The Nature and Causes for the Delayed Atmospheric Response to El Niño

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ABSTRACT

Remarkable among the atmospheric phenomena associated with El Niño–Southern Oscillation (ENSO) is the lag in the zonal mean tropical thermal anomalies relative to equatorial east Pacific sea surface temperatures (SSTs). For the period 1950–99, the maximum correlation between observed zonal mean tropical 200-mb heights and a Niño-3.4 (5°N–5°S, 120°–170°W) SST index occurs when the atmosphere lags by 1–3 months, consistent with numerous previous studies. Results from atmospheric general circulation model (GCM) simulations forced by the monthly SST variations of the last half-century confirm and establish the robustness of this observed lag.

An additional feature of the delay in atmospheric response that involves an apparent memory or lingering of the tropical thermal anomalies several seasons beyond the Niño-3.4 SST index peak is documented in this study. It is characterized by a strong asymmetry in the strength of the zonal mean tropical 200-mb height response relative to that peak, being threefold stronger in the summer following the peak compared to the preceding summer. This occurs despite weaker Niño-3.4 SST forcing in the following summer compared to the preceding summer.

The 1–3-month lag in maximum correlation is reconciled by the fact that the rainfall evolution in the tropical Pacific associated with the ENSO SST anomalies itself lags one season, with the latter acting as the immediate forcing for the 200-mb heights. This aspect of the lagged behavior in the tropical atmospheric response occurs independent of any changes in SSTs outside of the tropical east Pacific core region of SST variability related to ENSO. The lingering of the tropical atmospheric thermal signal cannot, however, be reconciled with the ENSO-related SST variability in the tropical eastern Pacific. This part of the tropical atmospheric response is instead intimately tied to the tropical ocean’s lagged response to the equatorial east Pacific SST variability, including a warming of the tropical Indian and Atlantic SSTs that peak several seasons after the Niño-3.4 warming peak.

1. Introduction

The zonally homogeneous response of tropical temperatures is the most robust impact of El Niño–Southern Oscillation (ENSO). Atmospheric warming occurs from the surface to the tropopause during ENSO’s warm phase (e.g., Newell and Weare 1976; Angell and Korshover 1978; Pan and Oort 1983; Reid et al. 1989), resulting in an elevation of pressure surfaces in the upper troposphere. Bjerknes (1972) and Horel and Wallace (1981) highlighted the spatial coherence of the geopotential height response throughout the Tropics, a pattern that is consistent with the horizontal redistribution of heat by tropical thermally direct mass circulations (e.g., Wallace 1992).

A remarkable feature of this tropical tropospheric response is its lag relative to the tropical eastern Pacific sea surface temperature (SST) forcing. This was an early discovery in ENSO research, being documented in Newell and Weare’s (1976) analysis of the relation between sea surface temperatures of the equatorial east Pacific and 700–300-mb thicknesses at 14 tropical stations during 1958–73. In numerous other accounts, the maximum correlation between the zonal mean tropical tropospheric temperatures and the SST in the equatorial east Pacific occurs when the SST leads the atmosphere by one to two seasons (Newell and Weare 1976; Angell 1981; Pan and Oort 1983; Reid et al. 1989; Yulaeva and Wallace 1994). This feature has been confirmed from the analysis of different subperiods within the historical archives of observational data, within case studies of individual ENSO cycles, and within climate model responses to SST boundary forcing related to ENSO. Yet, its origin continues to defy explanation. The purpose of this study
is to revisit the nature of the lagged behavior of the tropical response to east Pacific sea surface temperature variations, establish its robust features, and offer physical and dynamical explanations for its cause.

A recent example is offered in Fig. 1, based on results of Hoerling et al. (2001), that illustrates observed and simulated 200-mb heights during the life cycle of the 1997/98 strong El Niño. East equatorial Pacific SSTs reached their peak intensity during the fall of 1997 (Fig. 1, right), whereas the zonally averaged tropical height anomalies peaked in late winter (Fig. 1, left), a lag consistent with many earlier studies. For this particular case, an additional feature is the prolonged warm tropical troposphere throughout 1998 despite the reversal in sign of the tropical east Pacific SST anomalies. Many of these observed atmospheric anomalies during 1997/98 were reproduced in an ensemble of climate simulations (Fig. 1, middle), confirming the SST forcing’s strong controlling influence.

By comparison to the extensive documentation for a delay in the tropical atmospheric response, there has been a dearth of explanations for its source. It is generally believed that the tropical tropospheric warming during El Niño results from the increase in precipitation over anomalously warm SSTs in the central to eastern equatorial Pacific. Reid et al. (1989) argued for a slow (1 m s⁻¹) zonal propagation of the tropical thermal response from a point of origin in the eastern Pacific where convection is enhanced. But it appears to be an unlikely cause for the one season lag in view of the theoretical results on the rapidity with which warming can spread across the Tropics in response to isolated heating (e.g., Heckley and Gill 1984; Bantzer and Wallace 1996).

A combination of observational, statistical analysis, and general circulation model (GCM) experimentation is employed to identify the mechanisms for a lagged atmospheric response to east Pacific sea surface temperature variations. The paper first builds upon previous studies to document the tropical atmospheric thermal anomalies associated with ENSO by examining the lead–lag relationship between the interannual variability of Niño-3.4 (5°N–5°S, 120°–170°W) SST anomalies and the 200-mb height anomalies during 1950–99. In light of the realism of climate simulations as shown in Fig. 1, we perform a parallel analysis of such relationships using the ensemble mean output of GCM experiments forced by the global, observed SSTs during the same record. These allow us to establish the robust, characteristic features of the delay in atmospheric responses, and test hypotheses for their origin. An additional suite of GCM experiments with different specifications of SST anomalies is also performed in order to understand the role of ENSO-related SST anomalies in different ocean basins on the tropical atmosphere. The datasets and the experiments are described in section 2. Results are presented in section 3, the first section of which documents the lagged behavior, and the second section of which proposes and tests various causal mechanisms. A summary is provided in section 4.

2. Data and analysis procedure

a. Observational data

Our analysis focuses on anomalies in 200-mb heights related to the ENSO cycle. This variable is a useful measure for the vertically averaged temperature between roughly the surface and the tropopause, and thus serves as a proxy for the tropospheric thermal response to ENSO. We analyze the observed monthly mean 200-mb heights for 1950–99 provided by the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis that are available on a 2.5° latitude–longitude grid (Kalnay et al. 1996).

Monthly mean observed sea surface temperature for the same period is obtained on a 2° latitude–longitude grid from a blended dataset based on two sources. The first constructs global fields during 1950–80 by projecting the sparse, monthly in situ SSTs onto empirical orthogonal functions (Smith et al. 1996). The second is a global SST analysis constructed by combining in situ and satellite observations using optimum interpolation (Reynolds and Smith 1994) during 1981–99. Observed monthly anomalies are calculated for both variables relative to a 1950–99 climatology.

We employ the interannual variability of SSTs in the so-called Niño-3.4 region that encompasses the area 5°N–5°S, 120°–170°W as an index for the tropical east Pacific forcing associated with ENSO. This monthly index for 1950–99 is regressed against monthly 200-mb height anomalies at each grid point to obtain the global atmospheric linear response to the east Pacific SST variations. We perform two different regression analyses. The first is a simultaneous regression between the Niño-3.4 SST index and 200-mb heights. These are computed for each calendar month separately, and the spatial maps provide information about the simultaneous atmospheric response. The second is a monthly lead–lag regression between the Niño-3.4 index and 200-mb heights for up to 12 months preceding and 12 months after the index. We calculate the lead–lag relations with respect to January values of the Niño-3.4 index, corresponding roughly to the month of maximum east Pacific SSTs. We thus seek to describe the character of 200-mb heights that precede (lead) this SST peak by up to 12 months, and the height behavior that follows (lags) this peak for up to 12 months.

To complement the simultaneous regression analyses, ENSO composite maps for 200-mb heights are also analyzed. The warm and cold ENSO events are defined according to a one standard deviation exceedence of the Niño-3.4 SST index. Composite maps are constructed based on both simultaneous as well as lead–lag analysis. We should point out that for the simultaneous compos-
b. Atmospheric GCM experiments

We also apply the methodologies described above to an ensemble of atmospheric GCM simulations. One model used is NCAR’s Community Climate Model (CCM3; Kiehl et al. 1998) run at T42 resolution and forced with the observed monthly variations of global SSTs during 1950–99. A 12-member ensemble has been performed in which the individual realizations start from different atmospheric initial conditions, but are forced with the same prescribed SST and sea ice boundary conditions. Our analysis is based on the ensemble mean atmospheric anomalies of CCM3, which are computed with respect to the model’s 1950–99 climatology.

We also analyzed two sets of GCM simulations with different configurations of SSTs. These simulations are performed using the Geophysical Fluid Dynamics Laboratory (GFDL) R30 atmospheric GCM coupled to an ocean mixed layer model. The horizontal resolution of the R30 GCM is approximately 2.25° latitude by 3.75° longitude, and the GCM has 14 vertical sigma levels. In the first set of experiments, the observed SSTs for the 1950–99 period are specified only over the eastern tropical Pacific (15°S–15°N, 172°E–South American coast). SSTs at all the other ocean points are specified as a climatologically varying seasonal cycle. In this paper these simulations are referred to as POGA (the Pacific Ocean and Global Atmosphere). The atmospheric responses in the POGA experiment are solely due to the ENSO-related SST variability in the tropical eastern Pacific.

In a companion experiment, SSTs in the tropical eastern Pacific are again specified as in the POGA experiment, but SSTs elsewhere in the World Ocean are free to evolve according to coupled interactions with a mixed layer model. In the mixed layer model a vertical column of specified ocean depth is coupled to the atmosphere, and SST in that column evolves according to the exchange of air–sea fluxes. Details for these so-called POGA–mixed layer (ML) experiments are outlined by Alexander et al. (2002). The oceanic response outside the tropical eastern Pacific in the POGA–ML experiments is due to the “atmospheric bridge” mechanism linking ENSO SST variability in the tropical eastern Pacific with the rest of the global oceans (see Alexander et al. 2002; Lau and Nath 1996). An ensemble of 8 POGA and 16 POGA–ML realizations were available for the 1950–99 period.

Regression and composite analysis techniques described above are also applied to the ensemble mean atmospheric responses in the POGA and POGA–ML experiments. The difference in the lead–lag atmospheric responses between the POGA and POGA–ML is indicative of the atmospheric feedback of SST variability in the oceans outside the tropical eastern Pacific. At the same time, this SST variability itself is because of the ENSO SST variability in the tropical eastern Pacific, which are the only SSTs specified externally in these experiments.
3. Results

a. Evidence for a delayed atmospheric response

Figure 2 shows the lead–lag correlations of January values of the Niño-3.4 SST index with the tropical SSTs and 200-mb height. The correlations are calculated at each grid point. For 200-mb heights, these are zonally averaged to produce the time–latitude sections in Fig. 2 (left and middle). For SSTs, correlations are averaged for 10°N–10°S to produce the time–longitude sections in Fig. 2 (right). Simultaneous correlations during January are plotted for ordinate value of “0,” whereas correlations that precede (follow) the January Niño-3.4 index are specified by the leading (lagging) months on the negative (positive) ordinate.

The SST correlations (Fig. 2, right) illustrate the tropics-wide oceanic evolution during ENSO’s life cycle, having a characteristic timescale of about 18 months. Beginning with initial SST signals at 9-month lead in the west-central equatorial Pacific, the positive correlations indicate that the SST anomalies associated with ENSO typically appear during the spring prior to their boreal winter peak, a result consistent with earlier studies (e.g., Rasmusson and Carpenter 1982). By contrast, the termination of the ENSO event occurs in the tropical Indian and Atlantic Oceans at a 9-month lag. The positive correlations between Niño-3.4 SSTs and those over the Indian and the Atlantic Oceans are largest at a 1–2-season lag, and are in phase with the equatorial Pacific SST anomalies as noted in Pan and Oort (1983) and studied recently in Klein et al. (1999).

The zonal mean 200-mb heights lag the January Niño-3.4 SST index by 1–3 months in both observations (Fig. 2, left) and GCM simulations (Fig. 2, middle). During the evolution of ENSO, positive correlations between Niño-3.4 SSTs and 200-mb heights initially appear on the equator in late summer preceding the Niño-3.4 index peak, consistent with a tropical tropospheric warming (cooling) during the warm (cold) phase of ENSO. The correlations increase with time, and also broaden meridionally indicative of a tropospheric warming that spreads into the subtropics by the following spring and summer seasons.

A remarkable feature in Fig. 2 is the appreciable difference in evolution of the oceanic and atmospheric anomalies during the ENSO cycle. In particular, while the correlation between SSTs and the Niño-3.4 index peaks during December in the east equatorial Pacific, indicative of the mature phase of the tropical Pacific ocean expression of ENSO, the tropical height correlations do not peak until the following March.

The lagged relationship between the Niño-3.4 SSTs and the atmospheric tropical thermal response has further expression in terms of the prolonged positive height correlations that exist throughout the second summer season. Note that both observed and simulated zonal mean 200-mb height correlations exceed 0.4 at an 8-month lag, despite the fact that the Niño-3.4 SST anom-
Fig. 3. Same as in Fig. 2 but for the zonally averaged warm ENSO event lead–lag composite anomalies. Warm events are defined to be
the years when the Jan Niño-3.4 index exceeds a +1 std dev departure. Units are m for the 200-mb heights (contoured every 5 m), and °C
for the SSTs (contoured every 0.5°C).

The composite results reiterate the divergence between the tropical oceanic and atmospheric responses during the evolution of ENSO’s extreme phases. First, the peak amplitude of the tropical tropospheric warming (cooling) during warm (cold) events occurs in the February–April period, whereas the tropical east Pacific SST anomalies peak in the previous December. Second, the atmospheric response lingers with appreciable amplitude into the following summer and fall season despite the sharp decline in tropical east Pacific SST anomalies. Note, for example, that the zonally averaged 200-mb height anomalies at the 6-month lag (July) are equal to their values at the 1-month lead (December), whereas the tropical east Pacific SST anomalies decline by more than half their value.

To further illustrate the nature of the delay in atmospheric responses, we present in Figs. 5 and 6 the global patterns of the composite warm and cold event SST (left panels), observed 200-mb height (center panels), and GCM 200-mb height (right panels) anomalies. These are shown for three select months, one for the January base month of the Niño-3.4 index (middle rows), and the others for the Julys that lead (top rows) and lag (bottom rows) the January base, respectively. The two July composites have been selected to emphasize the asymmetry in the atmospheric response with respect to ENSO evolution, and we will subsequently refer to the previous summer as July(−) and the following summer as July(+).

Consistent with many previous studies, the composite maps show that the tropical Pacific SST anomalies during ENSO are already well established in July(−), with further strengthening by January. By July(+), these anomalies decay considerably, and a point to note is that the amplitude of SST anomalies in the tropical eastern Pacific during July(−) is larger than the amplitude during July(+). Elsewhere in the Tropics, SST anomalies are generally not strong in the Indian and the Atlantic
Fig. 4. The same as in Fig. 3, but for cold ENSO event lead-lag composites.

Fig. 5. The spatial maps of warm ENSO event composite anomalies for (left column) SST (°C), (middle column) observed 200-mb height (m), and (right column) CCM3 simulated 200-mb height (m). Warm events defined as in Fig. 3. The top (bottom) row shows Jul composites preceding (following) the Jan base month of the Niño-3.4 index. The middle row shows simultaneous composites for Jan. Negative anomalies are dashed.
Oceans during July(−), but these grow by winter and assume the same sign as the east Pacific SST anomalies. As such, a tropicalwide pattern of like-signed SST anomalies exists by July(+).

The tropical 200-mb height response is larger in January than either in the previous or following July. Likewise, a prominent response in the Northern Hemisphere extratropics can be seen in January, whereas a much weaker signal occurs in summer. With regard to the delay in atmospheric 200-mb height responses, it is evident from Figs. 5 and 6 that a stronger, and larger-scale height response exists in July(+) compared to July(−). This asymmetry across the consecutive summers of the ENSO cycle occurs despite the fact that SST anomalies in the tropical Pacific are actually weaker in July(+) compared to July(−). In many ways, the increase in homogeneity of the tropospheric thermal response by July(+) is consistent with the increased homogeneity SST anomalies throughout the Tropics.

To what extent does the temporal evolution of the zonal mean tropical response differ from the monthly varying simultaneous atmospheric response to Niño-3.4 SSTs? In other words, how important is the evolutionary aspect of tropical SSTs in determining the evolving zonal mean response? To answer these questions, we compare the lead–lag analysis with the monthly varying, simultaneous sensitivity to Niño-3.4 SSTs. The latter is calculated by regressing the 200-mb heights onto the Niño-3.4 index for each calendar month. The regressions are then scaled by the actual phase and amplitude of the Niño-3.4 SST index following the ENSO cycle. As such, identical Niño-3.4 SST forcing is employed in both approaches, but the temporal evolution of the tropical SSTs are preserved only in the lead–lag analysis. Results are shown for the (warm minus cold event) SST differences, the evolution of which is shown in Fig. 7 (right).

Comparing results from the simultaneous (Fig. 7, left) and lead–lag (Fig. 7, middle) analyses reveals very similar atmospheric thermal responses during the growing phase of Niño-3.4 SST anomalies. Furthermore, the 1–3-month lag in the maximum tropical atmospheric response is reproduced in the simultaneous regressions. By contrast, there is a large difference between the two analyses during the decay phase of Niño-3.4 SST anomalies, and it is clear that specifying the seasonally varying tropical response solely from knowledge of the simultaneous sensitivity to Niño-3.4 SSTs severely underestimates the atmospheric signal during the latter half of the ENSO life cycle.

b. Possible origins for the delayed atmospheric response

Our analysis of the origins for the delayed atmospheric response deals with two separate issues: 1) a lag
of 1–3 months between the maximum in 200-mb tropospheric height response and the maximum in the Niño-3.4 SST anomalies in the tropical eastern Pacific, and 2) a marked asymmetry between the temporal evolution of Niño-3.4 SST anomalies and tropically averaged 200-mb heights.

Regarding the origin of the lagged atmospheric response, one hypothesis is that it may reflect a seasonally varying sensitivity of the tropical troposphere to Niño-3.4 forcing. Indeed, the results of Fig. 7 indicate that the 1–3-month lag in the zonal mean 200-mb height response could originate from such a sensitivity. It is already well established that the extratropical response to ENSO varies strongly with the seasonal cycle, and that this results from changes in the background atmospheric state rather than changes in tropical east Pacific forcing (e.g., Webster 1982). The seasonal cycle of the tropical atmospheric circulation is considerably weaker, and it remains an open question if these variations modulate the tropical atmospheric response to ENSO as implied by Fig. 7.

In fact, further diagnosis indicates that the reproduction of a 1–3-month lag in atmospheric responses from the simultaneous sensitivity analysis is due to such seasonally varying sensitivity in the atmospheric response to Niño-3.4 forcing. It specifically involves a seasonal change in the tropical rainfall response, rather than a change in the atmospheric sensitivity due purely to the seasonal cycle of the tropical atmospheric circulation. The temporal evolution of zonal mean precipitation is shown in Fig. 8 (bottom left) based on the CCM3 simulations, and it too possesses a 1–3-month lag relative to the Niño-3.4 index. It thus maximizes simultaneously with the zonal mean 200-mb height response (Fig. 7, bottom middle). Calculations using a linear baroclinic model driven by time-evolving zonal mean diabatic heating that mimics this behavior of the zonal mean rainfall confirms that the 200-mb heights are indeed responding to such forcing (not shown).

A fundamental question regarding the 1–3-month lagged atmospheric thermal response is thus the cause for the lag in the tropical rainfall. It should first be noted
that the tendency of zonal mean tropical rainfall to maximize in late winter is not due to an increase in zonal mean SST anomalies, since the latter exhibit no lag relative to the Niño-3.4 index (see Fig. 8). The lagged zonal mean rainfall is instead related to a sensitive phasing of interannual SST anomalies with the climatological seasonal cycle. Of importance are the total, rather than anomalous SSTs, in forcing the tropical convection in the eastern Pacific. To illustrate, Fig. 9 superimposes equatorially averaged SST anomalies during ENSO’s life cycle (shading) onto the seasonal cycle of climatological SSTs (contours). When east equatorial Pacific SST anomalies approach their peak during fall, the seasonal cycle of SSTs is near a minimum. The climatological SSTs warm in subsequent seasons, and this occurs at a rate much faster than the decay of ENSO SSTs. As such, the total SSTs achieve peak amplitude in that region by early spring. Since the convective rainfall depends on the total SSTs (Graham and Barnett 1987; Zhang 1993), these weaker east Pacific SST anomalies following the Niño-3.4 peak are in fact more effective in exciting rainfall. The equatorial average of CCM3 rainfall following ENSO’s evolution (Fig. 8, bottom right) confirms the existence of such an increase in the east equatorial Pacific rainfall, in phase with the increase in zonal mean rainfall. It follows that the zonal mean diabatic forcing associated with the temporal evolution of ENSO does not decrease at the same pace as implied by the evolution of the Niño-3.4 index, but is phase locked with the seasonal cycle of climatological tropical SSTs.

Another factor contributing to a lag in the zonal mean rainfall relative to Niño-3.4 SSTs is the evolution of rainfall anomalies in the western equatorial Pacific. During the growth phase of Niño-3.4 SST anomalies in fall, these are out of phase with the east-central equatorial Pacific rainfall anomalies (Fig. 8, bottom left). By the following spring, these rainfall anomalies weaken and become in phase with the rainfall anomalies over the eastern Pacific, thereby adding to the amplitude of the
It is through the analysis of additional GCM experiments that the 1–3-month lag between the tropical heights and the ENSO SST variability can be attributed to the SST variability in the tropical eastern Pacific alone, and its interaction with the seasonal cycle. The zonal mean of the lead–lag composites for 200-mb heights based on the POGA experiment is shown in Fig. 10. The height response in these simulations (Fig. 10, left) also has a 1–3-month lag relative to the Niño-3.4 SST index. Recall that the observed SSTs variability are specified only in the tropical eastern Pacific in the POGA simulations, and SSTs outside this region evolve through a climatological seasonal cycle. This 1–3-month lag in the tropical height response is thus entirely due to the interannual variability in the tropical eastern Pacific.

The behavior of the equatorial zonal mean tropical rainfall cannot, however, be used to explain the lingering zonal mean 200-mb response throughout spring and summer following the peak of the Niño-3.4 index [hereafter denoted by year(+)], nor the striking asymmetry of the tropical atmospheric response between July(-) and July(+). Note in Fig. 8 that the equatorial zonal mean rainfall during July(-) is almost identical to July(+). One might thereby be tempted to interpret the slow decay time of atmospheric zonal mean thermal zonal mean rainfall forcing at that time. The origin for this variation in the rainfall anomalies over the western Pacific is not clear, and cannot be determined conclusively from these simulations.
anomalies in the Hovmöller results of Fig. 7 as reflecting an atmospheric memory of Niño-3.4 forcing. However, there is ample evidence to dismiss the notion of such an atmospheric thermal inertia. Theoretical and modeling considerations indicate that the tropical troposphere equilibrates rapidly to anomalous tropical heat sources (e.g., Heckley and Gill 1984; Jin and Hoskins 1995; Bantzer and Wallace 1996). The mechanism for this equilibration involves convectively driven Kelvin waves, which effectively communicate heating anomalies within the longitudinal expanse of the near-equatorial troposphere on the timescale of a week to 15 days. In the absence of Niño-3.4 SST and related tropical diabatic heating forcing, the atmospheric signal would be expected to return to climatological conditions on a similar timescale. Thus, while it is reasonable to seek a link between the lingering aspects of delayed thermal response and the steady tropical forcing, neither Niño-3.4 SSTs nor the equatorial zonal mean diabatic heating appear to be useful candidates.

Another possibility is the impact of the slow evolution of SST anomalies in the Indian and the Atlantic Ocean basins, which are linked to SST variability in the tropical eastern Pacific (see Fig. 2, right). Evidence for this comes from the comparison of the zonal mean lead–lag 200-mb height composites for the POGA and POGA–ML experiments in Fig. 10, where the meridional average of the lead–lag SST composites for the POGA–ML experiment are also shown. The POGA–ML run realistically simulates the lagged warming of the Indian and the Atlantic Oceans. It is evident from the differences in the atmospheric responses between the POGA–ML and POGA runs that this lagged ocean warming exerts a strong influence on the tropical atmosphere. The evolution of the tropospheric height response in the POGA–ML simulation is highly asymmetric relative to the January Niño-3.4 SST peak, as also seen in the observations, whereas such a marked asymmetry is lacking in the POGA run. The lingering response in the tropical tropospheric warming following the ENSO evolution is thus maintained by the lagged evolution of SSTs in the tropical ocean basins outside the Niño-3.4 index region.

To further illustrate this, Fig. 11 shows the temporal evolution of the POGA–ML minus POGA differences for select variables that have been area averaged over the entire tropical belt between 30°S and 30°N. By year(+) 200-mb heights for the POGA–ML are higher than for the POGA run, reflecting the warmer tropical troposphere in the POGA–ML. This elevated tropical tropospheric warmth in the POGA–ML is accompanied by an elevated warmth in the tropical SSTs (due to the coupled model’s response to Niño-3.4 SSTs), and also by an increase in tropically averaged rainfall.

The ultimate cause for the lingering tropical tropospheric warming is thus clearly the lagged warming in the tropical SSTs as a whole. In fact, it is seen from Fig. 11 that there exists virtually no lag between tropically averaged SSTs and tropospheric warming. Yet, a question remains as to how the influence of these SST anomalies is communicated to the free atmosphere. It has been previously argued that the tropical tropospheric temperature anomalies are associated with the SST anomalies at the lower boundary through a simple moist-adiaabatic adjustment process (Emanuel et al. 1994). This relationship between SSTs and the tropospheric temperatures is consistent with a physical model in which deep convection renders the tropical tropospheric temperature profile a moist adiabat starting from the near-surface moist static energy. Within this paradigm, the lingering of the tropical height response will be consistent with the above-normal SSTs in the tropical ocean basins, with deep convective rainfall anomalies communicating the near-surface layer warming to the

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1 SST composites for the POGA are same as for the POGA–ML except that the POGA simulation SST composites do not have any SST evolution east of 172°E, which is the easternmost boundary up to which interannual SST variations in these experiments is specified.
free atmosphere. It is not clear whether anomalous precipitation is a necessary condition for this to occur, however. Despite the strong relationship between tropically averaged tropospheric warming, SST warming, and increased rainfall seen in Fig. 11, the role of precipitation anomalies may be coincidental to achieving the tropospheric warming. For example, Su et al. (2003) find a poor relationship between tropically averaged tropospheric temperature and rainfall anomalies when comparing different El Niños, though they find a strong linear relation between tropically averaged tropospheric temperatures and SSTs.

4. Summary

Results of our observational analysis covering the 1950–99 record highlight two prominent features of the tropical atmospheric thermal response to SST variations in the east equatorial Pacific. One, well documented in earlier studies, is that the maximum correlation relating the zonal mean tropical 200-mb height response to Niño-3.4 SSTs occurs when the atmosphere lags by 1–3 months. The second, and hitherto less well-recognized feature, is a lingering of the tropical atmospheric response by up to 9 months after the winter (January) peak in Niño-3.4 SST anomalies. This latter behavior was a most singular and remarkable aspect of the zonal mean anomalies observed during the recent 1997–99 ENSO cycle (Hoerling et al. 2001; Kumar et al. 2001), and it is here confirmed to be characteristic of the atmosphere’s response during ENSO’s evolution of the last half-century. It is manifested by a strong asymmetry in the tropical zonally averaged thermal response between the antecedent summer when the Niño-3.4 SST anomalies are moderate and growing, and the following summer when they are weak and decaying. The zonal mean signal is shown to be threefold larger in July(+) compared to July(−) strength. A parallel analysis of the delay in atmospheric responses is conducted using climate model simulations forced by the monthly SST variations of 1950–99. These confirm the robustness of the observational results, both the one season lag in the maximum correlation between Niño-3.4 SSTs and tropical 200-mb zonal mean heights, and the strong asymmetry of those heights between the consecutive summers bracketing the peak in the Niño-3.4 forcing.

Causes for a one season lag in the atmospheric response are intimately linked with the temporal evolution of tropical rainfall during ENSO’s life cycle. The one season lag relative to the Niño-3.4 SSTs occurs because the tropical rainfall itself lags the Niño-3.4 index by one season. Therefore, while Niño-3.4 SST anomalies peak in early winter, the zonal mean rainfall anomalies peak in late winter. The associated zonal mean diabatic heating is thus in phase with the zonal mean tropical 200-mb heights, a relation that likely reflects the forcing of the latter by the former. The lag in zonal mean tropical rainfall anomalies can be attributed to the seasonality of the total SST variability in the tropical eastern Pacific. The decay in east equatorial Pacific SST anomalies in late winter of year(+) occurs while the climatological annual cycle of SSTs is approaching its warmest state. As such, total SSTs are greatest in late winter, and are most effective in forcing local enhancement of convection at that time. This attribution is further confirmed in additional sets of GCM experiments where the observed SST variability is specified in the tropical eastern Pacific alone.

The cause for the lingering of the tropical zonal mean thermal response for several seasons after the Niño-3.4 peak cannot be reconciled with a lingering in the equatorial tropical Pacific SST anomalies. In fact, the impressive asymmetry in strength of the zonal mean atmospheric signal between the consecutive summers that embrace an ENSO event has no counterpart in the behavior of the equatorial zonal mean rainfall, which is instead slightly weaker by July(+) compared to July(−).

Using a parallel set of coupled and uncoupled general circulation model experiments, the lingering of the tropical thermal response is shown to be due to a lagged evolution of tropical Indian and Atlantic SST anomalies relative to the Niño-3.4 index. In one set of GCM simulations (referred to as POGA), observed interannual SST variability was specified in the tropical eastern Pacific alone, and SSTs over other ocean basins evolve through a climatological seasonal cycle. In the companion set of GCM simulations, referred to as POGA–ML, SSTs outside the tropical eastern Pacific were computed using a mixed layer ocean model. SST anomalies outside the tropical eastern Pacific evolve much as observed in the POGA–ML simulations, consistent with an atmospheric bridge mechanism (see also Alexander et al. 2002; Lau and Nath 1996). The 30°N–30°S tropically averaged 200-mb height response retains high values well beyond the Niño-3.4 peak in the POGA–ML simulations as is seen also in the observations, but fail to do so in uncoupled POGA simulations. The comparison of these experiments reveals that the lingering of the tropical tropospheric signal relative to the Niño-3.4 peak is due to the SST response in the remote tropical ocean basins that emerge during the evolution of ENSO.

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