On the Forcing of ENSO Teleconnections by Anomalous Heating and Cooling

ERIC DEWEaver
Department of Atmospheric and Oceanic Sciences, University of Wisconsin—Madison, Madison, Wisconsin

SUMANT NIGAM
Department of Meteorology, University of Maryland, College Park, College Park, Maryland

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ABSTRACT

ENSO teleconnections were originally regarded as a single train of stationary Rossby waves generated by a compact region of enhanced (reduced for La Niña) equatorial convective heating. While more recent studies have greatly enhanced this dynamical picture, the dominant conceptual model of the teleconnections still identifies this monopolar convective heat source as the ultimate driver of the teleconnections.

This note presents evidence that the surrounding regions of diabatic cooling are just as important as equatorial heating in producing the ENSO teleconnections. In simulations with a linear diagnostic model, heating and cooling anomalies derived from the National Centers for Environmental Prediction—National Center for Atmospheric Research (NCEP—NCAR) reanalysis make comparable contributions to the upper-level eddy height anomalies. In particular, remote cooling is just as important as local heating in determining the central longitude of the subtropical El Niño anticyclones.

The same diagnosis is applied to the ENSO response of an atmospheric general circulation model (AGCM) forced by observed sea surface temperatures in an integration performed by the NASA Seasonal-to-Interannual Prediction Project (NSIPP). Despite differences in the climatological basic state and diabatic heating, positive and negative heating anomalies play the same complimentary roles for the simulated ENSO response as they do for the observed ENSO pattern.

1. Introduction

One of the best-known figures in climate dynamics is the schematic in which Horel and Wallace (1981, their Fig. 4) outlined the dynamics of El Niño–Southern Oscillation (ENSO) teleconnections. In it, equatorial convection forces a train of stationary Rossby waves that arcs across the Pacific–North American (PNA) sector on a great-circle route. Many subsequent studies have revised this account, citing the key roles played by vorticity gradients in the Asian jet (Sardeshmukh and Hoskins 1985), midlatitude storm tracks and stationary waves (e.g., Held et al. 1989; Ting and Hoerling 1993), and the barotropic instability of the background flow (Simmons et al. 1983).1 Yet despite these added complexities, the schematic remains in use as a conceptual model. For example, the review paper of Trenberth et al. (1998, their Fig. 4) offers a similar schematic as a teleconnection “protomodel.” The dynamical underpinnings have become much more sophisticated, but the figure still shows a wave train, forced—albeit indirectly—by a monopolar equatorial heat source.

To be sure, ENSO sea surface temperatures (SSTs) do produce a strong and uniform convection response over most of the equatorial Pacific. However, the local convection response is just one component of a much larger pattern of diabatic heating anomalies, as can be seen in Fig. 1 (details in section 2). In this figure, equatorial enhancement comes at the expense of the surrounding regions, as convection shifts from the sub-tropics to the equator and from the warm pool to the central and eastern Pacific.

ENSO-related shifts in tropical convection are, of course, well known. In particular, warm events are expected to cause droughts in locations surrounding the central tropical Pacific such as Indonesia and Hawaii. What is not known is the relevance of these negative
convection anomalies for the teleconnections. Branstator (1985) and Ting and Hoerling (1993) found Indonesian cooling to be an important mechanism in atmospheric general circulation model (AGCM) simulations of ENSO, but Ting and Hoerling found a much smaller role for the cooling in observations. Held and Kang (1987) and Rasmussen and Mo (1993) argued for the role of subtropical upper-level convergence in maintaining the teleconnections, but the convergence was not specifically attributed to local diabatic cooling; midlatitude transients and divergent outflow from the Tropics were also considered.

Here we use a steady linear diagnostic model to assess the relative importance of heating and cooling anomalies for ENSO teleconnections in the 0.25° eddy height field. The diagnosis is performed both on observational data and output from an AGCM integration. In both the observations and the model integration, the warm phase cooling anomalies are as important as the equatorial heating monopole in establishing the response. Our results thus suggest that the response to heating should be regarded not as a single wave train, but as a superposition of Rossby waves forced from several locations.

The remainder of this note consists of six sections. Section 2 describes the data sources used in the study and the methods for calculating diabatic heating, and section 3 presents the ENSO heating and cooling anomalies from the reanalysis and the National Aeronautics and Space Administration (NASA) Seasonal-to-Interannual Prediction Project (NSIPP) model integration. In section 4 we use a steady linear diagnostic model to assess the relative roles of heating and cooling in maintaining ENSO stationary waves in the reanalysis and the NSIPP integration. The effect of dissipation is also briefly examined using a time-marching diagnostic model, which can be integrated without horizontal diffusion. Section 5 considers separately the effect of cooling in the western Pacific (where cooling can be regarded as an integral part of coupled ENSO dynamics). Section 6 compares our results with Ting and Hoerling’s (1993) study, which found a much smaller role for western Pacific cooling in establishing the observed ENSO response. Concluding remarks follow in section 7.

2. Data and diabatic heating calculation

Observational data for this study come from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996). Monthly mean wind, height, and temperature data were obtained on 17 pressure levels and reduced to a 2.5° latitude × 5° longitude grid. In addition to the monthly mean fields, submonthly horizontal and vertical fluxes of heat and horizontal momentum were downloaded, and the forcing to the monthly mean flow anomalies by submonthly transients was calculated from the divergence of these fluxes. Monthly mean diabatic heating was calculated as a residual in the thermodynamic equation using the monthly mean wind and temperature fields, together with the submonthly thermal fluxes. A detailed description of the dataset and calculations can be found in DeWeaver and Nigam (2000).

A parallel diagnosis was performed on an AGCM integration from NSIPP. The AGCM is the atmospheric component of the coupled atmosphere–ocean–land model that NSIPP uses to produce seasonal forecasts (e.g., Pegion et al. 2000). It contains a finite-difference approximation to the primitive equations on an Arakawa C grid (2° latitude × 2.5° longitude), with σ = (level pressure)/(surface pressure) as a vertical coordinate (34 levels). Details of the model, together with an atlas of seasonal means simulated by the model, can be found in Bacmeister et al. (2000).

NSIPP data used here come from the fifth member (ens05) of a nine-member ensemble of AGCM integrations forced by observed monthly mean SSTs and sea ice for the period 1930–2000. The data were obtained from the NSIPP Web site (http://nsipp.gsfc.nasa.gov/data_req/atmos/nsipparc2.html), which provides monthly mean wind, height, and temperature fields, as well as submonthly transient fluxes of heat and horizontal momentum on 17 pressure levels.2 As with the NCEP–NCAR data, the forcing of the monthly mean flow by transients was calculated from the convergence of these fluxes.

Unlike the NCEP–NCAR reanalysis, the NSIPP Web site does not provide the vertical fluxes of horizontal momentum. However, we expect these fluxes to be small compared to the horizontal flux of horizontal momentum.
Diabatic heating for the NSIPP integration was available on the lowest 24 model sigma levels ($\sigma = 0.9925$–0.1875). In the heating dataset, the field “adv” contains the monthly averaged thermal tendency due to the adiabatic thermodynamic terms (i.e., convergence of resolved heat transport minus the work done by the pressure force) in the model’s dynamical core. On a monthly mean basis, adv balances the net heating due to all the diabatic heating parameterizations in the model, so we use $-\text{adv}$ to represent the diabatic heating for the NSIPP integration. The AGCM uses $p$, $\theta$ as its thermodynamic variable (i.e., at each time step the model calculates $\delta p$, $\delta \theta / \delta t = -\text{adv} + \text{diabatic heating}$; see Suarez and Takacs 1995), so adv must be multiplied by $(\sigma p / 1000)^{\kappa C} / p_s$ to convert it to a temperature forcing.

To eliminate small-scale orographic noise in the wind and potential temperature fields, the NSIPP AGCM uses a Shapiro filter, which is essentially an eighth-order diffusion term in the momentum and thermodynamic equations. The monthly mean output from the Shapiro filter in the thermodynamic equation is provided in the heating data (shp), and we add this field to adv to remove grid-scale noise near mountains. Thus, the NSIPP heating used in this study is computed as $Q = -(\text{adv} / p_s + \text{shp})(\sigma p / 1000)^{\kappa C}$. Note that unlike adv, the archived shp heating is not multiplied by $p_s$. We compute $Q$ in sigma coordinates and then transfer it to the 17 NCEP–NCAR pressure levels using linear interpolation in the logarithm of pressure.

Our diagnosis of the relative importance of heating and cooling for the observed ENSO response is, of course, limited by the accuracy of the residual heating calculation. In the Tropics, the residually derived diabatic heating is determined almost entirely by the divergent circulation, which is poorly observed (e.g., Newman et al. 2000) and difficult to assimilate (DeWeaver and Nigam 1997). In addition to the errors inherited from the divergence field, Hoerling and Sanford (1993) point out that substantial errors can be generated when, as in our case, the data are interpolated onto pressure levels before the residual heating is calculated.

Motivated in part by these concerns, Nigam et al. (2000) made a detailed comparison of the ENSO heating and cooling anomalies from the NCEP–NCAR reanalysis and the reanalysis of the European Centre for Medium-Range Weather Forecasts (ECMWF; see Gibson et al. 1997 for details). They found qualitative agreement between the two datasets regarding the structure of the heating anomalies. In particular, their regressions showed that the warm ENSO phase is accompanied by western Pacific cooling anomalies in both datasets, despite differences in data assimilation methods and the underlying numerical models. In fact, the western Pacific cooling anomalies are more prominent in the ECMWF reanalysis than in the NCEP–NCAR reanalysis.

The question of vertical interpolation errors is addressed by Nigam et al. (2000, their Fig. 10), in which a vertical profile of residual ENSO heating for the NCEP–NCAR reanalysis (the open squares) is plotted together with a comparable profile of diabatic heating generated by the assimilating model (the filled squares). The close agreement of the two profiles suggests that, at least for the reanalysis, relatively small errors are introduced by calculating the residual heating with vertically interpolated data. As a further check on the residual method, we computed the residual heating from the NSIPP thermodynamic budget on pressure levels and compared it to the NSIPP heating described above. The column mean of the residual heating (not shown) was in close agreement with the heating shown in Fig. 1b. Furthermore, although the vertical structure of the residual NSIPP heating was somewhat noisier than its model-calculated counterpart, the diagnostic model solutions produced with the residual heating (not shown) were virtually identical to the ones in the NSIPP AGCM diagnosis (see Fig. 3).

3. Identification of ENSO-related heating and cooling anomalies

ENSO-related anomalies are identified by regressing the variables of interest against the NCEP Niño-3.4 SST index (the average SST anomaly in 5° S–5° N, 170°–120° W; available online at ftp.ncep.noaa.gov/pub/cpc/wd52dg/data/indices/sstoi.indices). The regression is carried out on a monthly basis for the months December–March, and the period of record used for the regression, as well as the climatological basic state for linear model calculations, is December 1979 to March 2000. As discussed in DeWeaver and Nigam (2002), this period contains four warm events (1983, 1987, 1992, and 1998) and four cold events (1985, 1989, 1999, and 2000) that exceeded one standard deviation of the Niño-3.4 index for the period 1950–2000.

Figure 1a shows the mass-weighted vertical average of ENSO heating from the NCEP–NCAR reanalysis, which closely resembles the ENSO heating discussed in Nigam et al. (2000, their Fig. 4b) and DeWeaver and Nigam (2002, their Fig. 7c). As in these studies, the strong equatorial heating in the vicinity of the central Pacific ENSO SST anomalies is flanked by regions of cooling to the west, north, and south. In particular, the cooling center near the Philippines, while much smaller in size than the equatorial monopole, is equally strong, with a minimum value of $-0.8$ K day$^{-1}$. Also of interest are the cooling center to the northwest of Hawaii and the anomalous heating across Florida. Although these features extend into the midlatitudes, they have the deep vertical structure (not shown) of tropical heating rather than the surface-trapped profile of storm track heating. And while they are weaker than the equatorial center, they could be more effective at generating a rotational flow response, since the upper-level convergence as-
sociated with them is multiplied by the larger subtropical Coriolis parameter to produce a forcing term in the vorticity equation.

Figure 1b shows the ENSO heating anomalies from the NSIPP simulation, which strongly resemble the reanalysis heating in Fig. 1a. There are, however, important differences, including a much stronger equatorial center in the NSIPP output with a 1.4 K day$^{-1}$ maximum. The cooling anomalies in the Philippine/Indonesian sector are somewhat weaker and closer to the equator than in Fig. 1a, so that one might expect a weaker response to western Pacific cooling in the NSIPP data than in the reanalysis. Overall, differences between NSIPP and reanalysis heating are quite consistent with the differences between reanalysis and AGCM ENSO heating found by Nigam et al. (2000), who examined heating from an integration of NCAR’s Community Climate Model version 3 (CCM3). They found a more “Walker-like” structure in the reanalysis ENSO heating, in which enhanced convection in the central equatorial Pacific is accompanied by reduced convection in the western Pacific warm pool region. In contrast, the CCM3 heating formed a more “Hadley-like” structure, with enhanced equatorial Pacific heating accompanied primarily by cooling anomalies to the north and south. As in CCM3, the NSIPP equatorial heating is more zonally elongated, and a narrow band of cooling extends across the entire ocean basin just north of the heating. But despite their differences, the next section shows that the NSIPP model produces a successful simulation of the ENSO teleconnections. Furthermore, diagnostic modeling suggests that heating and cooling anomalies play the same complimentary roles in maintaining the NSIPP ENSO response that they do in the reanalysis.

4. Diagnosis of the eddy height response to ENSO heating and cooling

4a. Diagnosis of the ENSO response from NCEP reanalysis

To assess the importance of heating and cooling anomalies for ENSO teleconnections, we simulate the ENSO eddy height anomalies at the 0.25$\sigma$ level using a steady linear diagnostic model. The model is a linearization of the spectral $\sigma$-coordinate primitive equations about the wintertime climatology, with a basic state that includes the climatological stationary waves (a “wavy” basic state). A detailed description of the model is given in DeWeaver and Nigam (2000). To produce a complete diagnostic simulation, the model is forced with the global three-dimensional anomalies of diabatic heating and transient vorticity flux convergence. As in previous studies (e.g., Held et al. 1989; Ting and Hoerling 1993; Hoerling and Ting 1994; Peng 1995; see Trenberth et al. 1998 for additional references), we find that vorticity transients play an important role in generating ENSO teleconnections over the PNA sector.

Dissipation is always required in linear stationary wave models to produce stable, nonresonant responses to steady-state forcing, particularly when the basic state contains the climatological stationary waves (Branstator 1992). In the model responses shown here, dissipation takes three forms: a diffusive (Ekman) planetary boundary layer, a background (1000 day)$^{-1}$ vertical diffusion term as in Branstator (1992), and horizontal diffusion. The boundary layer and background vertical diffusion are applied with the same parameters as in DeWeaver and Nigam (2000). Horizontal diffusion is applied in the vorticity and divergence equations with a coefficient of $10^6$ m$^2$ s$^{-1}$, and $2 \times 10^6$ m$^2$ s$^{-1}$ in the thermodynamic equation. Following Hoerling and Ting (1994), the extra diffusion in the thermodynamic equation is intended to represent the dissipative effect of submonthly thermal transients.

Figure 2a shows the eddy height anomalies from the NCEP–NCAR reanalysis, which include anticyclones straddling the equatorial heating, a trough extending across the southern United States from the Gulf of Alaska, and a high over Canada. The ENSO anomalies also have a strong zonally symmetric component which, though not shown here, is documented in DeWeaver and Nigam (2002; see also Yulaeva and Wallace 1994).

The diagnostic simulation, shown in Fig. 2b, captures all of the prominent features in Fig. 2a, including the subtropical anticyclones, the Aleutian trough, and the Canadian ridge, although with some errors. The chief deficiency in this simulation, as with the diagnostic simulation from NSIPP data, is that most of the features are too weak. For example, the northern subtropical anticyclone has a maximum amplitude of 28 m, while its simulated counterpart has a maximum amplitude of 18 m. The weakness of the simulated features is primarily a consequence of the rather high values of dissipation used in the model. Strong dissipation is necessary to prevent resonance effects from corrupting the diagnostic solution, as noted by Branstator (1992). Ting and Yu (1998) argue that dissipation serves as a proxy for nonlinear effects. However, the resemblance between Figs. 2a and 2b, which have an area-weighted spatial correlation of 0.77, is sufficient to warrant further examination of the dynamics of the linear model response.

Since the model is linear, the response can be decomposed into components forced by the global heating anomalies (Fig. 2c), positive (Fig. 2d) and negative (Fig. 2e) heating anomalies, and the equatorial monopole (Fig. 2f). The difference between Figs. 2b and 2c reflects the influence of the transient vorticity fluxes, which are responsible for the high over Canada, about half of the amplitude of the Aleutian trough, and one-third of the amplitude of the northern subtropical anticyclone. The finding of an important role for transient vorticity fluxes is consistent with previous studies, as discussed above.

Comparison of Figs. 2c–e shows that heating and
cooling anomalies make comparable contributions to the eddy height response. The area of lowered eddy height in Fig. 2c extending westward from the date line to about 90°E is essentially a local response to anomalous cooling in the western Pacific (see also Fig. 5). Cooling also makes a substantial contribution to the Aleutian trough and to a similar trough in the Southern Hemisphere extratropics.

Of particular interest is the role of cooling in determining the location of the subtropical anticyclones. Gill’s (1980) simple model suggests that an equatorial heat source should produce a westward-shifted anticyclone pair. Studies like Sardeshmukh and Hoskins (1985) and Hendon (1986) invoked nonlinear processes to explain the placement of the observed anticyclones at the longitude of the equatorial heating. In our diagnostic simulation, equatorial heating by itself (Fig. 2f) does produce westward-shifted anticyclones; the placement of the anticyclones at the longitude of the heat source is largely the result of the cooling anomalies. It must be noted, however, that the use of the full wintertime basic state circulation is required for an accurate simulation of the anticyclones and other circulation anomalies. A model with a resting basic state, such as the Gill model, will not produce the correct anticyclone placement even when forced by the full suite of ob-

Fig. 2. Diagnostic simulation and decomposition of ENSO stationary waves at 0.25σ from NCEP–NCAR reanalysis. (a) ENSO eddy height anomalies from the reanalysis, obtained by linear regression and (b) eddy height simulation from the steady linear diagnostic model. Model response when forced by (c) all diabatic heating anomalies and (d) only the positive heating anomalies. Model response when forced by (e) the negative heating anomalies and (f) the positive heating center in the vicinity of the ENSO SST anomalies. Contour interval is 5 m, with dark (light) shading for positive (negative) anomalies in excess of 5 m, and zero contours suppressed in (c)–(f). In (d)–(f), regions where positive heating anomalies exceed 0.2 K day$^{-1}$ are enclosed in solid double lines, and regions of negative heating in excess of 0.2 K day$^{-1}$ are enclosed in dashed double lines.
observed December–January–February–March (DJFM) heating and cooling anomalies. As in previous studies (e.g., Ting and Hoerling 1993), our solution degrades even if the model is linearized about a zonally symmetric basic state.

b. **Response to ENSO heating and cooling in the NSIPP AGCM**

Our confidence in claiming a strong role for ENSO cooling is limited by the lack of definitive observations of tropical divergence, as discussed in section 2. The high-quality simulation of ENSO teleconnections by the NSIPP AGCM can offer a valuable second opinion, since the diabatic heating is known exactly in the model world. An additional source of uncertainty was identified by Ting and Sardeshmukh (1993) and Branstator (1992), who showed that the response of a linear stationary wave model to tropical heating can be sensitive to small changes in the climatological basic state. A diagnosis of the ENSO response in the NSIPP AGCM thus serves as a check on the sensitivity of our results to the details of the climatological flow (see Bacmeister et al. 2000 for an in-depth comparison of the AGCM and reanalysis climatologies). Further motivation for the NSIPP diagnosis comes from Ting and Hoerling (1993), who showed that the dynamics of the ENSO response in an AGCM can be very different from those of the observed teleconnections (see section 6 for discussion).

The NSIPP AGCM diagnosis is shown in Fig. 3. Figure 3a shows the 0.25σ eddy height pattern obtained by regressing the simulated monthly height against the Niño-3.4 index. Evidently, the NSIPP model produces a fairly realistic simulation of the ENSO pattern in upper-level eddy heights (Fig. 2a), capturing all the significant centers of action, including even subtle features such as the narrow high over central Asia. Deficiencies are also present in the NSIPP integration, most notably the lack of symmetry in the subtropical anticyclones. The area-weighted spatial correlation between the 0.25σ eddy heights from the NCEP–NCAR reanalysis (Fig. 2a) and the NSIPP ensemble member in Fig. 3a is 0.82. However, it would be more appropriate to evaluate the performance of the NSIPP AGCM by comparing the ensemble-mean ENSO response to the observations, a comparison which is beyond the scope of the present study (see Pegion et al. 2000 for comparisons of this sort).

Figure 3b shows the eddy height response to heating and vorticity transients in the linear stationary wave model. As in the diagnosis of observed eddy heights, the linear model response captures the features of interest, but with reduced amplitude. The El Niño anticyclones in the diagnostic simulation are somewhat too symmetric about the equator, and the southern anticy-
clone is $10^\circ$ too far west. Despite these deficiencies, the area-weighted correlation between the NSIPP anomalies and the diagnostic simulation is 0.78.

The response of the stationary wave model to global diabatic heating anomalies is shown in Fig. 3c, while the response to the negative heating anomalies is shown in Fig. 3d. Comparing these panels with their counterparts for the NCEP–NCAR reanalysis in Figs. 2c and 2e, we conclude that the relative contributions of positive and negative heating anomalies in maintaining stationary wave anomalies is the same for the NSIPP AGCM data as it is for the reanalysis. To the extent that can be determined by linear diagnostic modeling, the AGCM ENSO response is consistent with observations both in structure and dynamics.

c. Diagnosis with a time-marching diagnostic model

The diagnostic simulation is weak in amplitude, reaching only 50%–75% of the target values in both Figs. 2b and 3b. However, the amplitude deficiency—particularly in the region of the subtropical anticyclones—is not unique to our simulation since such models generally produce underestimated amplitudes in rigorous validation; a recent example is the modeling analysis in Watanabe and Jin (2003 their Figs. 3c and 5b, note that their column heating has a maximum of 3 K day$^{-1}$). This model behavior results at least in part from the excessive thermal and momentum dissipation needed for obtaining truly steady solutions in a dynamical environment permissive of hydrodynamic instabilities, wave guides, and potential resonance. A time-marching primitive equation model was developed in part to circumvent the need for such dissipation. In this modeling framework, we examine solutions at an early stage before they are overwhelmed by dynamical instabilities and/or resonant interactions. These models can, moreover, offer insight into the development of ENSO’s mid-latitude response.

The model was adapted from the idealized dry AGCM of Held and Suarez (1994). Model variables are discretized on an Arakawa C grid with $4^\circ$ meridional $\times 7.5^\circ$ zonal resolution; and we use the same 15 $\sigma$ levels as in the steady linear model. Time integration is carried out with an explicit leapfrog time step of 10 min. (from time $n-1$ to time $n+1$) and Brown–Campana pressure gradient averaging [Suarez and Takacs (1995) gives a complete description of the dynamical core]. The model is forced with the same heating and transient fluxes used in Fig. 2b, except that transient forcing is applied to the $u$- and $v$-momentum equations instead of the vorticity and divergence equations. Following Jin and Hoskins (1995), we also apply steady momentum and heat sources to the model equations that balance the basic state flow. In addition, the zonal-mean winds and potential temperature are reset to their basic state values at each time step, so that, as with the steady linear model, the model solves for only the eddy components of the flow anomalies.

The integration shown here (Fig. 4) was performed without the horizontal diffusion terms used in the steady linear model. However, the boundary layer and background vertical diffusion terms from the steady model were applied. In addition, the dynamical core requires a polar Fourier filter to satisfy the Courant–Friedrichs–Lewy (CFL) condition at high latitudes and an eighth-order Shapiro filter (which damps the $2\Delta$ wave with an $e$-folding time of 1.5 h) to control grid-scale horizontal noise. With this level of dissipation the model does not produce steady solutions. But as in Jin and Hoskins (1995), resonance and baroclinic transients do not dominate the solutions until about day 15.

Figure 4 shows the model solution obtained with full NSIPP forcing (ENSO heating and momentum transients) at day 12. The day-12 solution compares favorably with the solution from the steady linear model, and amplitudes in the Tropics and subtropics are now somewhat stronger (see Fig. 2b), but amplitudes in the high latitudes (over the Aleutians and northern Canada) are already a bit too strong. The mature response in nature would presumably be limited by further physical interactions and feedbacks, particularly at these sites. While the time-marching model supports the notion that strong dissipations are a factor in the amplitude deficiencies, more work is required to fully resolve the issue.

5. The role of western Pacific cooling

It is possible that the protomodel is correct on a more fundamental level, since the chain of causality could still begin with the local equatorial convection response to ENSO SSTs. In this scenario, the equatorial convection gives rise to compensating subsidence in the surrounding regions, which suppresses convection and produces the cooling anomalies shown in Fig. 1. In that case, the cooling anomalies could simply be added to the protomodel as another mechanism through which
the equatorial monopole generates remote flow perturbations. However, the recent papers of Su et al. (2001) and Watanabe and Jin (2003) present modeling evidence that the equatorial ENSO SST anomaly can at best produce the cooling anomalies found directly to the north and south of the equatorial monopole. Cooling in the western tropical Pacific is found to be a consequence of the local reduction of SSTs (in the warm ENSO phase) in that region. This local SST reduction is an integral part of the ENSO phenomenon, presumably related to the relaxation of the east–west thermocline tilt associated with ENSO events.

Following this line of reasoning, we examine separately in Fig. 5 the diagnostic model’s response to cooling when the cooling in the central subtropical north Pacific (11°–55°N, 170°E–120°W, roughly centered on Hawaii) is omitted (Fig. 5a) and when only Hawaiian sector cooling is retained (Fig. 5b). While our definition of this region is somewhat arbitrary, it is an area where cooling anomalies could reasonably be incorporated into the protomodel as an intermediary between equatorial heating and extratropical teleconnections.

Even discounting the Hawaiian sector cooling, we still find a substantial response to cooling, including a strong local response in the western Pacific and contributions to the Aleutian low and the subtropical anticyclones. Essentially the same diagnosis can also be made using the NSIPP data, as can be seen in Figs. 5c, d. Figure 5 thus argues against the notion that ENSO cooling should be regarded as merely an enhancer of the wave train forced by the equatorial monopole.

The day-12 solutions for the time-marching model forced by equatorial Pacific heating and western Pacific cooling are shown in Figs. 5e,f. Like the steady linear solutions, they indicate the considerable significance of western Pacific diabatic cooling anomalies in the development of ENSO’s midlatitude response. Key features of the upper-tropospheric ENSO response include the southeastward extension of the Pacific jet (with attendant storm track perturbations and feedbacks), a trough in the vicinity of the Aleutian Islands, and a ridge over the Hudson Bay and Great Lakes region. The western Pacific cooling anomalies evidently make a significant contribution in all these regions. They could, moreover, induce diabatic cooling in the subtropical central Pacific (Hawaiian sector) through anticyclonic development in that region. From vorticity balance considerations, the western flank of a subtropical anticyclone should be a region of descending motion, which would suppress local convection; Hawaii does experience drought conditions during El Niño winters. If our portrayal of the response of ENSO-related western Pacific diabatic cooling anomalies is realistic, these anomalies could also be implicated in the generation of anomalous coastal rainfall over southern California during El Niño winters, a region under the eastern flank of the above induced anticyclonic circulation and thus under ascending motions. The previously noted contribution of the western Pacific cooling anomalies in the correct placement of the subtropical anticyclones also stands out in the time-marching model solutions.

Further analysis and modeling are clearly needed to corroborate our findings on the role of ENSO-related diabatic cooling in the western tropical Pacific in setting up the ENSO teleconnections in winter circulation and hydroclimate—a role that has not been fully appreciated, particularly, in conceptual models.

6. Comparison with Ting and Hoerling (1993)

Our diagnosis shows a prominent role for western Pacific cooling in both the NSIPP AGCM and the NCAR–NCEP reanalysis. This finding differs from the earlier study of Ting and Hoerling (1993), in which anomalous cooling near Indonesia was quite important for the ensemble-mean response of an AGCM forced by El Niño SSTs, but marginally significant for the observed response. There are, of course, a variety of possible reasons for the difference between our results and theirs, including discrepancies in the vertical and horizontal structure of the residually diagnosed heating and differences in the period of record for the climatology (they use DJF 1986/87–1990/91).

One factor that may be particularly significant is the different anomalies examined in the two studies: we use a multiyear regression to identify the ENSO anomalies, while Ting and Hoerling looked specifically at the warm event of 1986/87. When the observed and simulated 200-mb streamfunction anomalies in their study (their Fig. 1) are compared with the regressed ENSO streamfunction anomalies corresponding to our Fig. 2, the best match is between the regressed anomalies and their ensemble AGCM simulation. As in the ensemble simulation, the regressed streamfunction shows anticyclones straddling the equator with centers east of the date line, together with a low near the Gulf of Alaska. In contrast, the observed 1986/87 anomalies show a poorly defined anticyclone pair and a wave train with centers over the western subtropical Pacific (a high), the end of the Aleutian chain (a low), and Hudson Bay (a high). These features are reproduced in Fig. 6, which shows a regression of the 200-mb streamfunction against the Niño-3.4 index (Fig. 6a) and the streamfunction anomalies for DJFM 1986/87 using the 1980–2000 climatology (Fig. 6b). In this figure we use an orthographic projection (120°W–60°E) for comparison with the simulated and observed anomalies in their Fig. 1 (note that their plot is centered at 20°N while ours is centered at 0°).

As Ting and Hoerling point out, the observed Aleutian low for the 1986/87 event is displaced about 30° to the west of the corresponding feature in the ensemble simulation. The 1986/87 Aleutian low also has a pronounced westward shift compared to its counterparts in the warm phase ENSO height composites of Hoerling et al. (1997, their Fig. 3a) and DeWeaver and Nigam (2002, their Fig. 4a). These comparisons suggest that
either the 1986/87 event was unusual in some way or that factors other than ENSO played a role in generating the flow anomalies of that winter. The latter possibility is further suggested by Ting and Hoerling’s finding that changes in the zonal-mean flow were influential in generating the 1986/87 stationary wave anomalies (their Fig. 8f). Hoerling et al. (1995) showed that such zonal–eddy coupling is not a prominent mechanism for generating the ENSO response.

A possible interpretation of the difference between the present results and those of Ting and Hoerling is suggested by Kumar and Hoerling (1997). They found that AGCMs forced by observed SSTs tend to produce essentially the same pattern for all warm ENSO events, despite the large interevent differences found in observations. This is presumably a consequence of the large internal variability of the atmosphere. Thus, the closer agreement between Ting and Hoerling’s single-event en-
semble simulation and our multi-event ENSO regression could be due to the internal variability that occurred during the winter of 1986/87. By failing to simulate the non-ENSO variability of that winter, it is possible that the simulation has inadvertently produced a more accurate representation of the basic dynamics of the canonical ENSO response.

7. Concluding remarks

Two factors contribute substantially to the difference between our diagnosis and the classical teleconnection protomodel. First, the reanalysis data allow us to construct diabatic heating anomalies that are more detailed than those available for the early teleconnection studies and include a larger sample of events. Second, the ENSO teleconnection pattern (Fig. 2a) does not strongly resemble a wave train propagating along a great-circle route (the resemblance becomes even weaker when the zonal-mean anomalies are included). Early studies like Horel and Wallace (1981) did not distinguish between the ENSO response and the PNA teleconnection pattern, which has a more wavelike appearance (e.g., Dai 1999; Straus and Shukla 2000). The theoretical advances of Hoskins and Karoly (1981) also produced an expectation that a simple wave train should appear in the ENSO response.

Given its elegance and simplicity, it is understandable that the protomodel continues to be the dominant—perhaps the only—conceptual model of ENSO teleconnections. It remains to be seen whether the results presented here can be incorporated into a conceptual model of comparable simplicity.

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