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Extratropical control of recent tropical Pacific decadal climate variability: a relay teleconnection

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Abstract Observations indicate that recent tropical Pacific decadal climate variability tends to be associated with the extratropical North Pacific through a relay teleconnection of a fast coupled ocean-atmosphere bridge and a slow oceanic tunnel. A coupled ocean-atmosphere model, forced by the observed decadal wind in the extratropical North Pacific, explicitly demonstrates that extratropical decadal sea surface temperature (SST) anomalies may propagate to the tropics through a coupled wind-evaporative-SST (WES) feedback. The WES feedback cannot only lead to a nearly synchronous change of tropical SST, but also force a delayed adjustment of the meridional overturning circulation in the upper ocean to further sustain the tropical SST change. The study further suggests that the extratropical–tropical teleconnection provides a positive feedback to sustain the decadal changes in both the tropical and extratropical North Pacific.

1 Introduction

The tropical Pacific ocean-atmosphere system exerts powerful controls on global weather and climate patterns, as documented in numerous studies of El Nino/Southern Oscillation (ENSO) dynamics and impacts (e.g., Trenberth et al. 1998; Alexander et al. 2002; Wang and Picaut 2004). In addition to its interannual ENSO variability, the tropical Pacific climate also reveals substantial decadal fluctuations. The anomalous warm conditions over the equatorial Pacific in conjunction with the relaxation of the trade winds since mid-1970s have been noted in many studies, and frequently described in terms of the Pacific decadal climate regime shift (e.g., Mantua et al. 1997; Zhang et al. 1997). The persistent tropical climate decadal variations not only have profound impacts on marine ecosystems (Mantua et al. 1997; Chavez et al. 2003; Miller et al. 2003), CO_2 uptake (Flückiger et al. 1999; Freely et al. 1999) and the climate of North America (e.g., Barlow et al. 2001; Cayan et al. 2001), but also may modulate the behavior of ENSO, for example, a shift toward more frequent, intense, and long-lasting El Ninos since the mid-1970s (Fedorov and Philander 2000). Thus, the potential predictability of such large-scale decadal climatic swings would have great societal and economical benefits. However, the causes and mechanisms of the tropical Pacific decadal climate changes remain not fully understood (see a review by Miller and Schneider 2000).

Several hypotheses have been put forward for the origins of the tropical Pacific decadal variability. Among some prevailing theories, extratropical–tropical teleconnections are frequently assumed to act as a conveyor to transmit the extratropical climatic anomalies to the tropics, by means of fast atmospheric bridges (Barnett et al. 1999; Vimont et al. 2001) and/or slow meridional overturning circulation in the upper ocean (Gu and Philander 1997; Kleeman et al. 1999). However, supporting evidence from both observations

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and climate models for these hypotheses remains elusive. For instance, it is not clear that what mechanisms caused the recent decadal changes of the trade winds that spun down the upper ocean circulation (McPhaden and Zhang 2002), and also it remains not well understood what ocean-atmosphere coupled processes in the subtropics bridge the extratropics and tropics via the atmospheric teleconnection. Furthermore, it is usually difficult to identify causality based only on the analyses of observations or a fully coupled ocean-atmosphere simulation because the climate is already the final product of the complex feedbacks.

Here, using both observations and a coupled ocean-atmosphere model with a novel modeling strategy, we explicitly show that the recent tropical Pacific decadal climate change may be associated with the surface wind changes over the extratropical North Pacific that are then communicated to the tropics through a combination of both a fast coupled ocean-atmosphere surface bridge and a subsequently forced slow adjustment of the meridional overturning circulation in the upper ocean. The paper is constructed as follows. In following Section presents some observational evidence. Modeling results are presented in Modeling evidence Section. Conclusions and discussions are given in Conclusions and discussions section.

2 Observational evidence

The atmospheric data used here are taken from the National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR) reanalysis (Kalnay and Coauthors 1996). The sea surface temperature (SST) data are taken from Hadley Global Sea Ice and SST (HadISST, Rayner et al. 2003). All data are linearly detrended before analyses.

2.1 Extratropical–tropical teleconnection: surface coupled wind-evaporative-SST feedback

Observations tend to provide some evidence of extratropical–tropical climatic transmissions. Since the mid-1970s climate shift, the North Pacific Ocean has been generally dominated by anomalous cooling in the west, surrounded by warming anomalies in the eastern subtropical Pacific with extension along the North American coast to the Gulf of Alaska (Zhang et al. 1997). Most warm SST events in the eastern subtropical Pacific since the mid-1970s climate regime shift tend to exhibit a nearly consistent equatorward drift, especially from the 1980s to the late 1990s (Fig. 1). During this period, the equatorial warming events (El Nino) including the long-lasting warming in the earlier 1990s tend to be associated with the eastern subtropical warming. This subtropical-tropical SST teleconnection appears to be characterized by a rapid equatorward drift of the subtropical SST anomalies, with a timescale

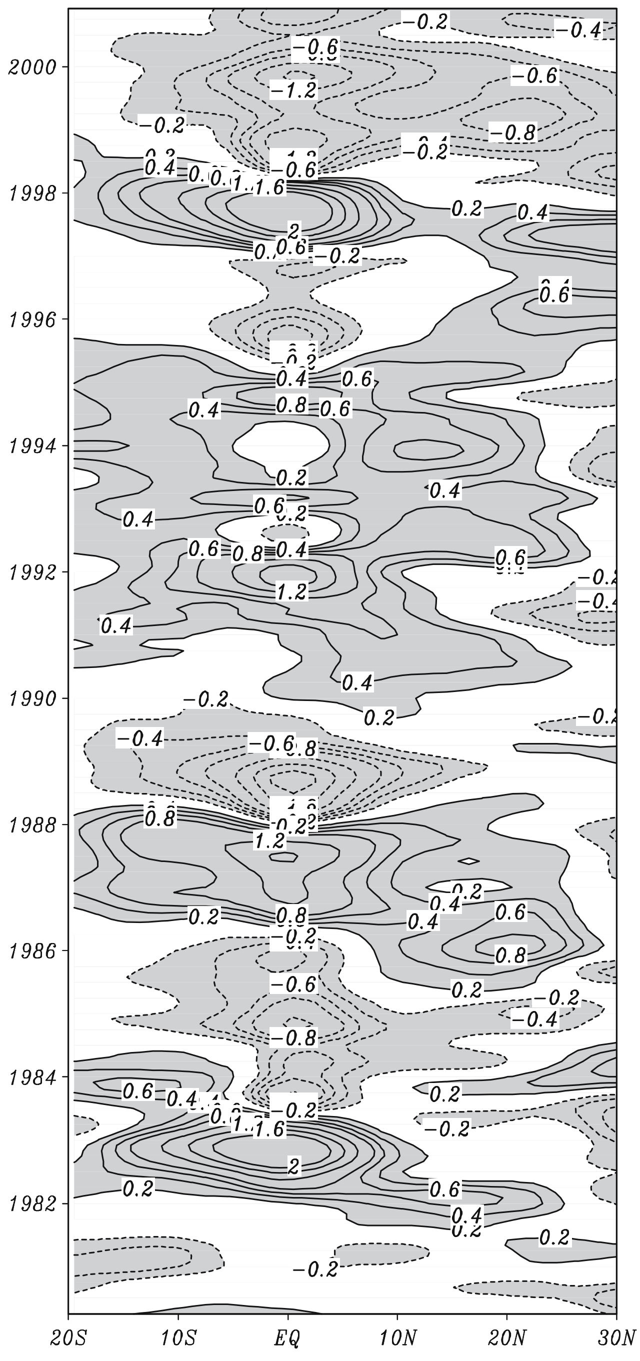


Fig. 1 History of the meridional distribution of monthly SST anomalies in the eastern subtropical–tropical Pacific. SST anomalies are smoothed using a 3-month running mean, and averaged over (160°W, 120°W) at each latitude

of only a few months from about 15–25°N to the equator.

The subtropical–tropical SST teleconnection is further demonstrated by regressing the Pacific basin SST and surface wind stress against the SST of the eastern subtropics. In winter, SST regression shows a pattern with anomalies predominant in the central and eastern subtropical North Pacific (Fig. 2a). This horseshoe-like pattern is sometimes referred to as the Eastern North Pacific Mode (Mestas-Nunez and Enfield 1999; Wu and Liu 2003), which shares great similarities to the pattern of the Pacific Decadal Oscillation (Mantua et al. 1997; Barlow et al. 2001). Over the North Pacific, associated with the SST anomalies is an anomalous cyclonic wind that intensifies the Aleutian Low and midlatitude westerlies, resulting in cooling in the western and central North Pacific and warming along the eastern boundary with a northward extension to the Gulf of Alaska. At lower latitudes, warming extends from the eastern subtropics to the west to 160°E and south to the tropics to 10°N. The warm SST anomalies in the subtropics are accompanied by anomalous southwest trades that may reduce the evaporative heat loss in this area. This lower subtropical branch is sometimes referred to as the Pacific meridional mode (Chiang and Vimont 2004). In the following spring and summer, the eastern subtropical warming extends further into the tropics and initiates warm SST anomalies in the central equatorial region between 130°W and 170°W (Fig. 2b).

What are the mechanisms leading to the subsequent warming in the tropics? It is noted that although the SST anomalies in winter stay mostly in the north of 10°N, anomalous southwesterly winds have developed in the equatorward of the warm SST anomalies in 140°W–170°W and 150°E–170°E (Fig. 2a). In this season, the mean ITCZ stays close to the equator, the anomalous southeasterly winds decelerate the trades to the north of the equator, reducing the evaporative heat loss to generate warming in the tropics in the following season (Fig. 2b). This coupled wind-evaporative-SST (WES) feedback between the atmospheric boundary layer and ocean (Xie and Philander 1994) has been invoked to explain the impacts of the extratropical annual variability on the equatorial annual cycle (Liu and Xie 1994). After reaching the equatorial region, the SST anomalies in the central equatorial Pacific set up a negative zonal pressure gradient to the west, which creates anomalous equatorial westerlies to further intensify the warming (Bjerknes's feedback). In fall and winter, the equatorial warming anomalies attain magnitudes comparable to the initial extratropical SST anomalies (Fig. 2c).

The equatorward propagation of subtropical SST anomalies due to the coupled WES feedback is further demonstrated in a lag-latitude plot of SST, surface wind and turbulent heat flux zonally averaged over the western and central Pacific from 160°E to 140°W (Fig. 3). In winter, the warm SST anomalies in the subtropics (north of 15°N) are associated with a weakening of the northeast trades. While the SST anomalies at this time mostly stay in the north of 10°N, the anomalous southwesterly winds and positive (downward) heat flux have extended to the northern deep tropics. This is because the warm SST anomalies in the north of the equator set up a negative meridional pressure gradient, inducing a C-form southerly cross-equator wind pattern in the tropics, a feature frequently seen in the tropical Atlantic basin (see a review by Xie and Carton 2004). The anomalous southwesterly winds north of the equator decelerate the northeast trade winds and reduce evaporative heat loss, leading to a development of warm SST anomalies in late spring and summer in the northern deep tropics. In this region, the SST anomalies peak in July and decay rapidly afterward together with the winds. In the tropics, the warm SST anomalies are initiated around May associated with a weakening of the equatorial easterlies and a reduction of turbulent heat loss, peak in October and decay afterward.

In summary, the above observational analyses demonstrate an apparent equatorward progression of the subtropical SST anomalies through the coupled WES feedback. Given decadal SST anomalies in the subtropics, it is conceivable that this fast atmospheric bridge can continuously bring subtropical SST anomalies to the tropics in each year to sustain decadal SST anomalies in the tropical region.

2.2 Extratropical–tropical teleconnection: subtropical–tropical cell

It is also conceivable that the persistent wind stress generated by the subtropical decadal SST anomalies through the WES feedback may further impact the ocean circulation. Observations show evidence of an extratropical–tropical subsurface oceanic teleconnection through the meridional overturning circulation in the upper ocean, referred to as the subtropical–tropical cell (STC) (Liu et al. 1994; McCreary and Lu 1994; Schott et al. 2004) that is: water masses in the subtropics, especially in the eastern subtropical Pacific, subduct into the pycnocline (where the water density increases rapidly with depth), and then move to the tropics in the subsurface within about a decade. Observations have indicated a weakening trend of the transport of the STC since mid-1970s climate shift (McPhaden and Zhang 2002). The weakening of the

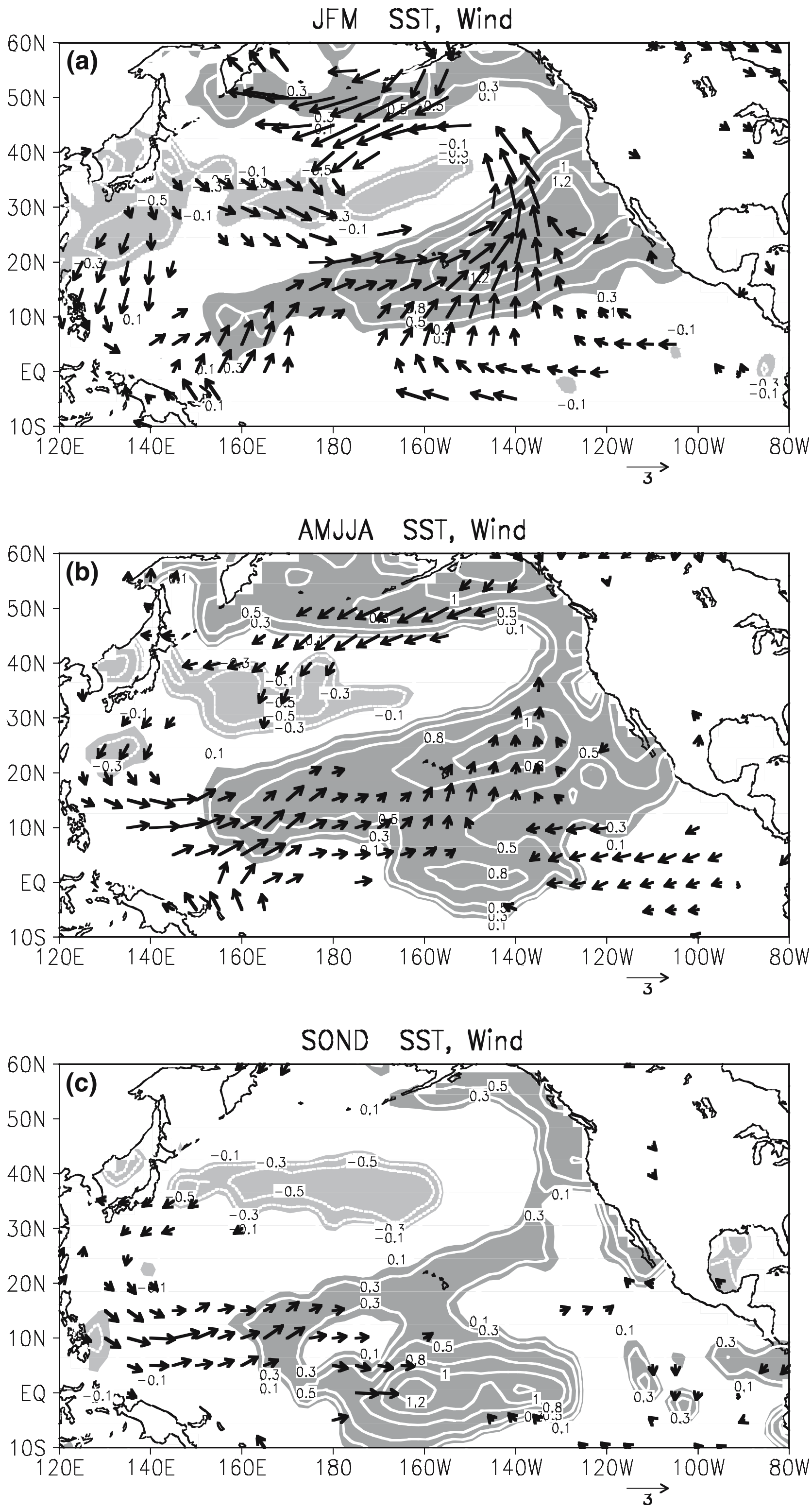


Fig. 2 Regression of SST (*shaded* and *contoured*) and surface wind (*vectors*) against the winter (DJF) eastern subtropical SST index. Shown are lagged regressions at **a** JFM, **b** AMJJA, and **c** SOND. The subtropical SST index is taken as an average of SST anomalies over the domain (150°W, 120°W) × (20°N, 30°N). Only regressions exceeding statistical significance are plotted (statistical significance levels for SST and wind are 90 and 85% using a two-tailed Student-*t* test, respectively). The wind data are from National Center For Environmental Prediction Reanalysis. Units for SST and surface wind are °C and m/s per degree (the SST index), respectively

STC can also be inferred from the vertical–meridional temperature changes (Fig. 4). In early 1980s, warm anomalies dominate both the surface and the subsurface of the eastern subtropical north Pacific within 5°N–20°N, while in the tropics and south Pacific cooling dominates the subsurface (Fig. 4a). The warm anomalies in the eastern subtropical north Pacific in the upper 100-m appear to penetrate to lower latitudes

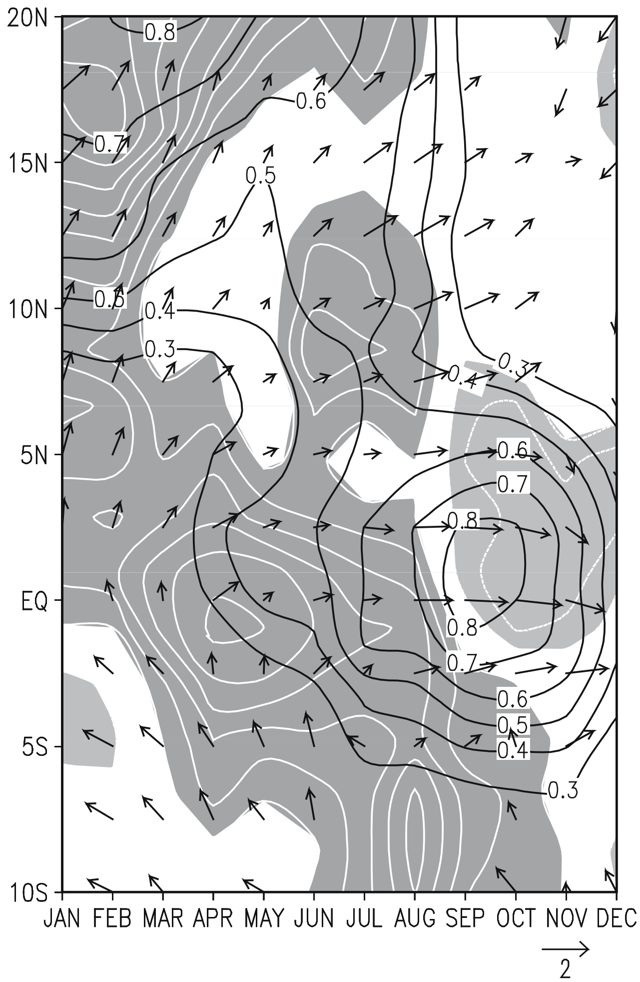


Fig. 3 Time–latitude plot of SST (*black contour* at 0.1°C interval, the minimum is 0.3°C), surface turbulent heat flux (latent + sensible, white contours at 2 W/m^2 interval, *shaded* >0.5 W/m^2), and surface wind (m/s) constructed from Fig. 2. All variables are zonally averaged over (160°E, 130°W)

(Fig. 4b) in the middle 1980s, and reach the equatorial region in early 1990s (Fig. 4c). Upon reaching the equatorial region, the warm anomalies are upwelled to the surface and amplified presumably by local ocean-atmosphere feedback. It is also noted there is an equatorward penetration of the cold anomaly within 20°N–30°N (Fig. 4a). This cold anomaly is largely generated in the central North Pacific and blocked in the north of 10°N (e.g., Schneider et al. 1999).

Some earlier studies ascribed this equatorward penetration of the subtropical subsurface temperature anomalies to the subduction process where the mean ocean current acts as a conveyor of these anomalies (Gu and Philander 1997; Zhang et al. 1998). However, recent observations and modeling studies indicate the subsurface anomalies in the lower latitudes are largely controlled by the local wind stress (Schneider et al. 1999; Nonaka et al. 2002; Schott et al. 2004), and therefore an indicator of the weakening of the mass transport of the STC: a weaker STC pumps less cold extratropical thermocline water into the tropics, leading to warming in the tropics. Observations indicate that the weakening trend of the STC since the mid-1970s climate shift is largely attributable to the weakening of the northeast trades (McPhaden and Zhang 2002). Over the subtropics, the weakening of the northeast trades persists from mid-1970s to early 1990s, except in the western subtropical Pacific where the trade winds started to relax somewhat in the late 1980s (Fig. 5). Nevertheless, the general trend of the trades relaxation is consistent with the slowdown of the STC during this period. However, the origins of the relaxation of the northeast trades remain not well understood. Here, our observational analysis suggests that the surface coupled WES feedback generates anomalous westerlies in the lower latitudes, which may potentially slow down the STC, and further enhance the tropical warming.

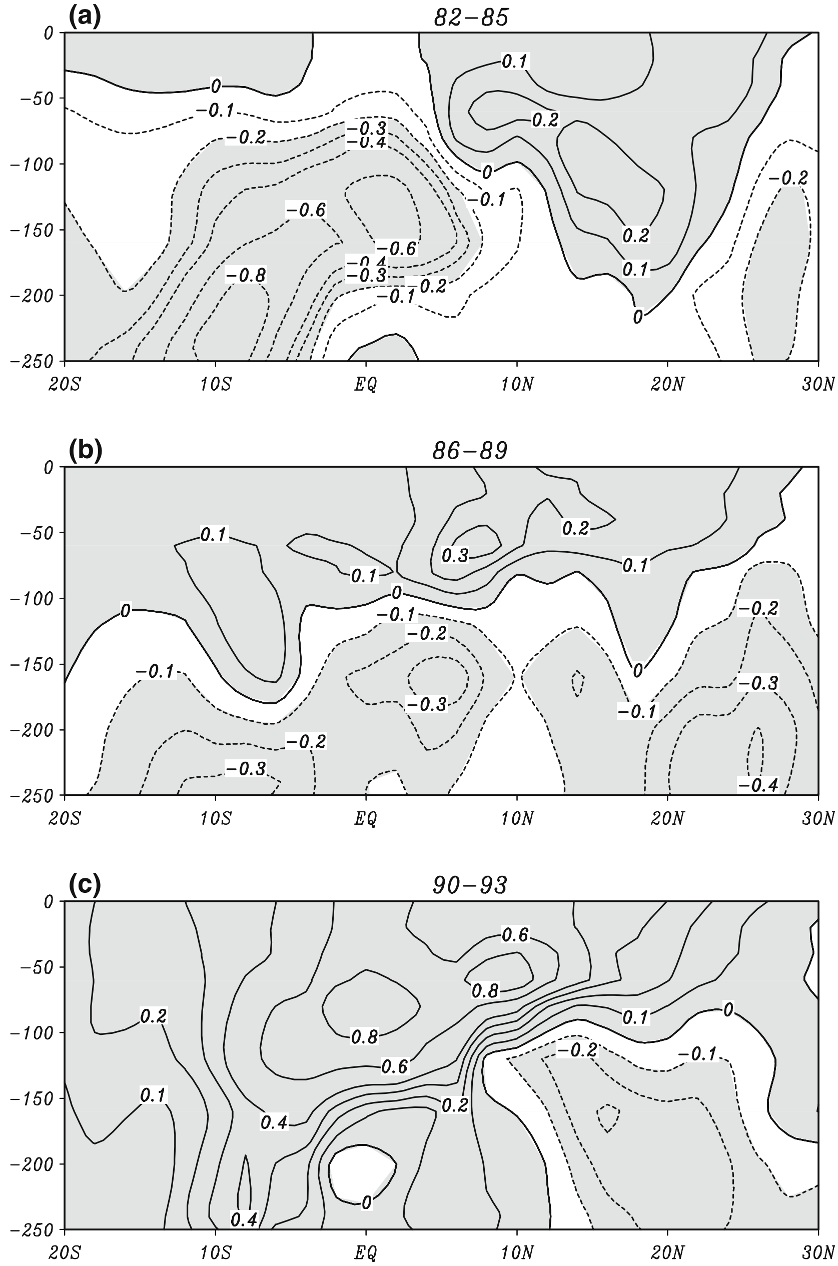
The observational analyses above lead us to hypothesize that the recent tropical Pacific decadal change may be associated with the extratropical North Pacific, which can drive the tropics through the fast surface coupled process and the subsequent slow change of the STC. We refer to this combined extratropical–tropical teleconnection of the fast surface coupled bridge and the slow oceanic tunnel as a relay teleconnection. In the following, we will further test this ‘‘Relay Teleconnection’’ hypothesis by using a coupled ocean-atmosphere model.

3 Modeling evidence

3.1 Model and experiments

The model we used is the Fast Ocean-Atmosphere Model (FOAM, Jacob 1997) Version 1.5. The atmospheric model is a parallel version of NCAR Community Climate Model (CCM) 2.0 at R15, but with atmospheric physics replaced by CCM3 Version, and vertical resolution of 19 levels. The ocean model is similar to the Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model (MOM) with a resolution of 1.4° -latitude × 2.8° -longitude × 24 levels. Without flux adjustment, the model has been run for over a 1,000 years without any apparent climate drift. FOAM produces reasonable ENSO (Liu et al. 2000) and Pacific decadal climate variability (Wu et al. 2003).

Fig. 4 Meridional–vertical profiles of temperature anomalies (°C) averaged over the eastern Pacific (170°W, 120°W) for three time periods: **a**1982–1985, **b** 1986–1989, **c** 1990–1993. The temperature anomalies are smoothed by applying a 4-year running mean. The subsurface ocean temperature profiles used are from XBT observations



To test above hypothesis, we performed ensemble experiments forced by observed decadal wind. In these ensemble experiments, the observed anomalous monthly wind forcing (including both wind speed and wind stress) is applied to drive the ocean component of the coupled system over the North Pacific (north of 20°N), while the ocean and atmosphere are fully coupled elsewhere. Therefore, the anomalous climatic response of the model outside of the North Pacific is predominantly caused by the imposed North Pacific wind. This new coupled modeling approach is different from traditional ocean-GCM studies because the tropical Pacific ocean-atmosphere are fully coupled. As such, we can explicitly assess the impacts of the North Pacific decadal climate change on the tropics through both atmospheric and oceanic teleconnections in the presence of full-coupled ocean-atmosphere interaction (south of 20°N). It should be noted that the experiment here cannot address the origins of the North Pacific decadal climate changes, because the wind specified here may include impacts from remote tropical SST changes.

The observed anomalous monthly North Pacific wind forcing is obtained by subtracting the monthly climatology of the period 1948–1976 from that of the period 1977–1996. In the experiments, this anomalous

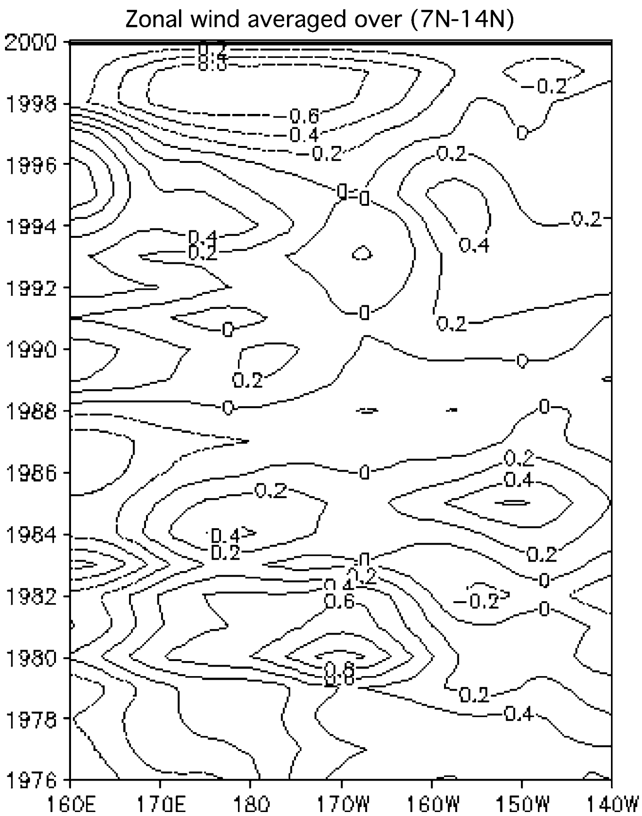


Fig. 5 Time-longitude plot of zonal wind anomalies averaged over the latitudinal band (7°N–14°N). Unit is m/s, and data are smoothed by applying a 3-year running mean

wind is superimposed onto the model mean climatological wind to drive the ocean component of the coupled system in the North Pacific. To eliminate the error caused by the model climatology drift due to the constraint of the wind stress, we first conducted a 150-year control simulation in which a zero wind anomaly is superimposed over the model mean climatological wind. Then we conducted a 30-member ensemble experiment, with each member starting from a different initial state of the control simulation (with zero wind anomaly), and being integrated for 10 years with the wind forcing repeated every year. The differences between the ensemble average of these experiments and the control simulation are taken as the response.

3.2 Extratropical–tropical teleconnection: surface coupled WES feedback

The experiment demonstrates that wind variations in the North Pacific generate a decadal-like SST pattern in the North Pacific and exert significant control over the tropics through both the surface coupled ocean-atmosphere process and the subsequent slow adjustment of the STC. In winter, the imposed cyclonic wind generates cooling in the central and eastern North Pacific and warming in the subtropics (Fig. 6a). The cooling is primarily associated with anomalous southward Ekman flow (not shown), which brings cold subpolar water to the midlatitudes. The warming in the subtropics is primarily caused by a reduction of the evaporative heat loss associated with the weakening of northeast trades. Although the warming is largely limited to be to the north of 20°N, anomalous southwesterly winds have been generated in the southern flank of the subtropical SST anomalies, particularly around 140°W and 160°E. As a result, in the following spring and summer, the warming in the subtropics quickly spreads toward the lower latitudes, and initiates warming in the eastern tropic Pacific (Fig. 6b). The equatorward spreading of the warm SST is coupled with anomalous southwesterlies, leading to a reduction of the evaporative heat loss. In the fall, the equatorward drift of SST appears to be less significant, and the SST is dominated by strong development in the equatorial region coupled with anomalous westerlies, with a magnitude comparable to the SST anomalies in the eastern subtropics (Fig. 6c). Overall, the equatorward propagation of the subtropical SST coupled with the surface wind simulated in this idealized experiment shares great similarities with the observational analyses (Fig. 6 versus Fig. 2), although there are some apparent discrepancies. For instance, in the fall, some cold anomalies develop in the latitude band 10–20°N, which do not appear in the observations (Fig. 6c versus Fig. 2c). This cooling appears to be associated with a reduction of the radiative heating in the model due to an increase of the cloudiness. Also the warming in the extratropics extends too far west and east, which may be associated with the model bias of a could tongue that extends too far westward, as also seen in most coupled GCMs (e.g., Davey et al. 2002).

To further demonstrate the role of the coupled WES feedback in the equatorward propagation of the subtropical SST, a time–latitude Hovmuller diagram of SST, surface wind and turbulent heat flux (similar to Fig. 3) is plotted (Fig. 7). In winter, although the warming stays largely in the north of 20°N associated with a reduction of latent heat loss due to the weakening of the northeast trades, anomalous southwesterly winds have extended down to the latitude band of 10–20°N and decelerate the northeast trades. This leads to a reduction of local evaporative heat loss, initiating warming in this latitude band in the following spring season. As the warming spreads equatorward to 10°N in late spring, some anomalous southwesterly winds

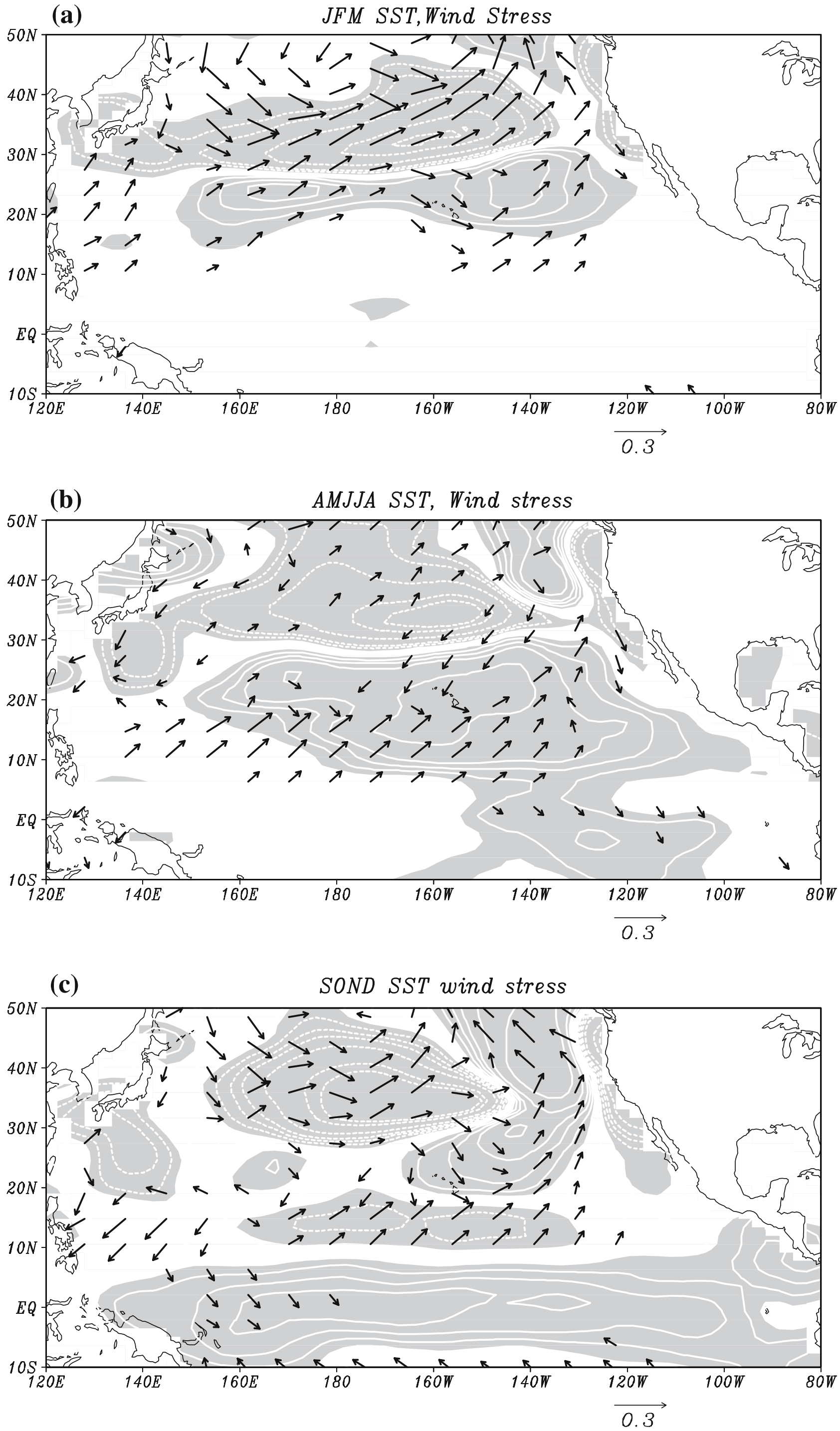


Fig. 6 Sea surface temperature (*white contour* at 0.1°C interval, *shaded* >0.05°C) and wind stress (*arrows*, N/m^2) for the first year of the ensemble experiments forced by the observed North Pacific decadal wind anomalies. **a** JFM (January–March), **b** AMJJA (April–August), **c** SOND (September–December). Plotted SST and wind stress exceed 90, and 85% statistical significance, respectively. The wind stress is square-rooted for plotting purpose

have been generated in its equatorward flank, leading to warming further into the tropics. In the tropics, substantial warming appears from summer to fall coupled with anomalous westerlies. The equatorial warming is largely associated with a reduced local upwelling, rather than the surface heat flux through the WES mechanism.

Overall, the model demonstrates a coupled equatorward propagation of SST, wind and heat flux from the extratropics to the tropics as seen in the observations. However, there are some differences, particularly the timing of the warming in the tropics. In the observations, the tropical warming starts in spring, but in the model experiment it starts in summer. This is

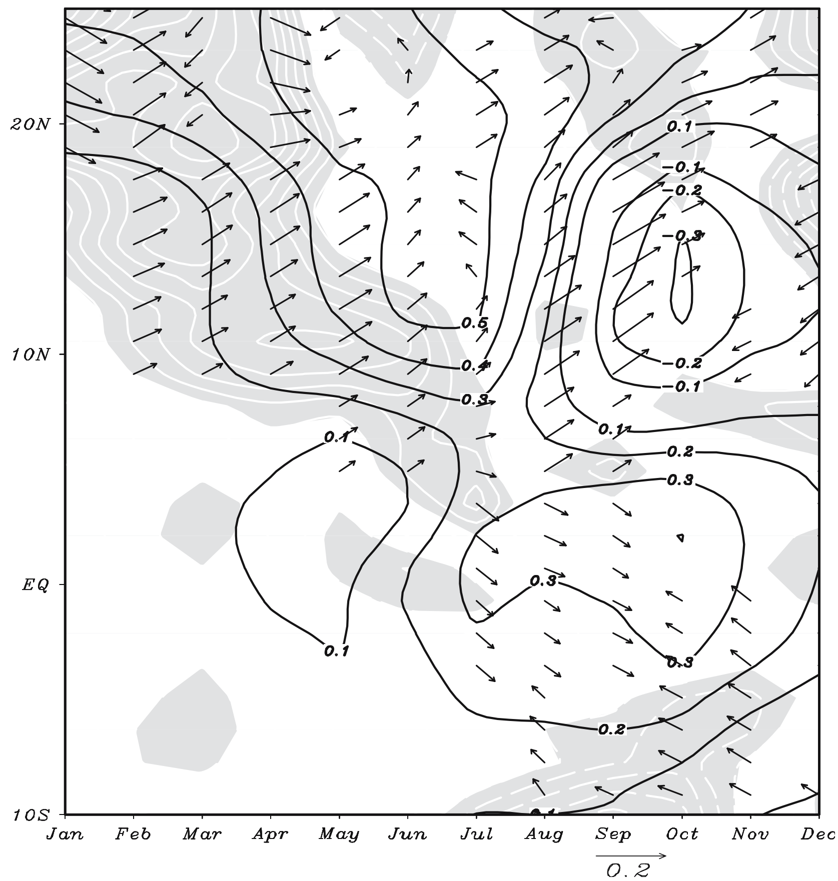


Fig. 7 Time–latitude plot of SST (black contour with interval at 0.1°C), surface turbulent heat flux (latent + sensible, *white contour* at 1 W/m^2 interval, *shaded* >2 W/m^2), and wind stress (*arrows*, N/m^2) constructed from Fig. 6. All variables are averaged zonally over (180°, 120°W)

possibly due to the design of the experiment. In the model, in order to see more clearly how the subtropical SST propagates to the tropics, the forcing is restricted to be the north of 20°N to leave more space to identify the WES feedback. In reality, the atmospheric forcing associated with the change of the Aleutian Low can extend to 10°N (Vimont et al. 2001), as seen in the observations (Fig. 2a), which could therefore produce warming earlier than the model experiment. A similar experiment with forcing set to the north of 10°N was also performed, which shows an equatorward propagation more similar to the observations (not shown).

In summary, this experiment appears to support the surface coupled ocean-atmosphere bridge (WES feedback) as an effective conveyor of SST anomalies from the extratropics to the tropics. This fast coupled process repeats yearly in concert with the decadal North Pacific wind forcing to sustain the decadal SST anomalies in the tropics. This annually repeating process is further demonstrated in Fig. 8a, which plots the time series of the tropical SST and the wind stress north of the equator over the model integration period. Both of them exhibit a repeating annual cycle, with annual averages dominated by anomalous westerlies and warming, respectively. The tropical SST appears to lag the off-equator wind by about a season as shown in the lagged correlation between the equatorial SST and the off-equatorial wind (Fig. 8b), indicating the role of off-equatorial westerlies in forcing the equatorial warm SST anomalies. The repeating westerly pulse in each year by the coupled WES feedback cannot only continuously feed the warming in the tropics, but also may further change the STC circulation. Also note that after year 6, the tropical SST appears to shift to a persistent warming regime. This sustained warming is consistent with a weakening of the STC by the persistent relaxation of the northeast trades associated with the WES feedback.

It is also noted that the equatorial warming may produce off-equatorial anomalous easterlies that can in turn reduce the anomalous westerlies induced by the WES feedback. This is shown in Fig. 8b, where a negative peak occurs when the equatorial SST leads the off-equatorial wind by a month. This anomalous easterlies may offset warming in the off-equatorial region, especially in the season followed the equatorial warming.

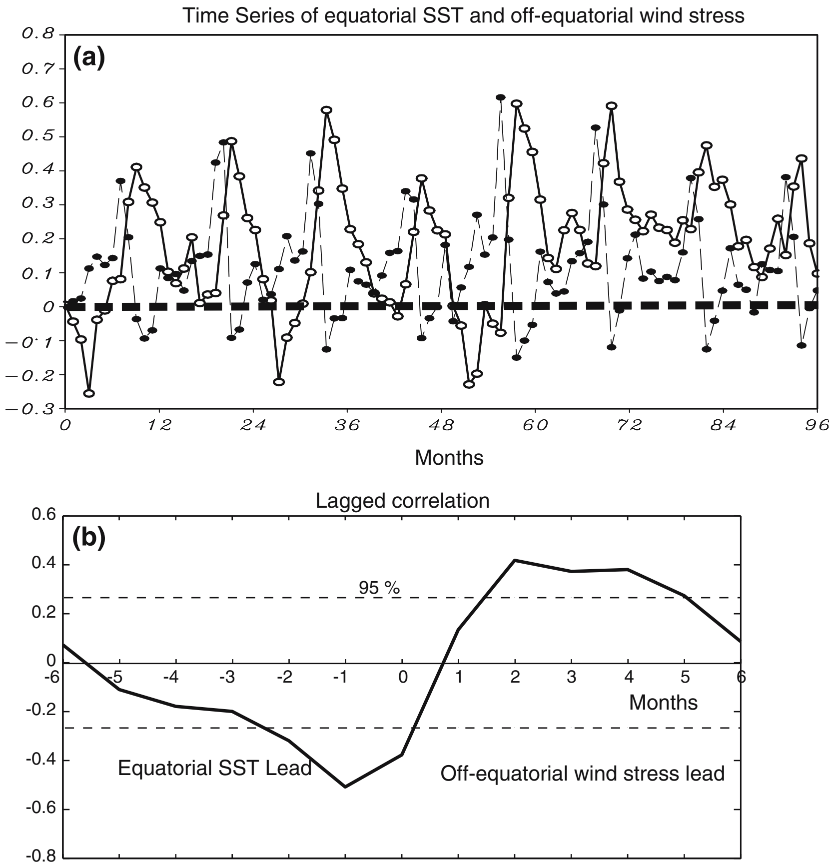


Fig. 8 **a** Time series of equatorial SST (*solid curve* with *open circles*) and off-equator wind stress (*dashed curve* with *bold circles*) over the model integration period. SST (°C) and wind stress (N/m^2) are averaged over (160°E, 130°W) × (5°S, 5°N) and (160°E, 130°W) × (7°N, 14°N), respectively. Wind stress is multiplied by a factor of 20 for plotting purpose. **b** Lagged correlation between the equatorial SST and the off-equatorial wind stress. *Dashed lines* represent the 95% statistical confidence level

3.3 Extratropical–tropical oceanic teleconnection: subtropical–tropical cell

The model further demonstrates that the anomalous westerly wind in the subtropics generated by the WES feedback leads to a weakening of the STC to further sustain the tropical warming. In the first two years, the warming in the tropics is generated predominantly by the fast surface coupled process, which enables the tropics to nearly synchronize with the extratropics (Fig. 9a). The SST in the tropics has the same sign as that in the subtropics, with the maximum warming trapped near the surface. During this period, there is no subsurface connection between the tropics and subtropics. In the following years, the tropical warming is enhanced as the subtropical–tropical subsurface teleconnection becomes apparent and the maximum warming occurs in the subsurface (Fig. 9b, c). As in the observations (Fig. 4), the subtropical warm anomalies in the upper 100-m appear to slowly penetrate to the tropics to intensify the warming in the equatorial thermocline (Fig. 9b), and subsequently are upwelled to further intensify surface warming (Fig. 9c). This oceanic subsurface teleconnection predominantly reflects a weakening of the STC in response to the surface anomalous westerlies generated by the coupled WES feedback (Fig. 10). In the first few years, the STCs have already shown a reduction of mass transport in both the north and south Pacific, but mostly in the tropical cell (TC) branch (Fig. 10a). The reduction of the mass transport is slightly stronger in the southern STC than its northern counterpart, with 2.5 and 3 Sv in the north and the south, respectively. The fast reduction of the TCs is driven predominantly by the local equatorial wind anomalies associated with the warming generated by the WES feedback. In the following years, the mass transport of the STCs in both hemispheres further declines, with a maximum reduction of 4 Sv (Fig. 10b). In the northern hemisphere, the anomalous STC shows a more broad meridional extension than the earlier stage, suggesting a slow adjustment of the subtropical cell branch. In the southern hemisphere, the STC adjustment is still dominated by the TC branch, indicting the contribution from the southern subtropical cell branch to the equatorial warming is less significant. This is because the forcing is imposed in the extratropical North Pacific.

It should be noted that although the model demonstrates the effects of equatorward propagation of

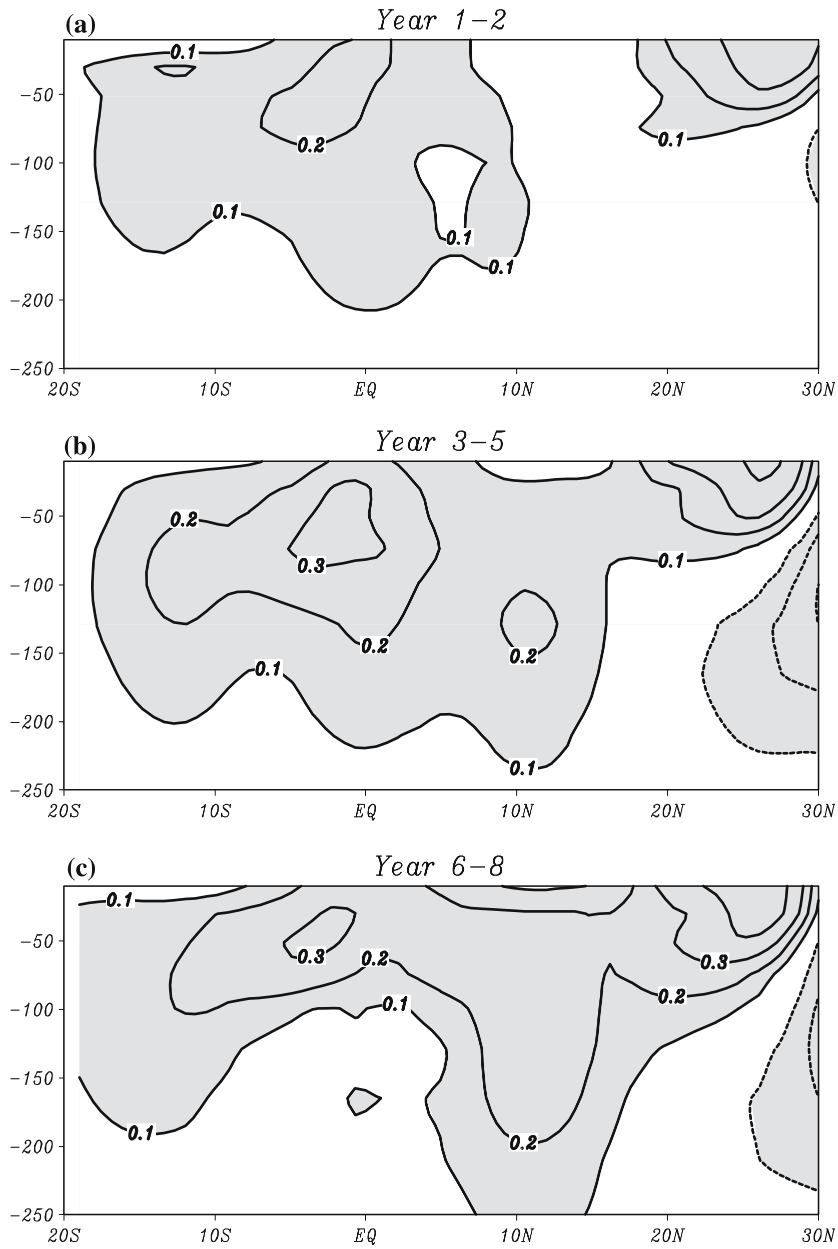


Fig. 9 Meridional–vertical profiles of temperature anomalies averaged over the eastern Pacific (170°W, 120°W) in the ensemble experiments forced by the observed North Pacific decadal wind anomalies. Results are shown for three time periods (**a**–**c**) to demonstrate the development

subtropical thermocline anomaly on the equatorial thermocline that is consistent with the observations, a direct comparison of the structure of the equatorial thermocline anomalies between the model and observations is not applicable. For example, the subsurface cooling seen in the observations (Fig. 4) was absent in the model simulations. This is because the modeling experiments here are targeted to study mechanism, rather than to make a hindcast. The difference between the model and the observations can be attributed to several factors including the initial state, influence from southern hemisphere, local wind stress, ENSO variability, etc. Nevertheless, the similarity of the tropical–extratropical oceanic subsurface connectivity between the model simulation and the observations suggests that the mechanism is the same.

4 Conclusions and discussions

Both the observational and modeling studies here tend to suggest that extratropical North Pacific decadal climate change can drive tropical climate change through a relay teleconnection carried by the rapid surface coupled ocean-atmosphere processes and the subsequently forced slow adjustment of the STC (Fig. 11). This relay teleconnection can be described as: a persistent strengthening (weakening) of the Aleutian Low

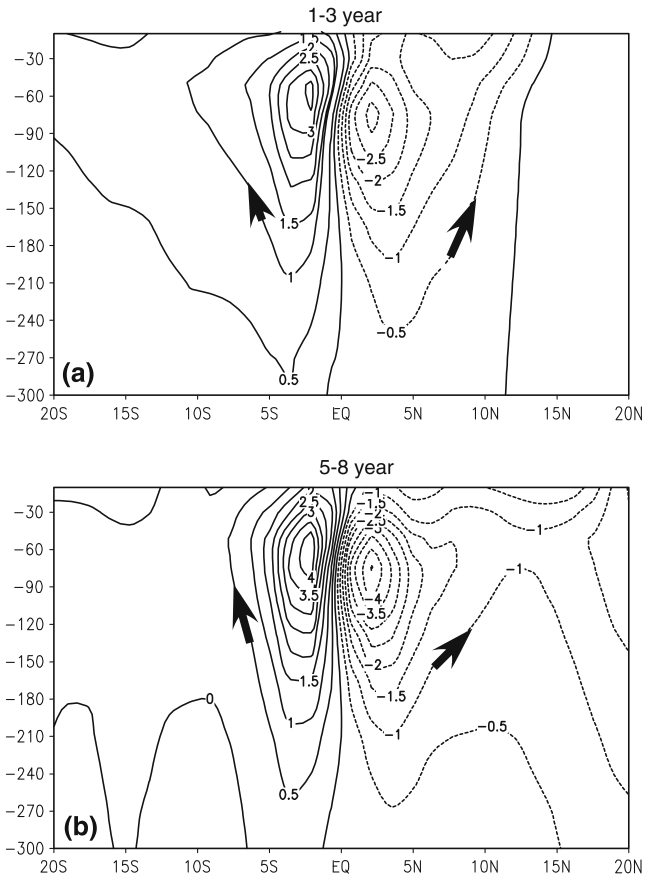


Fig. 10 Change of meridional overturning circulation in the subtropical–tropical Pacific calculated from the model. The contour is the streamfunction, and unit is Sv (10^6 m^3/s): **a** 1–3 years, **b** 5–8 years. *Arrows* denote the direction of the flow

creates cooling (warming) in the western-central North Pacific and warming (cooling) in the eastern subtropical Pacific; the latter rapidly propagates to the tropics through the WES feedback to create warming (cooling) in the equatorial region nearly synchronously; this fast surface coupled process operates annually and creates persistent westerly (easterly) anomalies in the subtropics, leading to a weakening (strengthening) of the STC to further sustain the warming (cooling) in the tropics.

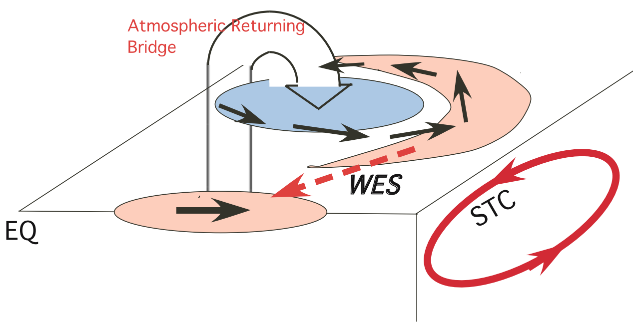


Fig. 11 Schematic diagram of the extratropical–tropical relay teleconnection. A persistent strengthening (weakening) of the Aleutian Low creates cooling (warming) in the western-central North Pacific and warming (cooling) in the eastern subtropical Pacific; the latter rapidly propagates to the tropics through the WES feedback to create warming (cooling) in the equatorial region nearly synchronously; this fast surface coupled process operates annually and creates persistent westerly (easterly) anomalies in the subtropics, leading to a weakening (strengthening) of the STC to further sustain the warming (cooling) in the tropics. As a result, the warming (cooling) in the tropics can further intensify (reduce) the Aleutian Low through return atmospheric bridge, forming a positive feedback loop

The proposed relay teleconnection here appears to unify different, prevailing hypotheses of the extratropical–tropical teleconnection. The WES feedback provides an alternative explanation of the extratropical influence on the tropical SST via the atmospheric bridges (Barnett et al. 1999; Vimont et al. 2001). The equatorward progression of the subtropical SST anomalies documented in this study is consistent with earlier studies by Vimont et al. (2003a, b), in spite of using different statistical methods and models. In our study, the equatorward progression in the northern deep tropics is attributed to the latent heat flux rather than the short-wave heat flux as in their study, which is perhaps associated with different parameterization scheme of tropical convection in the models (Vimont, personal communication). The fast atmospheric bridge also provides a potential forcing mechanism for the slowdown of the STC transport since 1970s (McPhaden and Zhang 2002).

The study here highlights the role of coupled ocean-atmosphere interaction in the eastern subtropical Pacific that acts as a conveyor to transmit the extratropical decadal climate variability to the tropics. The study, however, does not address what mechanisms drive the North Pacific decadal climate change, because the wind forcing is prescribed in the North Pacific. Coupled modeling studies indicate that the North Pacific ocean-atmosphere interaction itself can generate decadal variability (Barnett et al. 1999; Wu and Liu 2003). Our recent study also demonstrates that the mid-1970s climate shift may be attributed to the coupled ocean-atmosphere interaction within the North Pacific (Wu et al. 2005). Nevertheless, the study here, although highlighting the role of the North Pacific decadal climate variability in causing changes in the tropics, does not exclude the substantial influence exerted by the tropics on the North Pacific (e.g., Alexander et al. 2002; Deser et al. 2004). It is conceivable that the tropical decadal change induced by the North Pacific can further intensify the North Pacific SST anomalies through the returning atmospheric bridge. Therefore, the extratropical–tropical interaction including both the atmospheric bridge and the oceanic tunnel provides a positive feedback, rather than a negative feedback (e.g., Gu and Philander 1997), to sustain decadal climate changes in both the tropical and the extratropical North Pacific.

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