positive SST anomalies located in the (left) West, (middle) Central, and (right) East Pacific. Results are based on a 24-month ensemble average. Contours for height are drawn every 10 m, with positive (negative) values solid (dashed). Shading interval for rainfall is every 1 mm day$^{-1}$, with positive (negative) anomalies in blue (red).
with west and east equatorial Pacific forcing is 0.45, similar to that found between the height patterns associated with weak and strong equatorial Pacific warming in Fig. 13. Another point of similarity with results in section 4a is the enhanced sensitivity of North American precipitation. Shown in Fig. 16 are the simulated patterns associated with the three positions of the idealized equatorial Pacific ocean warming. For warming of the eastern Pacific basin, much of the Pacific West Coast becomes wetter than normal. This anomaly reverses sign, and drier than normal conditions dominate the U.S. Pacific Coast in response to warming of the equatorial western Pacific Ocean.

Fig. 16. Perpetual Jan MRF9 simulations of North American precipitation anomalies for varying positions of equatorial positive SST anomalies located in the (top) West, (middle) Central, and (bottom) East Pacific. Results are based on a 24-month ensemble average. Shading interval is every 1 mm day$^{-1}$, with positive (negative) anomalies in blue (red).
5. Summary and conclusions

In their review of progress on understanding teleconnections associated with tropical SSTs during the 10-yr Tropical Ocean Global Atmosphere (TOGA) program (1985–94), Trenberth et al. (1998) highlighted two ongoing challenges: (i) improved knowledge of the tropical forcings, and (ii) determining reliably the atmospheric response patterns to those forcings. From a mid-latitude forecasting perspective it is unclear how accurate the tropical SST forcing needs to be in order to capitalize on the atmospheric system’s seasonal predictability. Herein lies the merit of the second item because improved understanding of the atmospheric sensitivity to tropical SST anomalies will establish practical tolerance levels for both monitoring and prediction errors of these forcing fields. This issue of the sensitivity to different SST forcing cannot be resolved with a high level of precision from analysis of observations, however. As revealed over two decades ago (e.g., Madden 1976), the signal of SST-forced extratropical atmospheric variability is small compared to the internally generated variability. Only upon averaging numerous cases, for example, from a collection of ENSO years in the instrumental record (none of which are alike), can one identify the grossest manifestation of the atmospheric response pattern.

Within this paradigm of detecting atmospheric signals due to tropical SST forcing, Horel and Wallace (1981) aptly framed the problem when they asked whether the fraction of extratropical variability explained by the teleconnection of Fig. 1 is constrained by the high intrinsic variability, or instead by the sampling error due to the “inadvertent superposition of an ensemble of sharper patterns.” Our study has specifically addressed this question, using large suites of ensemble atmospheric climate simulations forced with the modern record of interannually varying tropical forcing. The answer, drawn from the model results, is that the observed estimate of the fraction of year-to-year PNA sector variability explained by ENSO is indeed limited by the intrinsic atmospheric variability. Further, much of the ENSO response is itself manifested as a single spatial pattern; though additional response patterns have been identified in our study, these accounts for much less of the boundary-forced variability over the PNA sector as a whole.

This leading extratropical response pattern of the GCM, derived from an EOF analysis of the winter season ensemble mean 500-hPa height response to global SST variations during 1950–99, is found to be very similar to the observed circulation pattern associated with ENSO during this same period. Indicated hereby is that the observed ENSO teleconnection, although assembled from all manners of equatorial east Pacific SST variations, embodies most of the potentially predictable boundary-forced upper-air signal. To quantify this, we have calculated the 500-hPa height spatial anomaly correlation skill score over the PNA sector for the cases of strong tropical Pacific forcing during 1950–99. The skill score of the grand GCM ensemble anomalies, averaged across the leading eight warm events and eight cold events in this period, is 0.49. That score falls only slightly to 0.47 when calculated from the leading EOF of the GCM’s 500-hPa ensemble response.

That the midlatitude seasonal predictability associated with tropical forcing should be embodied almost entirely by a single response pattern is consistent with theoretical expectations. Dynamical models predict an overall insensitivity of the Pacific–North American response to variations in tropical Pacific forcing due to a spatially fixed energy source in the exit region of the climatological east Asian jet (e.g., Simmons et al. 1983). A second effect of the wavy climatological basic state is that the so-called Rossby wave source associated with a tropical forcing is itself insensitive to the forcing location (Sardeshmukh and Hoskins 1988). A further issue is that the tropical forcing during ENSO, as represented by the SST-induced rainfall anomalies, is in fact not very different from one event to another. We showed from the GCM simulations that similar patterns of equatorial Pacific rainfall (and extratropical responses) occurred for a wide range of cold states of the tropical Pacific, and for weak–moderate warm states. This is consistent with the known physical link between tropical convection and total SST (e.g., Graham and Barnett 1987). The relevancy of “different flavors” of ENSO is thus judged to be low in the context of its impact on hindcast skill of 500-hPa heights averaged for ENSO events during 1950–99.

Nonetheless, the EOF analysis of the ensemble GCM 500-hPa height responses confirmed the existence of additional response patterns, and these explained up to 50% of the model’s SST-forced height variance over small portions of the PNA region that coincided with the nodal points of the linear ENSO signal. Two of these EOFs (EOF 3 and 5 of Fig. 11) resemble wave trains emanating from the Tropics. Regressions of the GCM rainfall on the PC time series of each eigenvector confirmed their coexistence with large-scale patterns of tropical Pacific and Indian Ocean rainfall anomalies. One of the teleconnections was associated with the transitional state of the tropical Indo-Pacific Ocean that often precedes by several seasons the occurrence of mature ENSO conditions. It was thus interpreted as an example of a non-ENSO teleconnection, and its contribution to forecast skill during “ENSO-neutral” winters has yet to be assessed. The other teleconnection described by EOF5 was interpreted as the modest “tuning” of the leading ENSO response pattern due to inter-ENSO variations in forcing, and in particular that related to an asymmetry in atmospheric response patterns to ENSO’s extreme opposite phases.

A physical basis for the empirical modes of variability was further established using case studies of Pacific–North American responses to specific ENSO events during 1950–99, as well as analysis of additional experi-
ments using idealized tropical Pacific SST anomalies. A change in the character of the midlatitude response occurred when rainfall was enhanced in the far eastern equatorial Pacific, as happens during the strongest warm events such as 1982/83 and 1997/98. This included a downstream shift of the North American anticyclone response relative to weaker warm events, and a considerable increase in spatial scale of the North Pacific cyclone response. The overall height pattern still resembled that forced by weaker warm events (0.5 spatial anomaly correlation); however, a disproportionate sensitivity occurred in North American rainfall. Along the Pacific West Coast and North Atlantic East Coast, the sign of precipitation anomalies reversed compared to those forced by weaker warm events.

Given the rarity of such cases, it is not surprising that the modification of upper-air response patterns relative to the leading EOF contributes little to the half-century average skill score. Yet, it is reasonable to argue that, when the necessary tropical forcing arises, an individual forecast may benefit by departing from the leading "canonical" linear ENSO signal. There are two caveats to bear in mind. One is that the ability to quantify the value added by incorporating such new information will require many cases to establish. A second related issue is the ability to verify the realism of the model behavior itself. We sought to reduce the influence of biases in any single GCM by performing our analysis on a combined GCM dataset consisting of four different models. From this data, the realism of the combined model variability in both tropical forcing functions and extratropical wintertime circulation was illustrated. Beyond such gross metrics, the brevity of the instrumental record does not permit an assessment of the high-order sensitivities. We can be certain that some biases are systematic across most current medium resolution climate models, for example, deficiencies in the structure and the statistics of intraseasonal tropical disturbances (e.g., Sperber et al. 2001). But it is unclear how important those are for the seasonal mean atmospheric response patterns. It goes without saying that the analysis performed herein needs to be repeated with improved, perhaps higher-resolution, climate models. A practical matter in doing so is to evaluate, for a fixed period such as 1950–99 as used herein, whether new GCMs yield new evidence on the sensitivity of atmospheric circulations to tropical forcing, and especially whether such behavior entails improvements in seasonal predictions. It is equally important to determine which physical factors are relevant for determining the character of tropically forced teleconnections in order to inform model development and clarify the tolerance of predictions to GCM errors.

Finally, if the results of the current study are robust, they have two specific implications for the prediction of tropical SSTs. On the one hand, since most of the potentially predictable signal in extratropics appears related to a single response pattern, and the fact that the time series of this pattern is a proxy for the time series of the leading EOF of tropical Pacific SSTs that describes ENSO, improvements in midlatitude skill will come from improved forecasts of the phase and the amplitude of that SST EOF with increasingly longer lead times. On the other hand, in order to capitalize on the additional atmospheric response patterns identified herein, predicting details of the SST forcing are required that have previously received little attention. For example, the results of our GCM ensembles indicate that different extratropical responses were forced during the two extreme warm events of 1982/83 and 1997/98 (see the different EOF 3 and 5 projections for these cases in Fig. 11) when forced with observed SSTs. To capitalize on the atmospheric sensitivity that goes beyond the dominant single ENSO signal will require that SST predictions themselves be skillful beyond the projection of their spatial anomalies onto the leading EOF of the tropical Pacific SSTs as shown in Fig. 2.

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