A Time-Varying Greenhouse Warming Pattern and the Tropical–Extratropical Circulation Linkage in the Pacific Ocean

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ABSTRACT

Recently, Cai and Whetton provided modeling evidence that the greenhouse warming pattern has undergone a systematic change from a pattern with maximum warming in subtropical and mid- to high latitudes to one that is El Niño–like from the 1960s onward. They suggest that the mechanism for the change is the transmission of the large extratropical warming to the equatorial east Pacific via modeled tropical–extratropical Pacific circulation pathways. The present study addresses several associated issues. How is the systematic change manifested in empirical orthogonal functions? How do the meridional heat balances respond to the systematic change? Does the proposed mechanism operate in the absence of greenhouse forcing? It is shown that the warming signals are represented by two empirical orthogonal functions, the first of these reflecting a long-term trend in the period considered, and the second showing the change in trend from the 1960s onward. Consistent with the time-varying warming pattern, the relative importance of various heat exchange processes in the tropical Pacific Ocean also undergoes systematic changes. Prior to the 1960s, advective heat flux from the extratropics is the heat source for warming the tropical subthermocline (80–270 m). This subthermocline warming weakens the thermocline and reduces the diffusive heat transfer down through the subthermocline. From the 1960s onward, as substantial subthermocline warming upwells, the El Niño–like pattern develops, strengthening the thermocline; consequently, the downward diffusive heat transfer to the subthermocline enhances reversing the trend prior to the 1960s, and eventually becomes the dominant source for subthermocline heating. The dynamical process, whereby extratropical anomalies are transmitted to the Tropics, operates in a run without external forcing, in association with a mode of ENSO-like interdecadal oscillation. In the equatorial central-eastern Pacific, the associated anomalies upwell and initiate an ocean–atmosphere feedback that changes the equatorial west–east sea surface temperature gradient and easterly winds, reinforcing the upwelled anomalies. The commonality of the modeled interannual ENSO cycles and the interdecadal ENSO-like variability is also discussed.

1. Introduction

The global mean surface temperature has increased about 0.3°–0.6°C since the late nineteenth century with rapid warming occurring over the past few decades [Intergovernmental Panel on Climate Change (IPCC) 1995]. In the Pacific, the spatial pattern associated with the recent warming is El Niño–like, in terms of a west–east sea surface temperature (SST) gradient along the equator, as greater warming occurs in the east than in the west. This rapid warming is often referred to as the “regime shift” (Nitta and Yamada 1989; Trenberth 1990; Graham 1994, 1997; Lau and Weng 1999). There have been more frequent severe El Niño–Southern Oscillation (ENSO) events over the past few decades (Trenberth and Hoar 1996, 1997). This raises the question of whether these features are manifestations of human-induced “greenhouse” warming, or simply a warm phase of a multidecadal ENSO-like oscillation (Gu and Philander 1997; Zhang et al. 1997; Polland et al. 1998; Latif 1998; Latif et al. 1997; Zhang et al. 1998; Kleeman et al. 1999). An El Niño–like warming pattern, that is, the equatorial eastern Pacific warms more than the equatorial western Pacific, has been produced by many climate models (Meehl and Washington 1996; Knutson and Manabe 1995; Timmermann et al. 1999). However, such a pattern is contradictory to that predicted by other studies (Cane et al. 1997; Clement et al. 1996; Seager and Murtugudde 1997), which suggest a response with slower warming in the equatorial eastern Pacific. Understanding the cause for these seemingly conflicting results is therefore important because these patterns could provide useful fingerprints of greenhouse warming, and can be expected to significantly influence other changes in climate.

The process that underpins an El Niño–like warming pattern in earlier coupled models is that the equatorial west Pacific is so warm that even modest additional warming would lead to a “cloud-shielding thermostat” effect (Meehl and Washington 1996; Ramanathan and
Collins 1991), reducing incoming solar radiation at the surface and inhibiting further warming. The difference in the strength of this feedback between the equatorial east and west Pacific will lead to a weaker warming in the west, generating an El Niño–like warming pattern, again in terms of the zonal SST gradient along the equator, with slower warming in the east.

The argument for a La Niña pattern is that the upwelling in the equatorial east Pacific will weaken the warming in the region, so that the strongest warming will occur in the west Pacific. This initiates an ocean–atmospheric feedback with stronger easterly winds along the equator, and stronger equatorial upwelling in the east Pacific, further reducing the warming in the equatorial east Pacific. This “dynamical thermostat” process will generate and reinforce such a warming pattern.

Several studies have employed empirical orthogonal function (EOF) analysis to identify leading modes of variability of observed SSTs. Often, a mode (usually the first EOF) with rising temperature trend is identified (e.g., Folland et al. 1998; Lau and Weng 1999; Cai and Whetton 2000). However, the spatial pattern of the mode varies significantly from one study to another. Folland et al. (1998) showed a trend mode (their EOF1 mode) for the period from 1900s onward that has maximum warming in the mid- to high latitudes, whereas an El Niño–like pattern is obtained from EOF analysis on SST anomalies since 1950 (Lau and Weng 1999). Can these results be reconciled? Does the first EOF alone pick up the global warming signal? Attributing the warming trend and the associated EOF to a specific cause implicitly assumes that the spatial pattern of the trend is time invariant. How does a time-varying warming pattern manifest itself in EOF analysis?

Based on results from simulations using the Commonwealth Scientific and Industrial Research Organisation (CSIRO) coupled atmosphere–ocean model, Cai and Whetton (2000) propose that the (El Niño like) warming pattern in the Pacific Ocean since the 1960s is a dynamical consequence of the warming pattern prior to the 1960s. The earlier pattern in their model has the largest warming in the subtropical and mid- to high latitudes. They demonstrate that the switch in warming pattern is present in each of an ensemble of three greenhouse warming simulations in which the model is forced by increasing levels of greenhouse gases as observed (1880–1990) and according to IPCC scenario IS92a (1990–2100; Houghton et al. 1992). The CO2 concentration doubles at about year 2030. Two of the warming simulations start several hundred years apart in the control simulation, and the other is carried out on a slightly different version of the model with an improved sea ice–ocean interaction process. In these versions of the coupled model, the ocean incorporates the Gent and McWilliams (1990) eddy parameterization, allowing the use of zero horizontal diffusivity.

In the first of three warming simulations, already reported by Hirst (1999), the atmospheric equivalent CO2 concentration is held constant for about 700 yr following tripling in year 2080. In the present study we concentrate on the period 1880–2030 and report results from all the three experiments for the period 1880–2030. In each experiment the coupled model is run with fixed CO2 at a level equivalent to that of year 1880 for 30 yr (for comparison, the CO2 level in the control simulation is also at the 1880 level; in any case, in the model it is the ratio of the increased CO2 level over that of the control climate that is important). Then the CO2 starts to rise from the level of year 1880. Our analysis uses outputs for the 30-yr period prior to the increase of CO2.
and for years 1881–2060. The extended data facilitate filtering to remove variability on timescales of up to 60 yr, while allowing a projection to year 2030 at which time the CO₂ has approximately doubled.

3. Observed and modeled time-varying warming pattern in EOF analysis

We apply EOF analysis to observed SSTs to illustrate the way in which a time-varying warming pattern manifests in EOFs. The first and the second EOF modes of the observed annual mean SST anomalies from 1881 to 1997 are displayed in Fig. 1. The observed SST data are those of an updated dataset of Rayner et al. (1996). The anomalies are constructed relative to the climatological mean of the period 1950–97 and are filtered to retain variability on timescales longer than 20 yr. It is appropriate to point out that the filtering procedure will cause the “regime shift” of the 1976 to be smoothed out over 20 yr. The two modes account for 76% and 9% of the total variance on the retained timescales. Temporal evolution of EOF1 shows a warming trend (Fig. 1b) and the spatial pattern shows that in the Pacific Ocean larger warming takes place in the subtropical and mid- to high latitudes than in the Tropics. The second EOF mode is La Niña–like (Fig. 1c), and the evolution (Fig. 1d) shows that there is an upward trend during the period 1900s–50s indicating development of a La Niña–like pattern. Since the 1950s there has been an downward trend, indicating the development of an El Niño–like pattern. Lau and Weng (1999) apply EOF to SST observations since 1950 and obtain an El Niño–like warming trend pattern. Is this upward trend since the 1950s influenced by the forcing that generates EOF1?

When the same analysis is applied to outputs of three individual greenhouse experiments, similar features emerge. Figure 2a shows the averaged spatial pattern over the three EOF1 patterns. The pattern correlation coefficients between patterns from the individual experiments are very high, in the range of 0.89–0.94. Figure 2b shows the three time series of temporal coefficients. Figures 2c and 2d are the corresponding pictures for EOF2 of the three experiments. In the Pacific, the trend mode (Fig. 2a) features larger extratropical warming than that in the Tropics, and the EOF2 mode (Fig. 2c) is La Niña–like but contains the evolving feature of initial development of a La Niña–like pattern, that changes to El Niño–like from the 1960s onward (Fig. 2d). We will discuss the model ENSO cycle and ENSO–like interdecadal variability in section 5. Here the fact that EOF2 in all three experiments share similar temporal evolution, and that these responses are in turn similar to the observed, suggest that EOF2 may be influenced by the forcing of EOF1.

Fig. 1. Results from EOF analysis on observed SST anomalies for the period 1881–1997 retaining variability on timescales longer than 20 yr. Shown are patterns of weights for (a) EOF1 and (b) the associated temporal coefficient, (c) EOF2 and (d) the associated temporal coefficient. They account for 76%, 9%, respectively, of the total variance on these timescales.
Applying EOF analysis to modeled SST anomalies for the period from year 1881 to year 2030 when the model CO₂ concentration approximately doubles yields a long-term trend mode. The pattern is similar to that obtained from results for stabilized CO₂ condition upon trebling (Whetton et al. 1998). The associated temporal coefficient shows that prior to the 1960s the trend rate is fairly low. The initial La Niña–like pattern in the Pacific Ocean appears in EOF2 (Figs. 3c and 3d), and the pattern changes to El Niño–like after about the 1960s.

The time-varying pattern can be confirmed by modeled linear trend maps averaged over the three experiments (see Figs. 2a–c of Cai and Whetton 2000) and the trend pattern of the observations (see Figs. 2d,e of Cai and Whetton 2000). Both the modeled and the observed show an initial warming pattern with larger warming in the extratropics, and a recent pattern that is El Niño–like. These results remain largely unchanged if the dividing year for the two periods is year 1960 or year 1940, or if year 1997 is excluded in the analysis. Maps of trend of mean sea level pressure (MSLP) and rainfall (figures not shown) show consistent patterns with those of SST anomalies. During the initial La Niña–like phase, MSLP decreases over the Australasia region and increases in the region near Tahiti; and rainfall in regions such as Indonesia and eastern Australia increases, and vice versa during the El Niño–like phase. The rainfall trend is similar to the trend map from observations for the period 1971–90 compiled by Morrissey and Graham (1996).

An important question arises. Is it possible that the greenhouse warming pattern in the model is simply a fixed pattern of EOF1 throughout, but it is suppressed initially due to a contrary trend arising through natural multidecadal-scale variability? This seems unlikely, given that the time evolution is shared by all three simulations, and it has been investigated further by Cai and Whetton (2000) using the 1000 yr of the control simulation. Cai and Whetton (2000) take sixteen, partially overlapping, 150-yr epochs from the control simulation, and add to data of each epoch a warming signal for 1881–2030 derived from a greenhouse simulation. The added warming signal is fixed in time in terms of pattern but grows in magnitude in line with the global average warming. The warming signal is obtained from the first EOF (already shown in Fig. 3) for the period of 1881–2030 from one of the greenhouse simulations. To assist in making a comparison between these synthetic greenhouse cases and the greenhouse simulations, an index, $R$ is formed, designed to measure the time-evolving behavior identified above. Here $R$ is defined as the ratio of $r_{\text{post}}/r_{\text{pre}}$, where $r_{\text{pre}}$ is pre-1950 linear warming rate for the eastern Pacific (EP) region (east of the inter-
national date line, 15°S–15°N) normalized by the non-EP warming rate for the same period, and \( r_{\text{post}} \) is the equivalent for the post-1950 period. If the warming rate in the EP region stays in the same proportion to the non-EP warming rate over the two periods, \( R \) equals one. When \( R \) is greater than one it means that the EP warming relative to that in the rest of the oceanic regions is slow in the pre-1950 period and accelerates in the post-1950 period. The distribution of \( R \) is plotted in the present study in Fig. 4. The value of \( R \) in the synthetic greenhouse cases is one on average, as would be expected because the imposed greenhouse signal has an invariant pattern, but has a range of 0.8–1.2 due to differences in natural variability between the cases. The three greenhouse simulations produce \( R \) values well beyond this range (1.7–2.2 when calculated using data to 1997; or 1.5–1.9 using data to 2030). The difference in these ranges indicates that the initial delay in warming in the EP region and the later acceleration is part of the greenhouse response of the model. The observations, with an \( R \) value of 3.9, also share this behavior. The result suggests that the change in warming pattern is statistically significant and is unlikely to be caused by internal variability.

4. The mechanism for the time-varying warming pattern and the associated heat balances

Both the observed and the modeled results show that initially larger warming takes place in the subtropical and mid- to high latitudes. Cai and Whetton (2000) suggest that the extratropical Pacific warming is transmitted to the Tropics via the tropical and extratropical oceanic circulation pathways identified by modeling and observational studies (Lu and McCreary 1995; Liu and Philander 1995; Lu and McCreary 1995; Deser et al. 1996; Zhang et al. 1998). In essence, water in the extratropics subducts, and flows equatorward and westward at depth, and equatorward in the western boundary current region and then eastward in the equatorial undercurrent to the subsurface equatorial east Pacific. Does the coupled model produce these pathways? Does the meridional oceanic heat balance also display corresponding changes?

a. The model circulation pathways

The modeled Pacific circulations averaged over 100 yr of the control run are presented in Figs. 5a and 5b, showing annual mean near-surface velocities averaged over the first three levels (levels 1–3, upper 80 m), and averaged over levels 4–7 (80–270 m), respectively. The low resolution leads to the absence of the north equatorial countercurrent, but other circulation features are reasonably simulated. Near the surface (Fig. 5a), the gyre circulation in the Pacific Ocean is evident, which extends from the westward flow of the Tropics and turns northward along the western boundary. The flow then veers eastward in the latitude band 30°–45°N, and its
speed reduces in the longitude band east of the international date line where the surface water subducts. At the subsurface (Fig. 5b), the subducted water flows southwestward, equatorward in the western boundary current, and some turns eastward in the equatorial undercurrent eventually reaching the equatorial east Pacific. The pathways in the Southern Hemisphere are somewhat different from those in the Northern Hemisphere. For example, previous studies have shown that a great deal of the water that subducts reaches the equator directly without first flowing into the western boundary current region (e.g., Johnson and McPhaden 1999). This feature appears to be reflected in Fig. 5b. The modeled undercurrent has a maximum velocity at level 6, at an averaged depth of 185 m. The velocities are weak compared with observations as would be expected from the low resolution of the model, but in terms of transport, it is about 25.7 Sv (10^6 m^-1) compared with about 30 Sv in observations. Thus the basic circulation pathways are reproduced by our coupled model. Other interesting features include poleward Ekman flow out of the Tropics (Fig. 5a), the water flowing out of the equatorial region through the Indonesian throughflow being fed by water from the south in the thermocline (Fig. 5b), and the transport to the Indian Ocean through the Indonesian throughflow taking place mostly in the surface as discussed by previous studies (e.g., Rodgers et al. 1999).

b. The generation of time-varying warming pattern

The equatorward transfer of extratropical warming to Tropics is plotted in Fig. 6a, which shows warming averaged over a 30-yr period (centered at the indicated decade) and over the three experiments, as a function of depth. The anomalies are averaged over the longitude band of 120Â°E–120Â°W. Consistent with EOF1 pattern (Figs. 2a and 3a), large warming at the subsurface originates from the latitude band about 25Â°–40Â° and flows equatorward to the Tropics. As extratropical warming proceeds, the subsurface warm anomalies intensify. After the 1950s, the growth of the anomalies at the Tropics accelerates.

The process responsible for the acceleration of tropical warming is outlined in Fig. 6b, which shows the evolution of the simulated warming along the equatorial Pacific as a function of depth, again averaged over the three experiments. Until the 1950s, there is stronger surface warming in the west than in the east, combined with strengthening equatorial easterly winds (Fig. 3 of Cai and Whetton 2000). This is consistent with what may be expected from theoretical studies, which have suggested that upwelling in the east equatorial Pacific would delay warming in that region relative to the west Pacific and induce atmosphere–ocean feedbacks with increasing easterly winds and upwelling, and a La Niña–like state. However, from the 1950s to the 1980s, warming in the equatorial east Pacific accelerates, indicating development of a pattern that is more El Niño–like. This change in the surface warming pattern has its origin in the strong warming that appears early at the subsurface in the west Pacific. As discussed in the Cai and Whetton (2000) study, this warm anomaly grows and extends eastwards with the equatorial undercurrent to the east Pacific where it upwells, and initiates the development of the El Niño–like warming pattern.

c. Heat budget of the tropical Pacific

To understand the associated heat exchange processes between Tropics and extratropics and between upper and lower east Pacific waters, we calculate various heat flux terms that contribute to temperature change in the tropical Pacific area 15Â°S–15Â°N. To resolve the subsurface warming, calculations are made for two layers in this region—one extends from the surface to 80 m and one from 80 to 270 m (see Fig. 7). For each layer the heat exchange may take place via four boundary surfaces, north, south, top, and bottom (a positive value means a heat gain to the layer). Note that the Indonesian throughflow passage is treated as a part of the south boundary but the boundary of this passage is along the latitude
Fig. 5. Climatological mean fields of vertically averaged ocean current vectors (cm s\(^{-1}\)), (a) over the upper ocean (0–80-m depth) and (b) over the subsurface ocean (80–270-m depth). These fields are obtained by averaging over 100-yr annual mean fields of the control run.

of about 5°S rather than 15°S. The east and west boundaries are land surfaces and no exchange is allowed.

For the upper layer, we write the heat budget as

\[ F_{St} = F + F_{Adv} + F_{VDif} + \text{Others}. \]  

(1)

Here, \( F_{St} \) is the heat storage rate, \( F \) is the surface heat flux, \( F_{Adv} \) is the divergence of advective heat flux including both the horizontal and vertical advective fluxes (Fig. 7), and \( F_{VDif} \) is the divergence of vertical diffusion.

For the upper layer, the budget equation (1) can be rewritten as

\[
\int_{-80}^{0} C_p \rho T_z \, dz = F + \int_{-80}^{0} C_p \rho \nabla \cdot (TV) \, dz \\
+ C_p \rho A_v T_{z=80} + \text{others}. \]  

(2)

Here \( V \) is velocity vector and includes the large-scale and eddy-induced components, \( \rho \) is the density of seawater, \( C_p \) is the specific heat, and \( A_v \) is the vertical diffusivity coefficient. The above equation states that the heat that is needed to change the oceanic heat storage rate is provided by the sum of surface heating, divergence of advective fluxes, divergence of vertical diffusive flux, and others, which may include convective and isopycnal fluxes. As will be shown, these other terms are very small in the tropical Pacific.

A similar equation can be written for the lower layer:

\[
\int_{-270}^{-80} C_p \rho T_z \, dz = \int_{-270}^{-80} C_p \nabla \cdot (TV) \, dz \\
+ \int_{-270}^{-80} C_p \rho (A_v T_z) \, dz \\
+ \text{others}. \]  

(3)

For the lower layer, there is no direct surface flux and the heat source derives from the oceanic processes such
as advection and vertical diffusion. Each term at each grid point for each year is calculated from the model outputs of the three experiments. The calculation starts from a time 30 yr prior to the commencement of the warming experiment (year 1851), and for each year, integration of each term over the above area is then carried out. The time series of each term for the upper and lower layers (filtered to retain variability on timescales longer than 60 yr), averaged over the three experiments are shown in Figs. 8a and 8b, respectively. The balance in each of the three experiments is displayed in Figs. 8c and 8d; it is seen that the fluctuations of each term in each experiment are similar.

Prior to the 1960s, the upper layer loses heat via the divergence of advective heat fluxes (Fig. 8a, dotted-dashed curve). Much of the advective heat that is lost is compensated by an increasing surface heating (Fig. 8a, dashed curve) and by a reduction in diffusive heat penetration (Fig. 8a, long dashed) to the underlying layer. Since the 1960s, however, the heat loss due to the divergence of advective fluxes starts to decrease, reversing the previous trend. In the meantime, diffusive heat loss from the upper to the lower layer increases and eventually becomes the major process for transferring heat away from the upper layer to the lower layer. The sum of these three terms is almost identical to the rate of heat storage, suggesting the neglected terms are small. For the lower layer (Fig. 8b), prior to the 1960s, it is the divergence of advective fluxes that is mainly responsible for the warming. After the 1960s, the diffusive heat (Fig. 8b, long-dashed curve) from the upper layer becomes increasingly important and eventually becomes the dominant heat source, consistent with the finding of previous studies (e.g., Bryan et al. 1984; Cai and Gordon 1998), which indicates that under greenhouse conditions, heat fluxes from vertical diffusive and convective processes eventually determine the heating rate of the ocean interior.

These changes around the 1960s are apparently associated with the change in warming pattern discussed.
above. To gain further insight, we decompose the divergence of vertical diffusive flux into components via the bottom and top boundary surfaces of each layer, and the divergence of the advection term into vertical and horizontal advection components (see Fig. 7).

The divergence of vertical diffusive heat flux for the upper layer can only take place via the bottom surface of the layer and is already shown in Fig. 8a (long-dashed curve). Prior to the 1960s, the large subsurface warming leads to a decrease in vertical temperature gradient across the boundary surface between the two layers, leading to a decreasing diffusive heat loss to the lower layer. Since the 1960s, the surface warming in the east Pacific accelerates and the vertical temperature gradient across the boundary surface increases. As a result, diffusive heat loss from the upper layer to the lower layer intensifies. The divergence of vertical diffusion for the lower layer takes place via both the top (80 m) and bottom (270 m) surfaces, and is shown in Fig. 9. Via the bottom boundary surface, diffusive heat loss to the water below 270-m depth (dashed curve) increases because the vertical temperature gradient across the boundary surface is enhanced, leading to warming in the water below 270-m depth. However, the large increase since the 1960s in diffusive heat input from the upper layer more than offsets the diffusive heat loss to the water below 270 m. Consequently, the contribution of the net divergence of diffusive heat fluxes is to a warming trend since the 1960s.

An area integration of the divergence of horizontal advection term yields the heat flux across the north–south boundary surfaces, because the east–west boundary surfaces are land, which allows no exchange. The associated time series for the two layers are shown in Fig. 10a. Throughout the course of the experiments, in the lower layer heat is transported from the extratropics to the Tropics via the north–south boundary surfaces (dotted–dashed curve), whereas in the upper layer heat...
is transported out of the tropical Pacific across the north–south boundaries (dashed curve). These results are consistent with the modeled circulation pathways described above, and suggest that the heat advected out of the Tropics by oceanic currents may contribute to the large extratropical warming. There is, however, no change in the long-term trend of horizontal advection term in either layer, indicating that the change in the divergence of the total advective heat flux in Figs. 8a and 8b is caused by the change in the vertical advection component alone.

The divergence of vertical advective heat flux for each layer is shown in Fig. 10b. There is a sharp increase in heat transfer from the lower layer to the upper layer after the 1970s (dashed curve). Given that vertical advection is the product of the upwelling and temperature at the base of the bottom boundary surface, and because upwelling decreases since the 1960s, the increased heat gain by the upper layer means that at this base (80-m depth) warming more than offsets the weakened upwelling. In the lower layer (dotted–dashed curve), the divergence of the vertical advective heat fluxes begins...
to offset the warming by horizontal advective fluxes from the 1970 onward. This is, in part, due to the increased vertical advective heat transfer to the upper layer discussed above and in part due to a decreased vertical advective heat input from water below.

Thus, the heat balance in the tropical Pacific Ocean also undergoes a systematic change in association with the change in warming pattern. Prior to the 1960s, the source of warming in the lower tropical Pacific Ocean is heat advected by the oceanic pathways from the extratropics. This leads to a large subsurface warming, reducing the vertical temperature gradient between the upper and lower layers. Consequently, diffusive heat transfer from the upper to the lower layer is curtailed, and the surface heat input to the upper layer in the Tropics is transported to the extratropics by the surface oceanic circulation, contributing to the large warming there. After the 1960s, horizontal advective heat input from the extratropics to the tropical lower layer becomes increasingly offset by the vertical advective heat loss of the lower layer. In the meantime, the accelerated surface warming in the east Pacific leads to an increased vertical temperature gradient, diffusive heat transfer to the lower layer intensifies and becomes the dominant process for heating the lower layer.

5. Tropical–extratropical linkage and variability in the Pacific Ocean

Does the above mechanism for transmitting anomalies from the extratropics to the Tropics also operate on other timescales? We would expect that this to be the case regardless of how the extratropical anomalies are generated, whether by greenhouse forcing or by internal variability. As will be shown below, the process operates in an ENSO-like interdecadal variability in the 1000-yr control run. The characteristics of model interannual ENSO cycles in the control run are also described to show that the interdecadal variability is many respects similar to the interannual ENSO cycles.

Two time series of Niño-3 (5°S–5°N, 90°–150°W) SST index are formed, one from the annual mean SST anomaly field relative to the climatological mean and the other from monthly mean SST anomalies. The former is just the annual mean version of the latter. A spectral analysis of both time series reveals an interannual oscillation with a dominant periodicity of about 25 yr; and an interdecadal oscillation with a periodicity of about 4 yr. Figure 11 shows the spectrum for the annual mean Niño-3 index. In the following we focus on variability on these two timescales.

The spatial and temporal evolution associated with the interdecadal oscillation is explored by regressing gridpoint annual mean SST anomalies upon a filtered version of the annual mean Niño-3 index, filtered to retain variability on timescales of 15–30 yr. We refer to this filtered index as the interdecadal Niño-3 index. Regression coefficients for a half cycle at each grid point at different time lags are computed; shown in Fig. 12 (left column) are regression patterns with time lags at 3-yr intervals. The remainder of the life cycle can be inferred from these patterns by reversing the signs of the anomalies.

Similarly, lagged regression of gridpoint monthly mean SST anomalies upon a filtered version of the monthly mean Niño-3 index (filtered to retain variability on timescales of 2–6 yr, hereafter referred to as interannual Niño-3 index) is carried out. Maps of regression pattern for about half of the oscillation cycle at intervals of six months are shown in Fig. 12 (right column). For a given lag, a positive coefficient at any location implies that the SST anomalies at this location increase in phase with the Niño-3 index but the index lags by the indicated time.

Focusing on the model interannual ENSO cycles first (right column of Fig. 12), we see that at zero lag, the pattern corresponds to an El Niño warm phase. The life cycle is similar to that of the observed ENSO cycle (Rasmusson and Carpenter 1982; Zhang and Levitus 1997) and to that simulated by a low-resolution coupled models (e.g., Knutson et al. 1997; Timmermann et al. 1999b). Several features that deviate from those of the observed are the following: first, the model ENSO has the largest anomaly in the central equatorial Pacific rather than in the equatorial eastern Pacific; second, the anomaly that should be in equatorial central eastern Pacific extends too far to the western Pacific; and third, an examination reveals that the amplitude of the anomaly in equatorial region is about one-third of the ob-
served, a feature typical of the low-resolution coupled models (e.g., Timmermann et al. 1999b). It is not the purpose of this study to describe the detailed model ENSO mechanism. Our description here aims to demonstrate the similarity between the model interannual ENSO and interdecadal ENSO-like variability.

The pattern for the interdecadal variability at zero lag indeed resembles that of the model ENSO, but is broader in meridional scale, and is “horseshoe” shaped in structure with anomalies in the Tropics extending in the eastern Pacific to latitudes about 45° poleward. Such a pattern and the attendant impacts in our model have been reported in an earlier study (Walland et al. 1999) although the dynamics have not been explored. The structure is similar to that of the observed ENSO-like interdecadal variability (e.g., Zhang et al. 1997); in particular, the development and decay of a positive anomaly in the central and eastern equatorial Pacific is accompanied by the establishment of a negative anomaly in the midlatitudes and western Pacific. In the model, this extratropical negative anomaly centers at about 35°.

The tropical and extratropical linkage plays an important part in the interdecadal oscillation, as can be seen from Fig. 13. Figures 13a and 13b respectively show regression coefficients, for a half cycle, meridionally averaged over the longitude band 120°E–120°W and zonally averaged over the latitude band 5°S–5°N as a function of depth. Fig. 13c shows regression coefficients for surface wind stress plotted as vectors for this latitude/longitude area. The process of transmitting extratropical anomalies to the Tropics and the propagation of these anomalies from the equatorial western Pacific to the equatorial eastern Pacific bear strong similarity to that shown Figs. 6a,b. Thus the tropical and extratropical linkage is already in place in our model without external forcing. In the presence of greenhouse forcing, the linkage simply plays a similar role.

A similar tropical–extratropical linkage has been suggested to be responsible for the observed interdecadal Pacific oscillation by Zhang et al. (1998). They demonstrate the transmission of the large extratropical Pacific positive SST anomalies during the late 1960s to the Tropics via the ocean subsurface, and suggest that these warm anomalies have induced the severe El Niño events during the 1980s.

Figure 13 also reveals aspects of the self-sustaining process of the interdecadal oscillation. About 12 yr (Lag 12 yr) prior to the peak of positive SST anomaly in the central and east Pacific, a warm anomaly appears in the subsurface west Pacific (Fig. 13b), which in turn can be traced to the extratropics (Fig. 13a). At this phase, the tilt of the equatorial thermocline increases, and the pattern is La Niña–like, with easterly wind anomalies in the tropical west and central Pacific (Fig. 13c). In the following 9 yr, the subsurface extratropical anomaly moves further equatorward, and in the Tropics the anomaly moves eastward with the equatorial undercurrent to the central and east Pacific where it upwells. The thermocline tilt reduces, deepening in the east but shoaling in the west. The ocean–atmosphere feedback ensues.

Fig. 10. Time series of area integrated divergence of (a) horizontal and (b) vertical advective heat flux for the upper (dashed curve) and lower (dotted–dashed curve) tropical Pacific layer. These time series are obtained by averaging over those of the three experiments.
with weakening easterly winds and equatorial upwelling, leading to the positive anomaly in the central and eastern equatorial Pacific. In the meantime, about 6 yr (Lag 6 yr, Fig. 13b) prior to the peak of the positive SST anomaly in the Tropics, extratropical cooling develops. This negative anomaly is the seed for the decay of the tropical positive anomaly. At Lag 0 year, substantial negative anomalies have already reached the subsurface west Pacific, ready to commence the opposite phase of the oscillation.

A detailed account of the modeled decadal oscillation mechanism is beyond the scope of the present study, however, a brief examination on the processes that generate the extratropical negative anomalies is in order. Gu and Philander (1997) suggest that an extratropical negative anomaly can be a direct consequence of the tropical positive anomaly. The argument is that the response of the atmosphere to the tropical anomaly involves an intensification of the extratropical westerlies, which would increase the evaporative heat loss from the ocean to the atmosphere generating the negative SST anomaly.

To examine whether this process operates in our model simulations, Figure 14 shows the regression pattern at zero lag upon the interdecadal Niño-3 index of annual mean (a) evaporation, (b) surface wind stress vectors, (c) MSLP, and (d) upper oceanic current vectors superimposed on the regression pattern of SST. Several features emerge from comparisons of these patterns. First, correspondence between wind stress and evaporation anomalies, and between evaporation and SST anomalies, suggest that the process proposed by Gu and Philander (1997) is not captured by our model. Let us examine the North Pacific in the latitude band 30°–50°N; Fig. 14b shows that westerly winds reduce in this latitude band. However, evaporation weakens only in the western half of the latitude band but in the eastern half the evaporation rate increases. The reduced evaporation anomaly in the west would lead to positive SST anomalies, but the SST anomalies there are negative (Fig. 14d). In fact, at this latitude band, it seems that the weakened evaporation in the west and the enhancement in the east is a consequence of, rather than the cause for, the SST anomalies. Second, there are strong extratropical atmospheric responses to the tropical warming. Indeed, the extratropical SST anomaly can be the consequence of these atmospheric responses. For example, the MSLP response pattern in the northern mid- to high latitudes shows a Pacific–North America (PNA)-like structure with a deeper Aleutian low (Fig. 14c), and is largely in geostrophic balance with the wind stress pattern (Fig. 14b). The accompanying change in oceanic circulation is a weaker gyre circulation (Fig. 14d). This appears to have contributed to the horseshoe structure of the SST anomaly: in the east North Pacific Ocean, anomalous poleward advection generates positive SST anomalies that connect with the tropical anomalies of the same sign; in the west, it is the anomalously weak western boundary current that leads to the negative SST anomalies along the flow path.

This modeled mechanism supports the hypothesis that the PNA pattern acts as a link or “bridge” whereby tropical SST anomalies interact with the atmosphere, and the attendant atmospheric anomalies in turn act on the ocean to produce extratropical SST anomalies (Lau and Nath 1994; Graham et al. 1994; see Lau 1997 for a review). The process is also similar to that inferred from observations by Zhang et al. (1997), and is consistent with results from another climate model described by Pierce et al. (2000). Thus it seems that while the model interdecadal mechanism is in many respects similar to that proposed by Gu and Philander (1997), the extratropical anomaly in our model appears to be driven by the “atmospheric bridge” process.

We now return briefly to the model interannual ENSO cycles. While along the equator eastward propagation of SST anomalies is similar to that in the interdecadal oscillation, the subsurface anomalies in the equatorial western Pacific do not originate from the extratropics, but from anomalies in the subtropical eastern Pacific, which propagate westward and extend to the subsurface. The westward process is shown in Fig. 15a (Lag 6 months). As the negative anomalies propagate westward, they are reinforced. This reinforcement of westward-propagating anomalies is also produced by other models (e.g., Schopf and Suarez 1988), and is mainly driven by the wind stress perturbations. As in the Knutson et al. (1997) and the Timmermann et al. (1999b)
FIG. 12. (left column) Regression pattern ($^\circ$C $^\circ$C$^{-1}$) obtained by regressing annual mean SST anomalies at each grid point upon the interdecadal Niño-3 index (annual mean Niño-3 index filtered to retain variability on timescales of 15–30 yr), at different lags, showing the pattern for a time when the index is lagging by an indicated time in years. (right column) Regression pattern ($^\circ$C $^\circ$C$^{-1}$) obtained by regressing monthly mean SST anomalies at each grid point upon interannual Niño-3 index (monthly mean Niño-3 index filtered to retain variability on timescales of 2–6 yr at different lags), showing the pattern for a time when the index is lagging by an indicated time in months. Contour interval is 0.3, and positive values are shaded.

Before we leave this section we conclude that the major aspects of the process whereby the warming pattern prior to the 1960s changes to the El Niño–like warming pattern also operates on interannual and interdecadal timescales. The common features include (i) a subsurface anomaly first appears in the west Pacific, moves eastward to the central and east Pacific where it
upwells generating ocean–atmosphere positive feedbacks that reinforce the anomaly; and (ii) the subsurface anomalies originate from those outside the Tropics.

6. Summary

Recently Cai and Whetton (2000) provide modeling evidence that the greenhouse warming pattern since the late nineteenth century have undergone a systematic change from a pattern prior to the 1960s with maximum warming in mid- to high latitudes to one that is El Niño–like from the 1960s onward. They suggest that the underlying dynamical process for the systematic change is the transmission of the large extratropical warming to the equatorial east Pacific via modeled tropical–extratropical Pacific subsurface circulation pathways. The present study addressed several important issues previously unanalyzed.

We demonstrate that EOF analysis only separates the pattern of modes but may not separate the dynamics of these modes. In other words, modes other than EOF1 can be dynamically related to EOF1. In the context of greenhouse warming pattern, the time-varying pattern is represented by at least two EOFs. Thus, approaches

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Fig. 13. Regression pattern of gridpoint anomalies (°C °C⁻¹) upon the interdecadal Niño-3 index at different lags with the index lagging by an indicated time in years. (a) Pattern of temperature anomalies zonally averaged over the longitude band 120°E–120°W. (b) Pattern of temperature anomalies meridionally averaged over the latitude band 5°S–5°N. In both (a) and (b), contour interval is 0.2, and positive values are shaded. (c) Pattern of surface wind stress anomalies (dyn cm⁻² °C⁻¹) over the same lat–long area plotted as vectors.
assuming that only EOF1 is greenhouse related may not be appropriate.

We show that the tropical–extratropical Pacific circulation linkage whereby extratropical anomalies are transmitted to the subsurface equatorial east Pacific are already in place in the modeled circulation state without greenhouse forcing. The pathways are important ingredients of the modeled interdecadal ENSO-like oscillation. Under greenhouse forcing, the extratropical warming, which is initially larger than that in the Tropics, is transmitted to the Tropics in a similar way. In the east equatorial Pacific the warm anomaly upwells, and initiates the ocean–atmosphere feedback with decreasing equatorial west–east SST gradient and easterly winds, reinforcing the warm anomaly and leading to the development of the El Niño–like warming pattern.

Heat budget analysis further demonstrates the mechanism for the change of warming pattern. Throughout the course of warming experiments horizontal advection transfers heat from the extratropics to the lower-layer tropical ocean (80–270 m), and prior to the 1960s this advective heat flux is the primary heat source for warm-
ing the tropical lower-layer ocean, mostly in the subsurface west Pacific. By contrast, in the upper layer, heat is advected from the Tropics to the extratropics. Consequently, the vertical temperature gradient across the boundary surface between the upper- and lower-layer ocean decreases, reducing the diffusive heat transfer from the upper to the lower layer. From the 1960s onward, as the subthermocline warming upwells, the El Niño–like warming pattern develops, dramatically changing the relative importance of these oceanic processes. First, the horizontal advective heat input to the lower-layer ocean from the extratropics continues, but much goes to compensate for the heat loss arising as a result of a change in the divergence of vertical advection. Second, vertical diffusive heat transfer from the upper- to the lower-layer ocean increases, and eventually becomes the dominant heat source for warming the lower-layer tropical Pacific ocean.

Our result also indicates that the El Niño–like warming pattern will continue into the twenty-first century. This will have significant climatic impacts. For example, rainfall in the Australasia/western Pacific regions, where droughts occur during El Niño events and floods in La Niña years (Philander 1990), will start to decrease. This will contribute to a long-term depletion of water resources in the regions (Meehl and Washington 1996).
On the other hand, the eastern Pacific island countries will begin to experience a trend of increasing rainfall. In summary, our result greatly strengthens the argument that the El Niño–like warming pattern in recent decades is, in part, greenhouse induced, and may well persist.

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