Wind-driven shifts in the latitude of the Kuroshio-Oyashio Extension and generation of SST anomalies on decadal timescales[1][2]

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Abstract

The causes of decadal variations of North Pacific sea surface temperatures (SSTs) are examined using a hindcast performed with an ocean general circulation model thermodynamically coupled to an atmospheric mixed layer model (OGCM-AML model) and forced by the time history of observed surface winds. We focus on the 'shift' in North Pacific ocean climate that occured around 1976/77 since this is the best observed example available. After the 1976/77 shift the Aleutian Low deepened and moved to the southeast of its previous position. This placed anomalous cyclonic flow over the North Pacific. The SST response, as simulated by the ocean model, consisted of two components: a fast and local part and a delayed and remote part. In the central Pacific stronger westerlies cool the ocean by increased equatorward Ekman drift. Here the dynamical cooling is sufficiently large that the surface fluxes damp the SST anomaly. This Ekman response is fast and local and cools the SSTs beginning in 1977 and persisted through 1988. In the early 1980s cool SSTs emerge in the latitude of the Kuroshio-Oyashio extension east of Japan and persisted until 1989. It is shown that this region of cooling is associated with a southward displacement of the latitude of the confluence between the subpolar and subtropical gyres. This is consistent with the southward shift in the zero wind stress curl line. The timescale for the gyre adjustment is no more than four years. These results compare favorably with observations which also first show the central Pacific cooling and, later, cooling east of Japan. Observations show the cooling in the Kursohio-Oyashio extension region to be damped by surface fluxes, implying an oceanic origin. The timescale of adjustment is also supported by analyses of observations.

The delayed response of the ocean to the varying winds therefore creates SST anomalies as the latitude of the gyre confluence varies. The delayed SST response is of the same sign as the locally forced SST signal suggesting that, to the extent there is a feedback, it is positive. Implications for the origins of decadal climate variability of the North Pacific are discussed.

1. Introduction

The search for the causes of decadal climate variability continues. Climate variability on decadal timescales has been demonstrated in the North Atlantic Oscillation (NAO) and the North Atlantic Ocean (e.g. Hurrell and Van Loon 1997), the tropical Atlantic (e.g. Rajagopolan et al. 1998) and the mid-latitude and tropical Pacific Ocean (e.g. Tourre et al. 1999). It is unclear whether this variability is simply the red noise that might be expected on the basis of the ocean's ability to integrate the atmosphere's white noise forcing (Hasselmann 1976, Frankignoul and Hasselmann 1977) or whether it exhibits distinct spectral peaks at preferred timescales. If the latter is true then it is typically concluded that an active role for the ocean must be involved. This is because changes in ocean heat transport could, it is supposed, introduce preferred timescales. However, proving that the ocean plays an active role in decadal climate variability has been difficult for two reasons. First, models and data agree that mid-latitude SST variability is strongly influenced by surface flux variability forced by the changing atmospheric circulation (e.g. Cayan 1992a, 1992b) while it is less clear how changes in ocean heat transport impact the SST. Secondly, even if the ocean circulation does impact SST variability, it is not clear that the mid-latitude atmospheric circulation subsequently responds (see Kushnir and Held 1996 for a review).

In this paper we will attempt to understand how decadal timescale changes in North Pacific SSTs are generated. A prominent feature of decadal variability of North Pacific SSTs is the apparent 1976/77 'climate shift'. SSTs after this shift, relative to the prior period, were warmer in the tropical Pacific and cooler in the central North Pacific around $30°N$ (e.g. Graham 1994, Trenberth and Hurrell 1994). Zhang et al. (1997) have argued that the 1976/77 shift was not unique in the Twentieth Century and instead represents part of a multidecadal oscillation that involves the entire Pacific Ocean and atmosphere. This Pacific Decadal Oscillation (PDO) has spatial patterns of SST variability, sea level pressure and surface wind stress, that are similar to those associated with interannual ENSO variability (Zhang et al. 1997, see also Mantua et al. 1997). The differences are that the PDO has a tropical SST pattern that is broader meridionally than that associated with ENSO and that the magnitudes of the North Pacific and tropical variability are about the same in the PDO while the tropical component dominates for the ENSO variability. Also, for the PDO the North Pacific SST anomaly is more zonally elongated and extends all the way to Japan whereas for ENSO it is centered in the central Pacific and tilts from northwest to southeast (Zhang et al. 1997). The same differences have been noted by Tanimoto et al. (1997).

ENSO clearly has a huge impact on North Pacific climate but there also appears to be a mode of variability that is linearly independant of ENSO (Deser and Blackmon 1995, Zhang et al. 1996). This North Pacific mode is characterized by zonally elongated SSTAs around $30°N - 40°N$ with opposite sign anomalies encircling this region to the northeast, east and southeast. In this mode, when the central North Pacific SST is cool, there is a cyclonic surface wind anomaly representing a strengthening and southeast shift of the Aleutian Low. The SST anomaly pattern is similar to the North Pacific part of Zhang et al.'s (1997) PDO. The time evolution of the North Pacific mode reveals interannual variability, decadal variability and a long term trend consistent with a cooling of the central North Pacific Ocean (Zhang et al. 1996). These two regions of SST anomalies, one in the central Pacific at about $30°N$ and the other further north and west, seem to encompass the important decadal SST variability of the North Pacific and to have different origins (e.g. Nakamura et al. 1997). The 1976/77 climate shift in the North Pacific was possibly associated with both modes.

An active role for the North Pacific Ocean would consist of generation of SST anomalies by either advection by the mean flow (e.g. Saravanan and McWilliams 1998), a strengthening of the gyre circulation (which is argued for by Latif and Barnett 1994, 1996) or a shift in the locations of the subtropical and subpolar gyres. In both the North Atlantic and the North Pacific the subpolar and subtropical gyres are divided by a front in the SST field along which the Gulf Stream/North Atlantic and Kuroshio extension currents flow. It is well established that the latitude at which the western boundary currents leave the coast and their subsequent path into the ocean interior is largely dictated by the pattern of wind stress over the ocean to the east (Parsons 1969, Veronis 1973). It would therefore be expected that, as the wind stress pattern varies, the latitude of separation and the path into the interior will vary. This adjustment is achieved by Rossby waves propagating west and will take some time. In the regions where the gyre location varies, occupancy is switched between subpolar and subtropical waters and we would expect large changes in SST to occur. We would also expect these SST anomalies to be meridionally confined and longitudinally extended. In contrast, SST anomalies associated with changes in gyre strength would be expected to be small since, to first order, the circulation flows along isotherms.

Changes in the gyre locations are known to occur in both the North Atlantic and North Pacific. In the North Atlantic Gangopodhay et al. (1992) have related the Gulf Stream separation latitude to the wind stress forcing of the previous three years and Taylor and Stephens (1998) have found that the latitude of the north wall of the Gulf Stream lags two years behind the NAO. Anomalous southward intrusions of the subpolar waters of the Oyashio current are familar to Japanese oceanographers and have been related to prior anomalous wind forcing (Sekine 1988, 1999, Yamada and Sekine 1997)

Czaja and Marshall (2001) have invoked shifts in the gyre locations in their theory of North Atlantic decadal variability but, to our knowledge, it has not yet been clearly invoked in the Pacific. Existing theories of decadal variability of the North Pacific have often appealed to variations in gyre strength (e.g. Latif and Barnett (1994, 1996) using the coupled general circulation model (GCM) of the Max Planck Institut für Meteorologie (MPI) or Venzke et al. (2000), using a forced ocean GCM in a follow up study) but the exact mechanisms responsible for generating SST anomalies have remained obscure. Recently, Schneider et al. (2001, see also Miller and Schneider 2001) have demonstrated that within the MPI coupled GCM, North Pacific SST anomalies east of Japan are indeed created by a latitudinal shift in the gyre confluence which lags behind the wind forcing.

The purpose of the current paper is to examine the mechanisms that control the interannual and longer timescale variability of North Pacific SSTs. In particular we will look for changes in SSTs that are related to changes in the gyre circulation that are delayed behind changes in wind forcing. Demonstration of the ability of the ocean to slowly respond to wind forcing and create a lagged SST response would be one step towards demonstrating that coupled modes of mid-latitude climate variability do indeed exist. The next step, not yet taken, must show that the mid-latitude atmosphere will subsequently respond to these

SST anomalies.

We will use a long run of an ocean GCM coupled to a simple model of the atmospheric mixed layer (AML) and forced by the observed wind speeds and direction. The current work expands on previous ocean GCM studies (Miller et al. 1994, 1998, Venzke et al. 2000) by using an ocean GCM coupled to the AML model and performing a detailed breakdown of the mechanisms of SST variability. The ocean GCM-AML model computes ocean temperatures and currents as well as the SST, surface fluxes and the atmospheric mixed layer air temperature and specific humidity. This experimental arrangement allows us to determine the relative influences that ocean heat transports and surface heat fluxes exert on the SST variability. In addition we can decompose the ocean heat transport into meridional, zonal and vertical advection and also decompose into parts associated with the locally forced Ekman circulation, and the nonlocal barotropic and baroclinic response to variable wind stress. We will focus on the causes of the mid-Pacific cooling that began in the 1970s, and the cooling off Japan that began in the early 1980s, and compare the modeled SST shifts to those observed. The ability to directly compare the model results against the historical record, and to separate the model response into its component parts, provides an ideal means to assess whether the ocean plays an active role in North Pacific climate variability. In particular we will look for a delayed adjustment of the latitudinal extents of the subpolar and subtropical gyres and creation of SST anomalies near their confluence. This will allow an assessment of whether adjustments of the location of the subtropical-subpolar gyre confluence can account for the observed SST variability east of Japan.

2. Model description

The Lamont ocean GCM has been described in several recent publications (Visbeck et al. 1998, Seager et al. 2000, 2001). For these experiments the GCM spans the Pacific Ocean down to $40°S$ with a variable meridional resolution of $0.5°$ near the equator to $2°$ in mid-latitudes and a fixed $2°$ resolution in longitude. There are 23 fixed vertical levels, 15 of which are in the upper 300m. The model includes basin geometry and bathymetry consistent with the resolution. Temperature and salinity are restored to climatology at the southern boundary. The model includes a simple one and a half layer thermodynamic sea ice model, a bulk wind-driven mixed layer model, convective adjustment and isopycnal thickness diffusion.

The ocean GCM is coupled to a simple model of the atmospheric mixed layer (AML) described by Seager et al. (1995). It represents the well-mixed layer that underlies the cloudy portion of the marine boundary layer (e.g. Augstein 1978, Betts 1976). It computes the air temperature and air humidity by balancing the surface fluxes, advection, atmospheric eddy transports, entrainment from above and radiation. In nature, atmospheric temperature and humidity are closely tied to the SST so that specifiying them in an ocean model's heat flux boundary conditions ensures that the model will reproduce the observed SST. Computing the air temperature and humidity is essential if the SST is not to be overly constrained. The ocean GCM-AML model has been used successfully in similarly motivated studies of Atlantic climate variability (Seager et al. 2000, 2001).

The ocean GCM-AML model is dynamically forced by a combination of NCEP-NCAR reanalyzed climatological surface wind stress (Kalnay et al. 1996), wind speed and direction and anomalies of these quantities taken from DaSilva et al. (1994) after removing a basin mean trend[1]. The period covers 1945 to 1994 but we discard the first five years to allow the model time to adjust from the climatological initial conditions. We use International Satellite Cloud Climatology Project estimates of the surface solar radiation (Bishop and Rossow 1991) and cloud cover. The cloud cover is used in the computation of the longwave cooling of the surface according to a bulk formula (Seager and Blumenthal 1994). The solar radiation and cloud cover are held at their climatological, seasonally varying, values. Therefore only changes in surface wind stress, speed and direction can create SST anomalies.

[1]Generally we always use NCEP-NCAR Reanalysis winds for the climatological forcing to avoid the need to spin up the model each time we change the anomalous forcing. Anomalous winds from different wind products are then added to the NCEP-NCAR climatology to create the total winds that are used to force the ocean model. In this study we did not use NCEP-NCAR anomalies due to conspicuous temporal inhomogeneities found in the surface variables. The DaSilva et al. data appear more homogeneous in time except for an overall positive trend in the wind speed which was corrected by removing a basin mean trend at each grid point.

3. The observed shift in North Pacific climate from the 1960s through to the 1980s

The focus of our analysis will be the climate changes that occured in the North Pacific region centered on 1976/7. The atmospheric circulation and SST were distinctly different before and after this shift (Trenberth and Hurrell 1994, Mantua et al. 1997). While significant changes occured in the central Pacific during 1976/7 itself there were also many subsequent changes, with SST changes occuring east of Japan only in the early 1980s. These changes will be taken to represent the decadal climate variability studied here.

Around 1976 the climate of the entire Pacific Ocean sector went through a noticeable shift. The easiest way to see this is to look at the difference of climate variables between periods before and after the shift. For example Zhang et al. (1997) present a map of Pacific SST for the period 1977 to 1993 minus the period 1950 to 1976. The SST differences look much like a 'fat-ENSO' in that SSTs warmed in the equatorial and eastern subtropics, and along the North American coast up to Alaska, but cooled in the mid-latitude North Pacific centered around $40°N$. Deser et al. (1999) looked at the difference, in just the extratropical North Pacific, between 1977 to 1988 and 1968 to 1976. Their SST change is different from Zhang et al.'s in that the North Pacific SST cooling shows two lobes, one stretching east along $40°N$ from Japan and another in the central Pacific around $30°N$. These two regions also emerge as the first two EOFs of decadal Pacific SST variability in the analysis of Nakamura et al. (1997). The two epochs we choose are the period from 1982 to 1990 and the period 1967 to 1975. These two were chosen to allow time for the ocean circulation to adjust to the mid 1970s change in wind stress forcing.

How did this decadal timescale transition occur? In Figure 1a we show the time history of observed SST and zonal wind speed anomalies across $30°N$ and in Figure 2a we show the time history of SST anomalies and the curl of wind stress across $40°N$. (Plotting wind anomalies at $30°N$ and curl anomalies at $40°N$ anticipates our conclusion that the SST anomalies are generated by different mechanisms in the two regions.) The data are from Da Silva et al. (1994). Looking first at $30°N$ we see that cool SSTs began in 1977 in the central Pacific associated with increased zonal winds and remained until the late 1980s, and were replaced by warm anomalies in 1989. During this period westerly anomalies were more frequent than not but winters of easterly anomalies corresponded with warmings that interrupted the general tendency towards cool waters. The data here are consistent with local, instantaneous, atmospheric forcing of the SST. At $40°N$ the SSTs were also generally cool after 1977 but developed into a protracted period of stronger cooling near the western ocean boundary from 1983 through 1989. This contrasted with generally warm SSTs from the mid 1960s to 1975. Although it is dominated by atmospheric noise, it is discernible that the wind stress curl tended to be more negative in the earlier period and more positive in the later period. The shifts in curl are consistent with the shifts in zonal wind at $30°N$. Consequently the persistently anomalous cyclonic flow over the Pacific went hand in hand with the persistence of the SST anomalies east of Japan and in the central Pacific.

Looking at the raw data in this way makes clear just how tenuous the concept of climate regimes is. Certainly the astute observer can see that the North Pacific in the mid 1980s was different to that in the 1967 to 1975 period. However it is equally clear that this 'regime shift' or epochal change was a quite small signal amongst a tremendous amount of year to year noise. Within each epoch, the wind forcing readily switches sign from year to year. The decadal variations of SST are more noticeable than the decadal variations of the winds, testifying to the ocean's ability to integrate the atmosphere's higher frequency variability, but considerable interannual variations of SST remain.

Figure 3a shows the difference in observed SSTs, surface winds, and surface latent plus sensible heat flux for the period 1982 to 1990 minus the period 1967 to 1975. The surface wind shift is characterized by a cyclonic circulation that strengthens the mean circulation but also shifts it somewhat to the south. SST cooled in two regions: a zonal strip stretching east of Japan at $40°N$ and a central Pacific region at around $30°N$. This is very similar to the pattern of Deser et al. (1999). In the central Pacific the anomalous surface fluxes are not well matched to the SST anomaly. Between $30°N$ and $50°N$ the flux anomalies are forcing the SST anomalies but west and south of this area they damp the SST anomaly. In particular fluxes are damping the portion of the zonal strip of SST anomalies immediately east of Japan and, by implication, variations in ocean heat transport must be the forcing.

In summary, it might be argued that anomalous cyclonic flow over the Pacific immediately generates cool waters in the central Pacific at about $30°N$ and, at a later date, cool waters further north at $40°N$ east of Japan. The central Pacific response is expected as a result of increased winds either cooling the ocean by increased surface heat loss or by equatorward Ekman advection while the later, delayed and remote response, needs explanation.

4. The response of the ocean GCM-AML model to varying winds

Figures 1b and 2b show the time history of modeled SSTs along $30°N$ and $40°N$ where they can be compared with the observations. The match is overall quite reasonable although the modeled SST anomalies tend to be weak and at $40°N$ they tend to be too confined towards the east.

Figure 3b shows the changes in observed winds and modeled SST and surface fluxes, for 1982 to 1990 minus 1967 to 1975, from the OGCM-AML model. Broadly speaking the SST change is consistent with that observed. However, SST anomalies are stronger than observed to the east of Japan (this is a result of the choice of averaging period and is not general) and weaker elsewhere and the model fails to join together the northwestern and central Pacific anomalies. In the region east of Japan the modeled fluxes damp the cold SST anomaly as observed. Figure 3c shows the change in ocean heat transport, integrated down to mixed layer depth, and model SST. Clearly the zonal anomaly off Japan is created by a change in ocean heat transport. How is the ocean creating these heat transport and SST anomalies? First we separated the changes in ocean heat transport into horizontal and vertical components. The total change was dominated by the horizontal terms. Next we split the horizontal transport into the mean currents advecting the anomalous SST, the anomalous currents advecting the mean SST and anomalous currents advecting anomalous SSTs. The last, nonlinear, term was universally small. Additional terms involving anomalies in mixed layer depths were not crucial to establishing the pattern of SST anomalies. We found that advection of anomalous SSTs by the mean currents (not shown) acts to move SST anomalies in the region east of Japan further east along the Kuroshio-Oyashio extension. Advection of the mean SST by the anomalous currents (also not shown, but see below), on the other hand, was clearly the process creating the SST anomaly in the Kuroshio-Oyashio extension region.

Next we need to discover which components of the time varying ocean circulation are responsible for changing the ocean heat transport and creating the SST anomaly. We will attempt to divide this impact into a part that is rapid and a part that involves the slow adjustment of the ocean to changing wind stress.

a. Decomposition of the ocean response to variable wind stress into the Ekman, barotropic and baroclinic components

The ocean's response to a variable wind stress can be divided into three parts. First the wind stress anomalies drive an anomalous Ekman drift that is largely confined to the mixed layer and which is set up essentially instantaneously. Anomalous Ekman convergence and divergence create a pressure gradient that is uniform with depth and which drives a barotropic anomalous current. In addition there is a baroclinic response associated with pressure gradients that vary with depth. The anomalous baroclinic currents have zero vertically averaged motion. The barotropic flow is governed by the vertical integral of the momentum equations. Let the zonal and meridional currents be comprised of a barotropic component, $\bar{u}$ and $\bar{v}$ and a baroclinic component u' and v' with $u = \bar{u} + u'$ and $v = \bar{v} + v'$ and

$$\bar{u} = \frac{1}{D}\int_{-D}^{0} u dz, \tag{1}$$

$$\bar{v} = \frac{1}{D}\int_{-D}^{0} v dz, \tag{2}$$

where D is the variable depth of the model ocean. The barotropic flow is subtracted from the total flow in the mixed layer to derive the baroclinic flow. Next we divide the baroclinic flow in the mixed layer into contributions from the Ekman current and those related to vertical and horizontal variations in density. The Ekman adjustment is fast while the remainder of the baroclinic adjustment is slow. The Ekman flow is given by:

$$v_{Ek} = -\frac{\tau^x}{\rho_0 f H}, \tag{3}$$

$$u_{Ek} = \frac{\tau^y}{\rho_0 f H}, \tag{4}$$

where H is the mixed layer depth. Subtracting the Ekman flow from the baroclinic flow allows us to derive the slowly adjusting baroclinic flow, u_s and v_s, in the mixed layer, as a residual using $u_s = u' - u_{Ek}$ and $v_s = v' - v_{Ek}$.

b. Division of the SST forcing into parts associated with the anomalous Ekman, barotropic and slow baroclinic responses to the wind stress shift

Once the anomalous zonal and meridional Ekman, barotropic and slowly adjusting baroclinic currents are computed as above, they are multiplied by the zonal and meridional mean temperature gradients, and $\rho c_p H$, to derive the associated heat transport anomaly in the mixed layer in Wm^{-2}. Figure 4 shows the heat transport shift associated with the Ekman, barotropic and slowly adjusting baroclinic responses, overlaid on the SST shift. Anomalous Ekman currents flow equatorward, forced by the anomalous westerlies, in the central Pacific region around $30°N$ and account for the cooling there. The Ekman cooling is sufficiently large that the surface fluxes damp the SST anomaly in some of this region. This is despite the fact that the winds strengthen here and increased surface fluxes alone would cool the ocean. Apparently the Ekman cooling is sufficiently large that it drives the SST below what the increased surface fluxes alone would allow and, therefore, the surface fluxes begin to damp the SST anomaly.

In the Kuroshio-Oyashio extension region matters are more complex. The altered heat transport due to the shift in Ekman currents is small here. The shift in the barotropic flow is locally important immediately off the coast of Japan where a southward anomalous boundary current introduces a strong cooling. Elsewhere the change in heat transport by the slowly adjusting baroclinic flow is largely responsible for creating the zonal strip of cooling along

$40°N$. The shift in the total current is shown in Figure 5 and its components in Figure 6 together with the shift in SSTs. As expected, the shifts in the barotropic and baroclinic flows are spatially very similar although the baroclinic component is much the larger. Both show increased eastward flow along about $35°N$ which is about 5° to the south of the climatological mean position of the Kuroshio-Oyashio extension in our model. The change in currents, heat transport and SSTs can be understood as a southward shift in the confluence of the subtropical and subpolar gyres. Subtropical waters no longer reach to $40°N$ and instead the zonal strip at that latitude becomes occupied by subpolar waters and cools.

The changes in the barotropic and baroclinic circulations and the gyre locations are consistent with the change in the wind forcing. The change in the atmospheric circulation places the maximum of the shift in the surface westerlies at about $30 - 35°N$. The shift in wind stress curl is therefore zero at that latitude. The shift in curl is positive to the north, driving a northward interior flow and southward boundary flow, and negative to the south, driving a southward interior flow and northward boundary flow. Consistent with this simple reasoning, the shift in the model circulation looks like an 'inter-gyre' gyre or a southward displacement of the subtropical-subpolar confluence and Kuroshio- Oyashio extension. The two locations of SST cooling are also consistent with this theory. Increased westerlies in the central Pacific around $35°N$ locally cool the SSTs by anomalous Ekman advection. The same wind forcing shift, with some delay, leads to a relocation of the gyre circulations that introduce a cooling to the immediate north, at $40°N$, of the confluence of the intergyre gyre. Since the confluence of the intergyre gyre occupies much the same latitude as the central Pacific cooling the relatively more northerly position of the SST cooling north of the mean Kuroshio-Oyashio extension is expected. Therefore the entire SST shift is consistent with, in the central Pacific, a local and instantaneous ocean response and, east of Japan, a delayed and remote response of the ocean to the same wind forcing.

5. Comparison with results from a mixed layer ocean model

If we are correct that the delayed response of the wind driven gyres creates the SST anomaly in the region east of Japan, then it should not occur in an ocean model that neglects variations in OHT. To check this we coupled the AML model to a uniform 75m deep slab ocean mixed layer (OML) model. The OML-AML model is forced by the detrended DaSilva winds. The SST anomalies in the OML-AML model are created only by anomalies in the surface sensible and latent heat fluxes. The flux anomalies are the difference between the total fluxes and the climatological fluxes. The total fluxes are computed using the observed climatological SSTs plus the modeled SST anomaly and the observed full winds. The climatological fluxes are computed using the climatological observed SSTs and the climatological winds (see Seager et al. (2000) for more details).

Figure 7 shows the change in SST and surface winds from the AML-OML model for the 1982 to 1990 period minus that of the 1967 to 1975 period. The OML-AML model succeeds in reproducing the cooling in the central Pacific, where it must be forced by increased surface heat loss associated with increased wind speed (rather than by Ekman drift), but fails to produce the cooling east of Japan.

6. The timescale of the delayed response: model and observations

We have demonstrated that the SSTs in the Kuroshio-Oyashio extension region east of Japan change as a delayed response to winds over the central Pacific. What is the timescale of the delay? To look at this we created a timeseries of the observed SST anomaly in the region east of Japan bounded by $34°N$ and $46°N$ and $139°E$ and $180°E$. Regression with the wind stress curl over the North Pacific indicated that the SSTs are influenced by the winds at different lags up to about five years previously. We therefore applied a three year running mean to the curl of the wind stress at each grid point over the North Pacific. The filtered time series of wind stress curl were then regressed together with the SST timeseries using different lags of the SST behind the winds. The highest correlations occur with the observed SST lagging three years behind the observed wind stress curl. The regression pattern for a three year lag is shown in Fig. 8a. Negative SST anomalies in the region east of Japan are associated with positive wind stress curl over most of the mid-latitude North Pacific. Positive curl would drive a northward interior Sverdrup drift and require a southward western boundary flow, an equatorward shift of the subtropical-subpolar gyre confluence and subsequent cooling in the region east of Japan. Next we zonally averaged the filtered wind stress curl across $40°N$, through the center of the region of large regression coefficient. In Fig. 8b we show this curl time series, multiplied by minus one, together with the SST index, with the latter moved three years backward. The general agreement between the time series is impressive with the shift towards cooler SSTs in the 1980s clearly being associated with the shift to more positive wind stress curl around $40°N$ in the preceding years. This time delay is close to the five years identified by Schneider et al. (2001).

Fig. 9 shows the equivalent regression pattern and timeseries for the OGCM-AML model. The spatial patterns, and the temporal evolution of the relationship between curl of wind stress and SSTs east of Japan, are very similar between the model and the observations indicating that the essential aspects of the delayed adjustment of the gyre circulations to the wind forcing are being captured.

As a final demonstration of the temporal evolution of decadal SST anomalies in the North Pacific we performed a lag correlation analysis of winds and SSTs. In order to have the longest period of data possible we performed this analysis using NCEP-NCAR reanalysis winds and SSTs for 1948 to the present (the problems of spurious trends in the tropical winds is not a problem in this analysis as it would be if we used these winds to force the model). We perfored two lag correlations. The first was between the SSTs in the central Pacific $(27.5°N - 32.5°N, 170°E - 140°W)$ and the Kuroshio-Oyashio extension region $(37.5°N - 42.5°N, 150°E - 180°E)$. The second was between the curl of wind stress over the north central Pacific $(37.5°N - 42.5°N, 170°E - 140°W)$ and SSTs in the Kuroshio-Oyashio extension region. Results are shown in Fig. 10. The SST correlations are shown by the solid line with positive lag meaning the central Pacific regions leads. It is clear that SST anomalies develop first in the central Pacific and that anomalies in the Kuroshio-Oyashio extension region lag behind for the next few years. Similarly the wind stress curl leads the SSTs in the Kuroshio-

Oyashio extension region by the same amount. Clearly central Pacific SSTs develop before those east of Japan and there is no indication that the latter subsequently influence the central Pacific anomalies.

The delayed response of the gyre circulation and SSTs to the time varying wind forcing is often thought of in terms of the quasi-geostrophic potential vorticity equation for a baroclinic ocean. In terms of the horizontal streamfunction, ψ, this can be written:

$$r\psi + \left[\nabla^2 - \lambda^{-2}\right]\psi_t + \beta\psi_x = \alpha curl\tau, \tag{5}$$

where α is the Lighthill coefficient, r is an inverse decay timescale, $\lambda = c/f$ is the Rossby radius of deformation with $c = (gh_e)^{\frac{1}{2}}$ as the baroclinic wave speed where h_e is the equivalent depth. In mid-latitudes for large scale wind forcing $\nabla^2 \ll \lambda^{-2}$ so:

$$r\psi + \psi_t - c_R\psi_x = -\alpha\lambda^{-2} curl\tau, \tag{6}$$

where $c_R = \beta\lambda^2$ is the Rossby wave speed. The solution is:

$$\psi(x,t) = \frac{\alpha\lambda^{-2}}{c_R}\int_{x_E}^{x} e^{-r\left(\frac{x'-x}{c_R}\right)} curl\tau\left[x', t - \left(\frac{x'-x}{c_R}\right)\right] dx'. \tag{7}$$

We have assummed that ψ is zero along the eastern boundary which ensures no normal flow. This equation states that the streamfunction, which has the same spatial pattern as the upper ocean heat content and hence, by reasonable inference, as the SST, is determined by the curl of wind stress integrated eastward back along Rossby wave characteristics. Because of dissipation the curl of wind stress is of most influence immediately prior, and immediately to the east, and is of declining importance back, in space and time, along the characteristics. If we assume $c = 3ms^{-1}$ for the first baroclinic mode then, at $45°N$, the Rossby wave speed, $\beta\lambda^2$, is about 7° of longitude per year. On the basis of a five year maximum delay between SST change in the Kuroshio-Oyashio extension region and the wind forcing, the region of wind influence extends only 35° to the east of the western boundary.

However, this simple theory does not work very well. We have solved Eq. 7 in the context of a linear reduced gravity model forced by the detrended DaSilva et al. (1994) wind stresses (this continues work presented by Karspeck and Cane (2001)). The modeled variations in upper ocean heat content (not shown) only modestly resemble those simulated by the OGCM-AML model or observations. It is hard to argue with the applicability of the quasi-geostrophic physics contained within the simple theory. Instead the poor simulation of the single baroclinic mode model probably argues for the importance of both the barotropic response and the higher baroclinic modes. This cautions against placing too much faith in theories of decadal variability that rely on only the first baroclinic mode (e.g. Jin 1997, Neelin and Weng 1999).

7. The nature of the delayed dynamical response of the ocean

Venzke et al. (2000) argue, following Latif and Barnett (1994, 1996), that the delayed response of the ocean to the varying wind stress acts as a negative feedback on North Pacific SSTs. This is attributed to 'anomalously strong horizontal advection'. It is not stated clearly in the text but it appears that the proposed negative feedback works as follows. A stronger circulation around the Aleutian Low immediately cools the waters on its southern flanks, by surface fluxes and equatorward Ekman drift, but also increases the velocities of the subtropical gyre. The gyre takes some time to speed up, but once it has, it brings more warm subtropical water north and then east in the Kuroshio-Oyashio extension region. The warm water eventually spreads through the North Pacific and, therefore, is a negative feedback on the locally wind forced SST anomaly. The problem with this theory is that, to first order, the gyre flow is along isotherms and speeding it up achieves little extra heat flux convergence. As discussed before, a shift in the latitude dividing the subtropical and subpolar gyres is a more likely way to create noticeable SST anomalies.

In our theory of wind-induced shifts in the latitudes where the subpolar and subtropical gyres meet, the delayed response of the ocean is a positive feedback on the initial locally forced SST anomalies. As the Aleutian Low strengthens and moves slightly southeast, it forces a central Pacific cooling at about $30°N$. The delayed response involves a southward shift of the subpolar-subtropical gyre boundary and creation of a cold anomaly further west at about $40°N$. This delayed SST anomaly could advect east and south and, if it did, would reinforce the original central Pacific anomaly. Thus, to the extent that the two areas interact, the purely ocean response is a positive feedback. However, as we have shown, there is no evidence that central Pacific anomalies are subsequently influenced by anomalies in the Kuroshio-Oyashio extension region. The cold anomaly created as a delayed response to the winds is too far south to change the sign of the warm anomaly that local wind forcing places to the north (as well as east and south) of the central Pacific SST anomaly. A negative feedback would only occur if the atmosphere responded to cool waters east of Japan by weakening the Aleutian Low and shifting it back northwestward.

8. Summary and Conclusions

We have presented the results of an hindcast of surface ocean circulation and SST completed with an ocean GCM coupled to a simple thermodynamic model of the low level atmosphere and forced with the observed history of surface winds. Our aim was to explain the mechanisms that caused the decadal changes in central and northwest Pacific SSTs that occured in the late 1970s and 1980s.

The results make it clear that the SST changes can be understood in terms of the dynamic and thermodynamic response of the ocean to varying wind forcing. In the central Pacific the strengthening and southward shift of the mid-latitude westerlies that occured in the 1976/77 winter and persisted thereafter caused an immediate cooling due to enhanced southward Ekman drift. Mean SST gradients are quite strong in this area so the enhanced Ekman drift caues a strong cooling of the ocean. The stronger surface winds might be expected to also cool the ocean by increased surface latent heat loss but, apparently, the anomalous Ekman cooling is sufficiently large that it dominates and the surface fluxes provide a mild damping. This is in contrast to the situation in the North Atlantic where surface fluxes and anomalous Ekman drift are about equally responsible for the local cooling in response to strengthened westerlies (Seager et al. 2000). This difference between the two oceans occurs because meridional SST gradients in the Pacific are stronger than those in the Atlantic (the reasons for which remain in dispute).

Some years after the wind shift over the North Pacific, SST anomalies begin to develop in a zonally elongated, but meridionally narrow, area east of Japan. During the 1980s SST anomalies were cold here, and they remained cold in the central Pacific too. In the OGCM-AML model the cold SSTs east of Japan are forced by a reduction in ocean heat flux convergence and are damped by reduced surface heat loss to the atmosphere. We were able to show that the reduced ocean heat flux convergence is associated with a southward shift in the latitude of the confluence between the subpolar and subtropical gyres. The region that became cooler than normal came under the influence of subpolar waters that extended further south along the coast of Japan than in the period before 1976. In terms of the components of the ocean's mixed layer heat budget the anomalous cooling is associated with anomalous meridional advection of the mean SST. The anomalous currents are predominantly accounted for by the slowly propagating part of the baroclinic response of the ocean to varying wind stress.

The southward shift of the gyre confluence is consistent with the southward shift of the westerlies on the southern flank of the Aleutian Low that occured in 1977 and afterwards. The latitude of the line of zero wind stress curl also moved south. Consequently the relocation of the gyre can be understood in terms of Sverdrup theory. Alternatively, application of Veronis' (1973) reasoning would lead from the observation of the strengthening mid-latitude westerlies to the conclusion that the western boundary current will need to separate from the boundary at a more southerly latitude in order that it can still balance the southward flow integrated from its eastern flank to the eastern boundary. Using both the results of the wind-forced OGCM-AML model and observations it is concluded that the gyre adjustment, and generation of associated SST anomalies, depends on the history of prior winds over the previous few years. There is no single time delay but, instead, wind forcing over the previous one to four years appear to be the most important.

Since the SST anomalies created as a delayed response to the central Pacific wind forcing are of the same sign as the locally forced anomalies the delayed feedback is positive.

Furthermore, the SST anomalies associated with the delayed response are of a much smaller meridional scale than the basin scale anomalies locally forced by the atmosphere. It appears unlikely that, by advection, they will be able to subsequently influence the directly forced anomalies. It is abundantly clear that the delayed response of the ocean to the varying winds cannnot provide a negative feedback that will subsequently alter the sign of the SST anomalies locally forced by the atmosphere.

We do not know how the atmosphere will respond to the SST anomalies associated with the gyre adjustment. The gyre adjustment will shift the exact latitude at which the strongest meridional gradients of SST occur but it will only do this by about 5° of latitude which is sub-synoptic scale. It seems unlikely that this subtle shift will have much impact on the atmosphere especially given the very large mean zonal temperature gradients that lie upstream of this location. If there is little feedback of the delayed response of the ocean onto the atmosphere it is nonetheless still true that the gyre adjustment will cause a reddening of the spectrum of ocean climate variability. It is also possible that the gyre adjustment is predictable some years in advance which would prove useful in studies of the ocean's biological resources.

Our results are broadly consistent with the mechanisms of SST variability that Schneider et al. (2001) describe occuring in the MPI coupled GCM. They also see shifts in the latitude of the subtropical-subpolar gyre confluence and the associated creation of SST anomalies. The lag of at most a few years between the wind forcing and the gyre response is consistent between our ocean hindcast and their coupled GCM. The dominance of the locally forced anomalous Ekman drift in the central Pacific is also a result common to the two models. The agreement between the different models and the observations strongly suggests to us that the delayed response of the ocean to variable winds over the North Pacific has now been indentified. It consists of monopoles of zonally elongated, meridionally confined SST anomalies that lie in the region of the subtropical-subpolar gyre confluence that shift the latitude of the maximum SST gradient east of Japan.

Given that the work described here and in Schneider et al. (2001) correctly describes what the North Pacific Ocean is capable of doing as a delayed reaction to the winds where does this leave our understanding of North Pacific decadal climate variability? It now seems highly unlikely that there is a coupled feedback loop in operation in which ocean dynamics play a delayed negative feedback. On the other hand the gyre response will provide a means whereby the spectrum of ocean climate variability can be reddened. If there is a distinct timescale of variability it must come from outside the North Pacific. It is well known that winds over the North Pacific respond strongly to tropical Pacific forcing. The central Pacific cooling that began in 1977 occured in combination with a shift to warmer SSTs in the tropical east Pacific. The demise of cool SSTs in the North Pacific similarly occured with the onset of the strong La Niña of 1989. Decadal variations of the North Pacific Ocean are forced by decadal variations in the surface wind forcing. Nothing we have seen in the current study is inconsistent with the North Pacific Ocean responding both locally and instantaneously, and remotely and delayed, to wind variations that are a combination of intrinsic mid-latitude variability and forcing from the tropical Pacific. But it has not been shown here, or elsewhere, that North Pacific decadal variability is primarily driven from the tropics.

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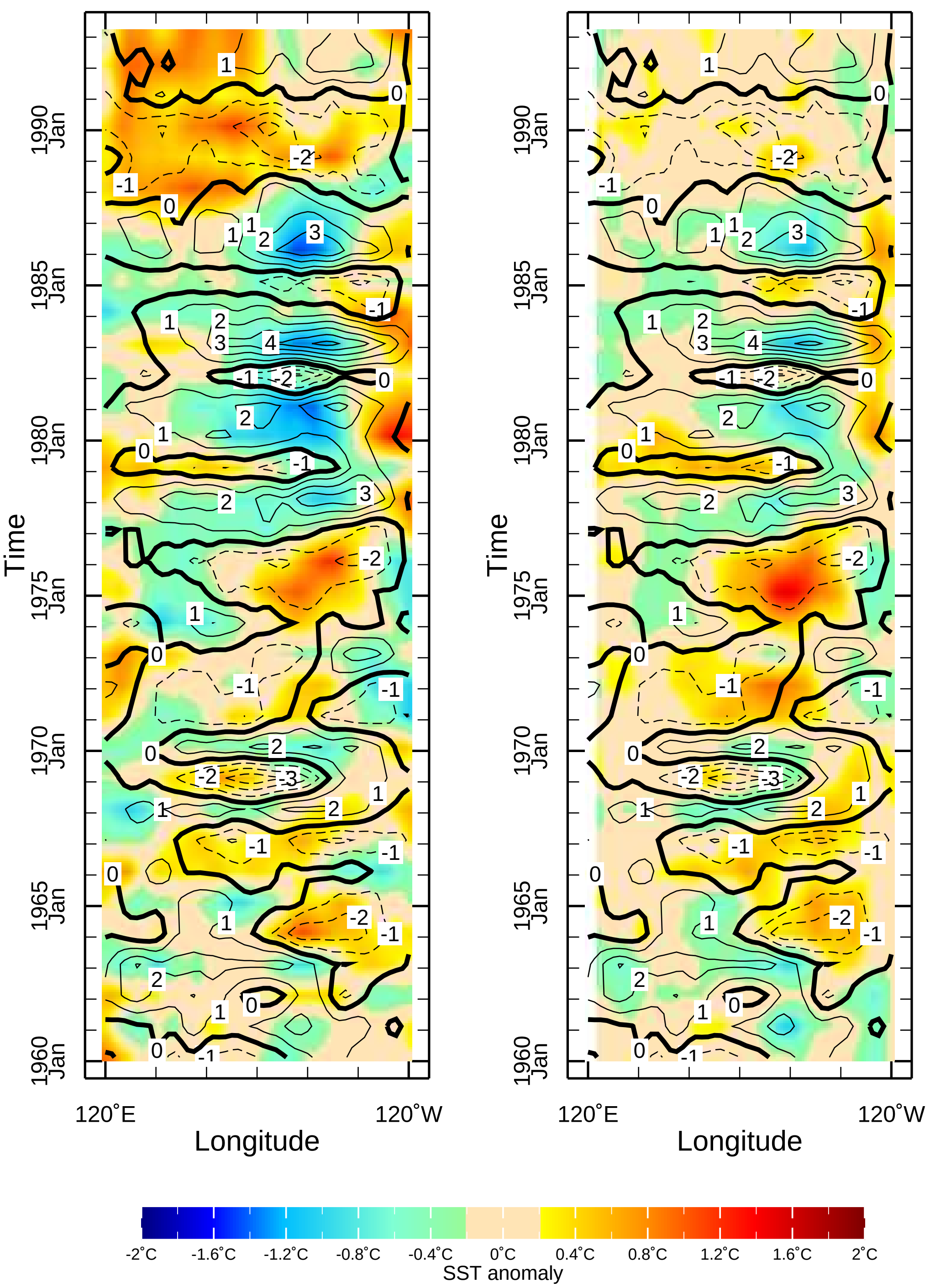
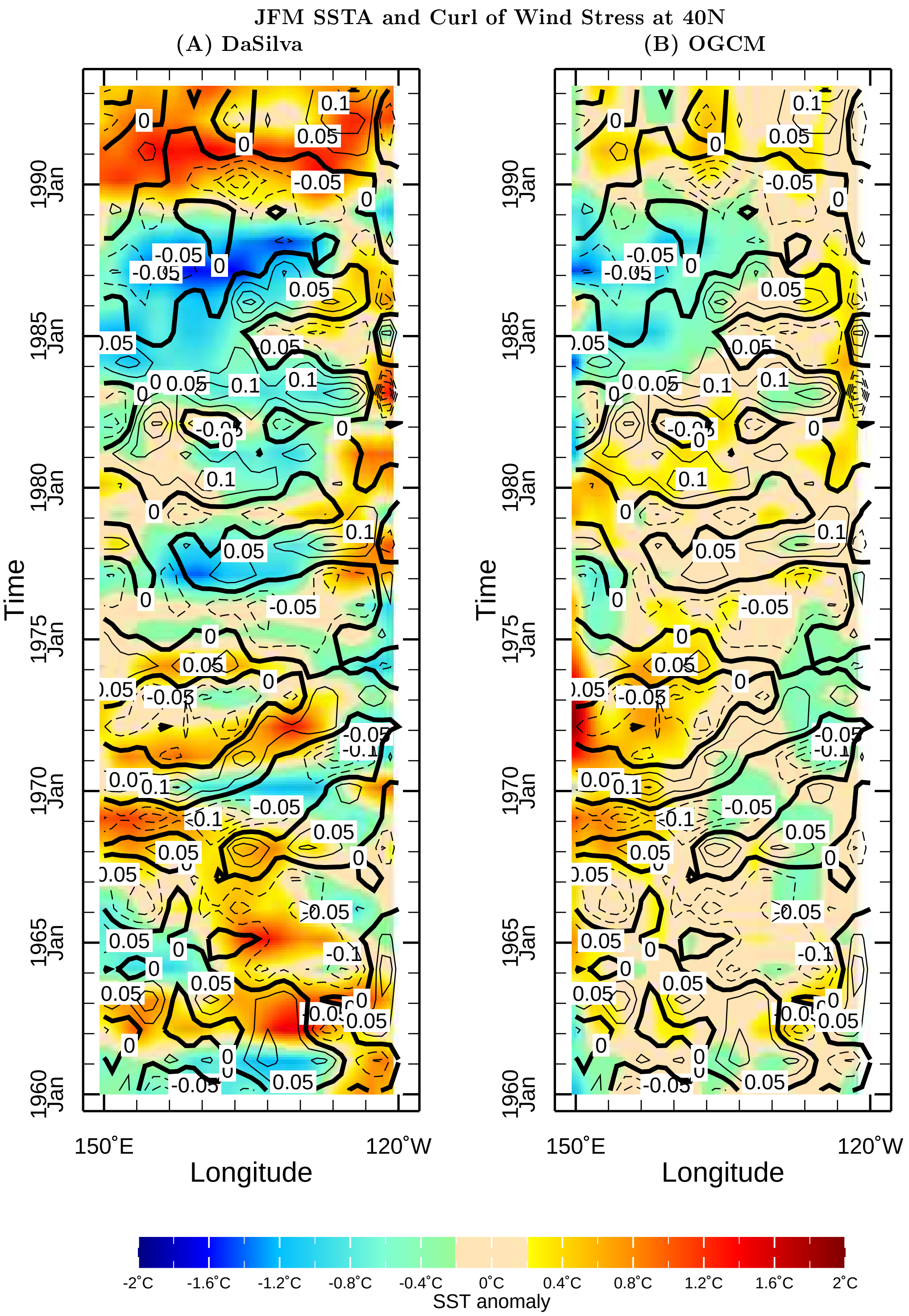
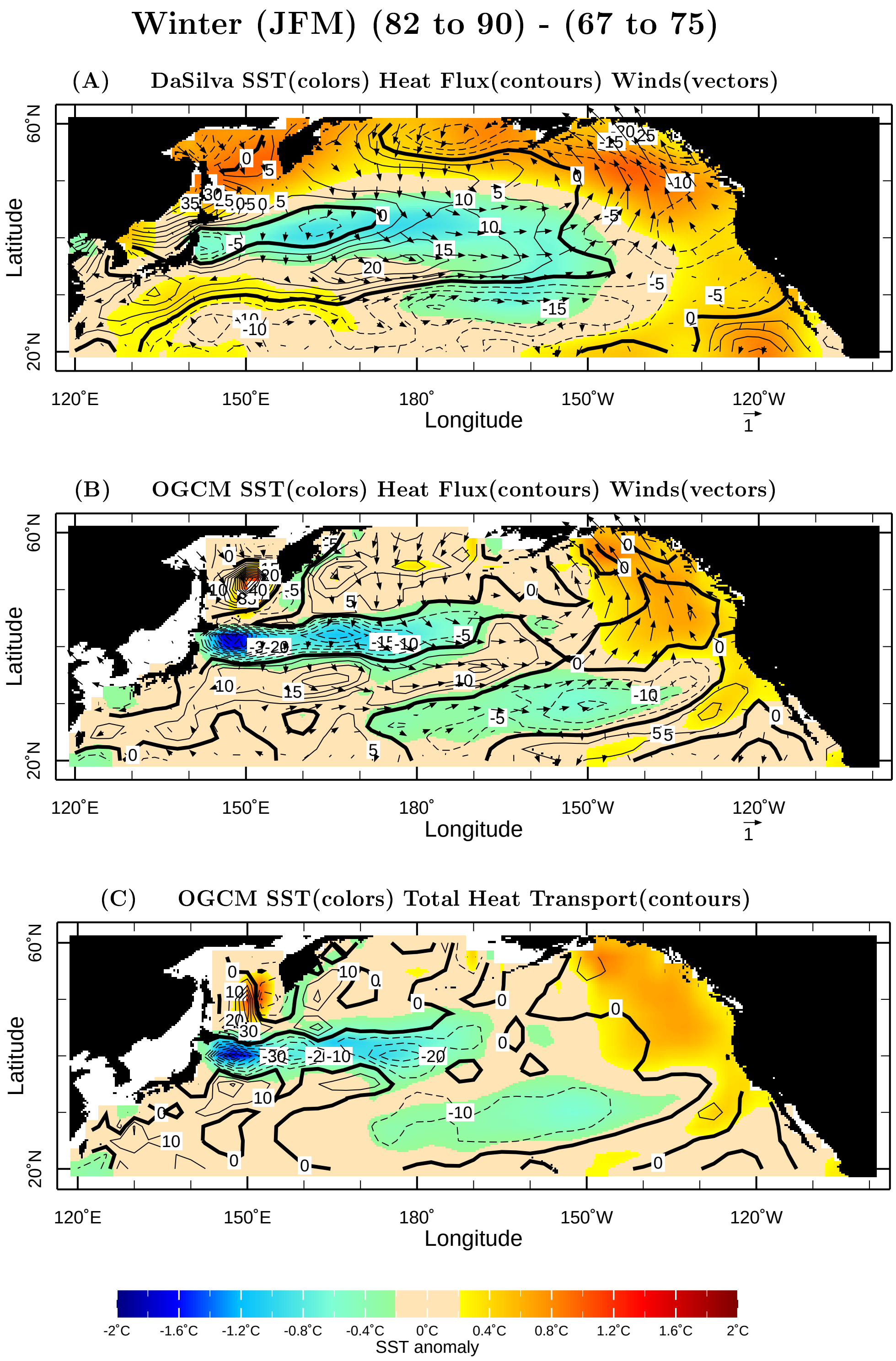
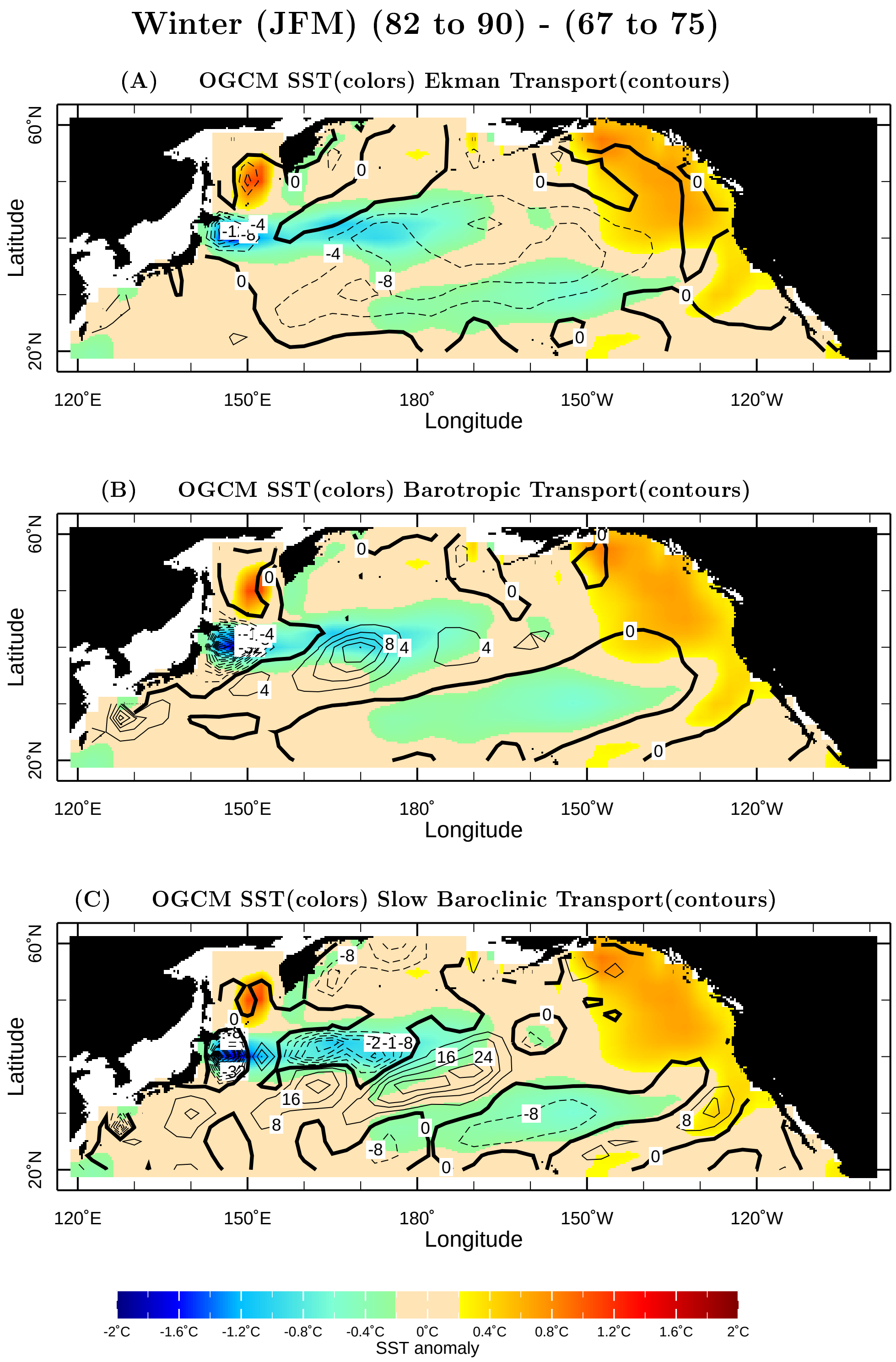
JFM SSTA and Zonal Wind Anomalies at 30N



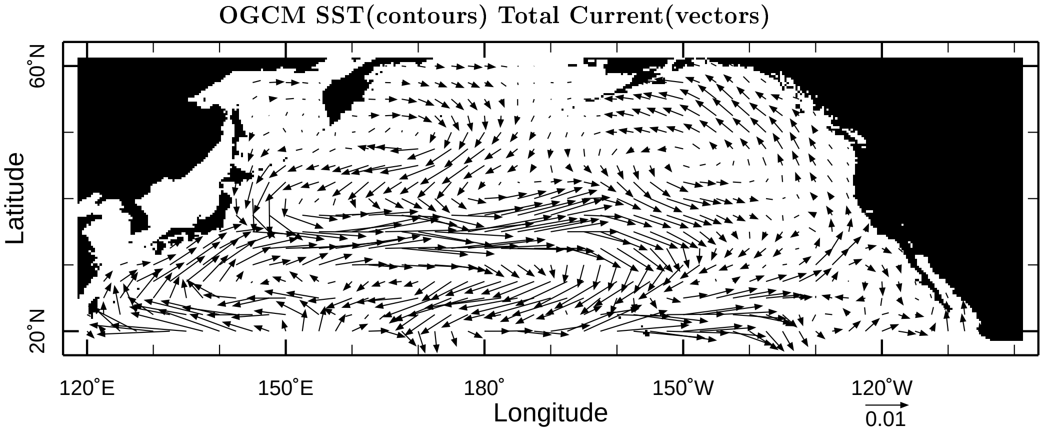


JFM SSTA and Curl of Wind Stress at 40N
(A) DaSilva
(B) OGCM
Time
Jan 1960
Jan 1965
Jan 1970
Jan 1975
Jan 1980
Jan 1985
Jan 1990
150˚E
120˚W
Longitude
-2˚C
-1.6˚C
-1.2˚C
-0.8˚C
-0.4˚C
0˚C
0.4˚C
0.8˚C
1.2˚C
1.6˚C
2˚C
SST anomaly


Winter (JFM) (82 to 90) - (67 to 75)
(A) DaSilva SST(colors) Heat Flux(contours) Winds(vectors)
(B) OGCM SST(colors) Heat Flux(contours) Winds(vectors)
(C) OGCM SST(colors) Total Heat Transport(contours)
Latitude
60°N
20°N
Longitude
120°E
150°E
180°
150°W
120°W
-2°C
-1.6°C
-1.2°C
-0.8°C
-0.4°C
0°C
0.4°C
0.8°C
1.2°C
1.6°C
2°C
SST anomaly

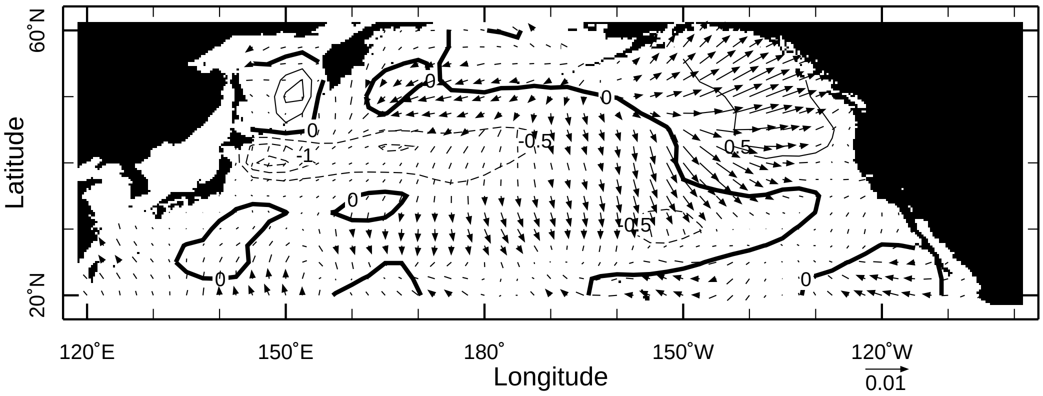

Winter (JFM) (82 to 90) - (67 to 75)
(A) OGCM SST(colors) Ekman Transport(contours)
Latitude
60°N
20°N
120°E
150°E
180°
150°W
120°W
Longitude
(B) OGCM SST(colors) Barotropic Transport(contours)
Latitude
60°N
20°N
120°E
150°E
180°
150°W
120°W
Longitude
(C) OGCM SST(colors) Slow Baroclinic Transport(contours)
Latitude
60°N
20°N
120°E
150°E
180°
150°W
120°W
Longitude
-2°C
-1.6°C
-1.2°C
-0.8°C
-0.4°C
0°C
0.4°C
0.8°C
1.2°C
1.6°C
2°C
SST anomaly

Winter (JFM) (82 to 90) - (67 to 75)

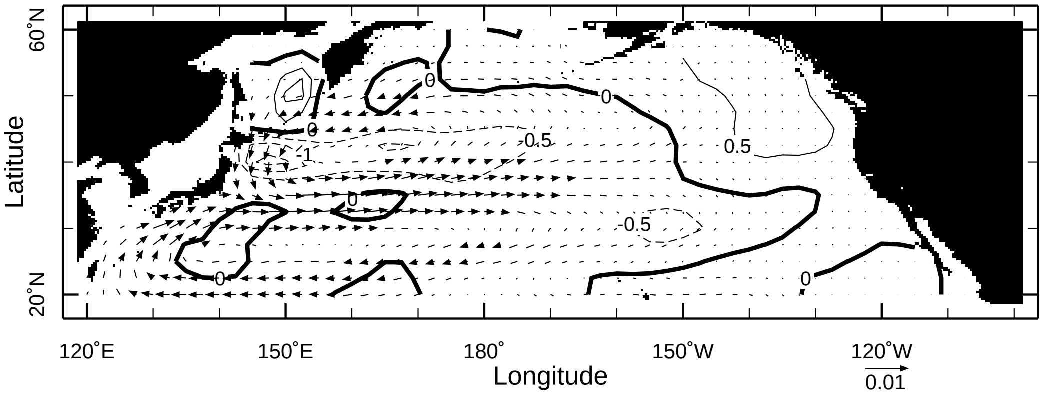


Winter (JFM) (82 to 90) - (67 to 75)

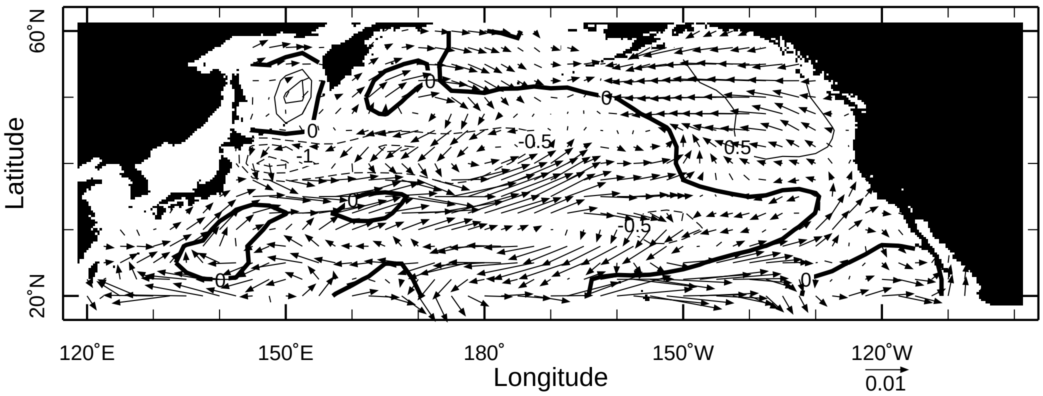
(A) OGCM SST(contours) Ekman Current(vectors)



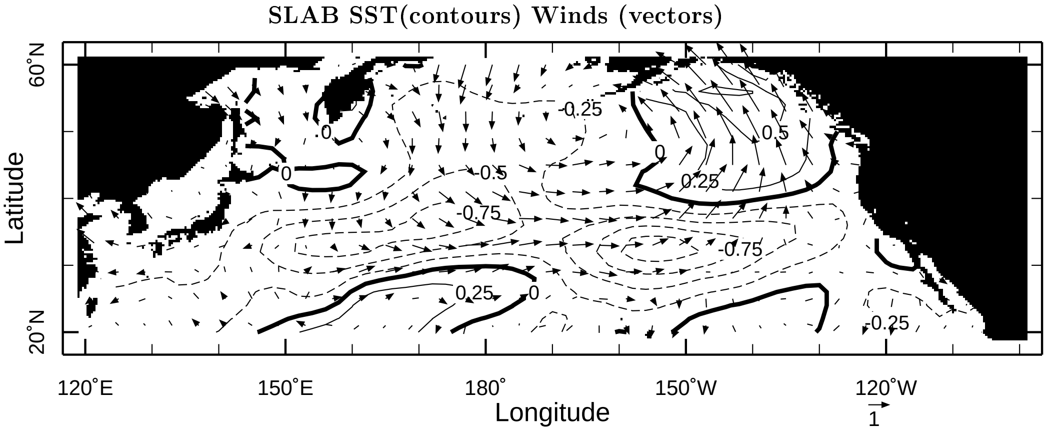
(B) OGCM SST(contours) Barotropic Current(vectors)

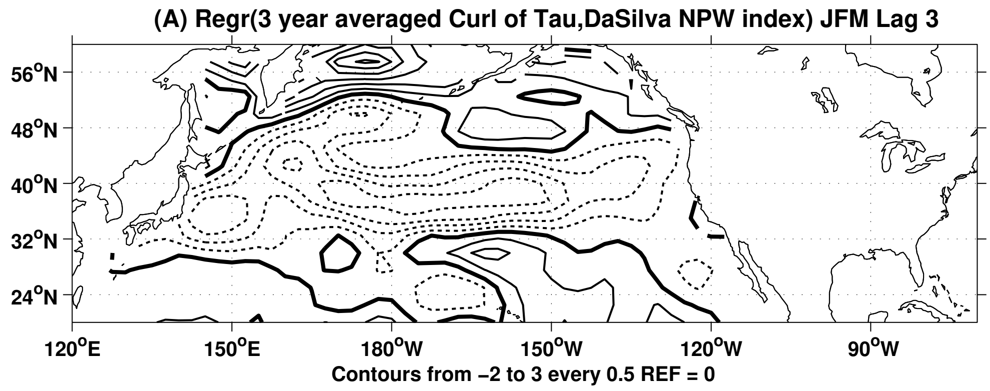


(C) OGCM SST(contours) Slow Baroclinic Current(vectors)

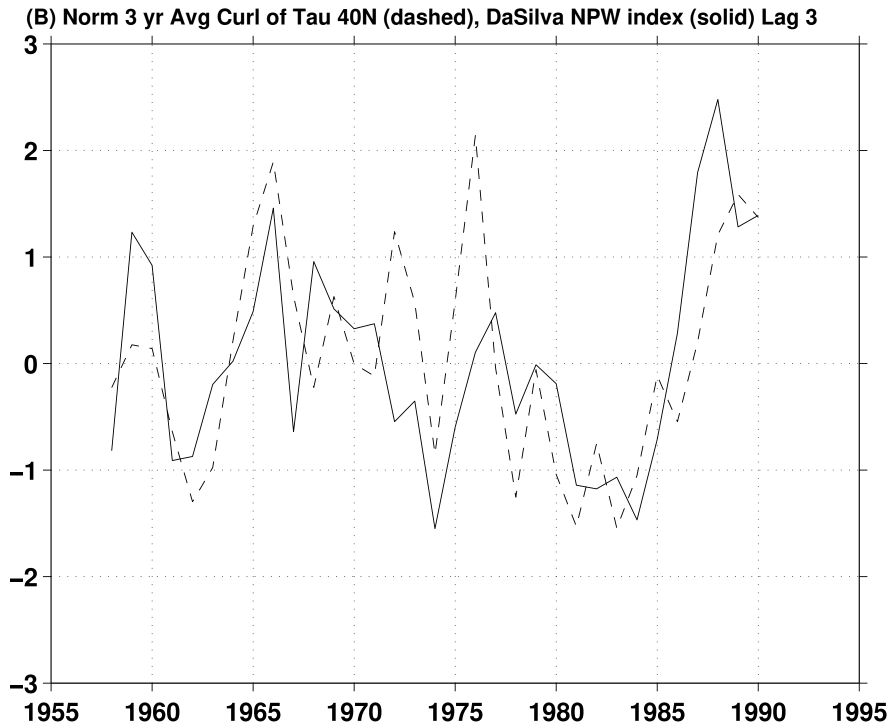


Winter (JFM) (82 to 90) - (67 to 75)

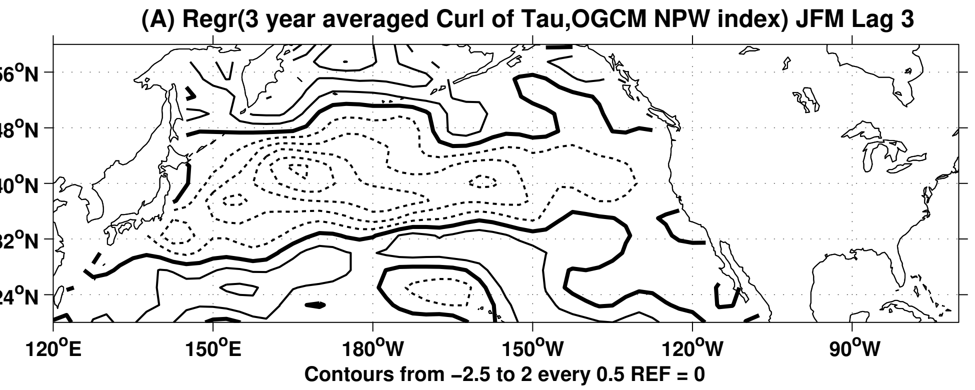


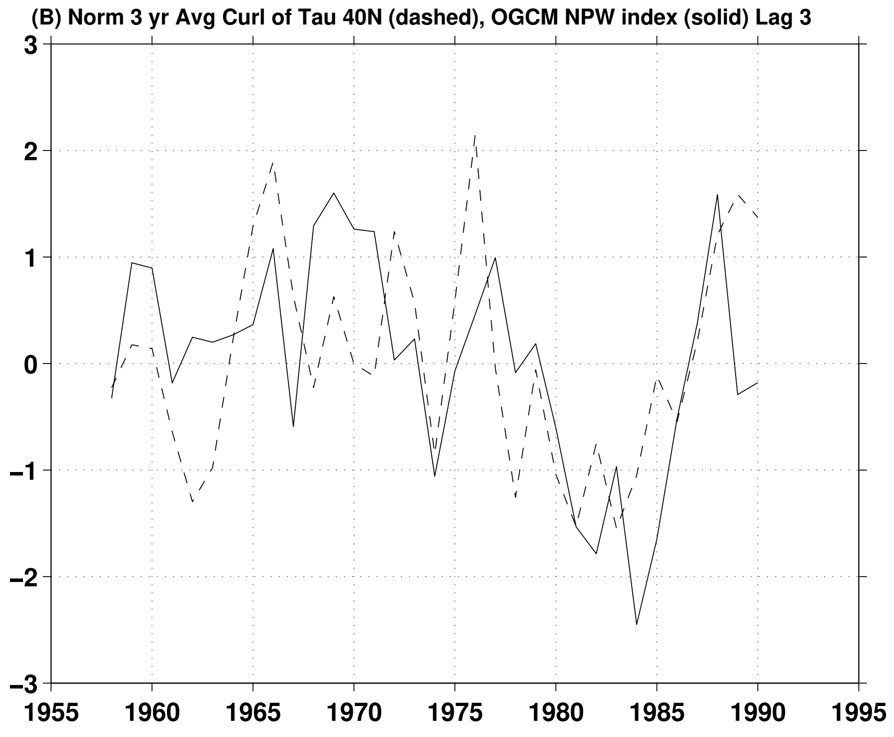


(A) Regr(3 year averaged Curl of Tau,DaSilva NPW index) JFM Lag 3
56°N
48°N
40°N
32°N
24°N
120°E
150°E
180°W
150°W
120°W
90°W
Contours from −2 to 3 every 0.5 REF = 0



(B) Norm 3 yr Avg Curl of Tau 40N (dashed), DaSilva NPW index (solid) Lag 3
3
2
1
0
−1
−2
−3
1955
1960
1965
1970
1975
1980
1985
1990
1995


(A) Regr(3 year averaged Curl of Tau,OGCM NPW index) JFM Lag 3
56°N
48°N
40°N
32°N
24°N
120°E
150°E
180°W
150°W
120°W
90°W
Contours from −2.5 to 2 every 0.5 REF = 0


(B) Norm 3 yr Avg Curl of Tau 40N (dashed), OGCM NPW index (solid) Lag 3
3
2
1
0
−1
−2
−3
1955
1960
1965
1970
1975
1980
1985
1990
1995

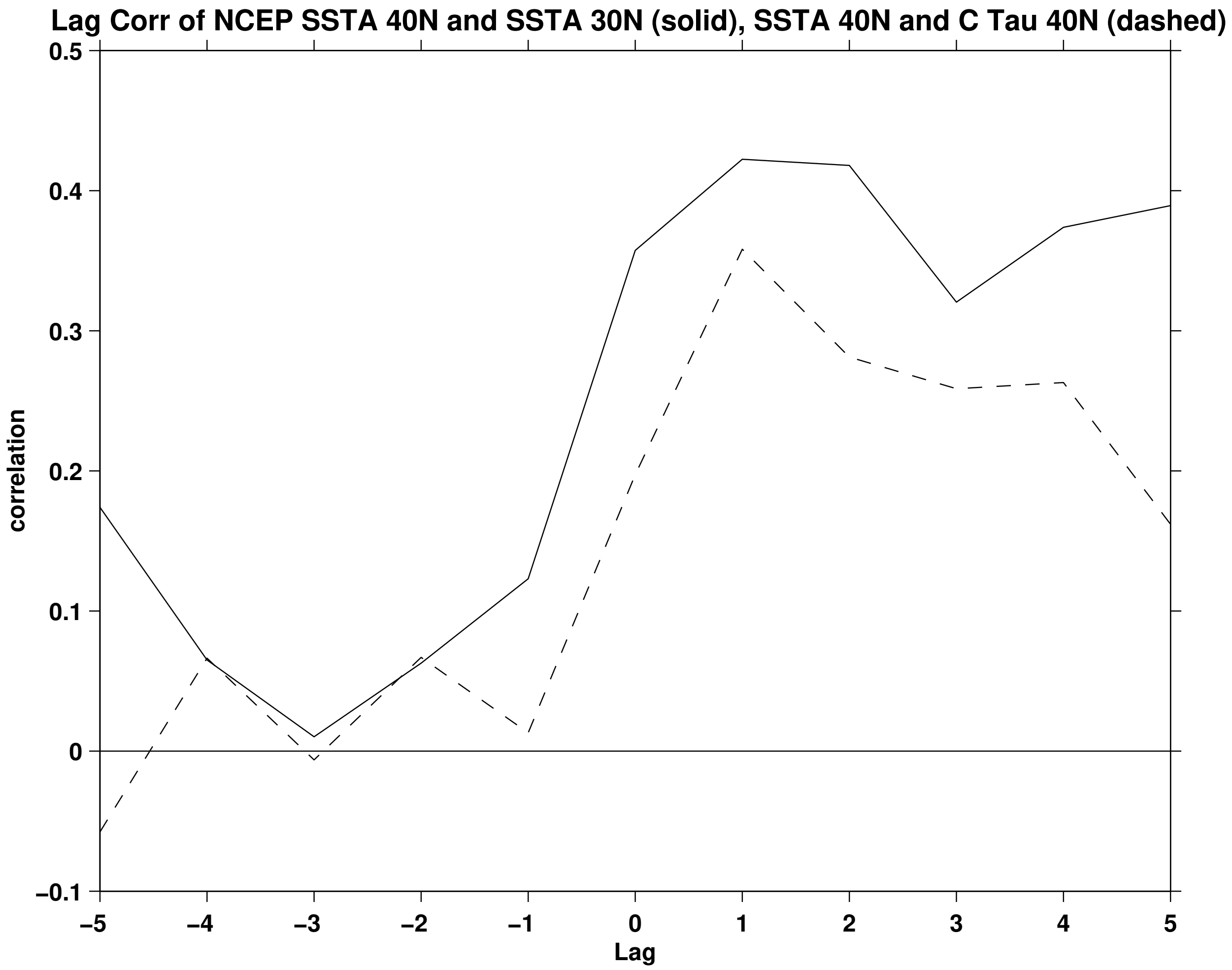

Lag Corr of NCEP SSTA 40N and SSTA 30N (solid), SSTA 40N and C Tau 40N (dashed)
correlation
0.5
0.4
0.3
0.2
0.1
0
–0.1
–5
–4
–3
–2
–1
0
1
2
3
4
5
Lag

Figure Captions

Figure 1. Longitude-time plots of the anomalies of zonal wind speed (ms^{-1}) and SST (K) along $30°N$ for (a) the observations and (b) the OGCM-AML model. Wind speed anomalies are in ms^{-1}. (Data from DaSilva et al. 1994).

Figure 2. Longitude-time plots of the anomalies of curl of wind stress ($\times 10^6$) and SST along $40°N$ for (a) the observations and (b) the OGCM-AML model (data from DaSilva et al. 1994).

Figure 3. The difference in (a) observed SSTs (K), surface winds and surface latent plus sensible heat flux (Wm^{-2}), for the January to March averages of the period 1982 to 1990 and the period 1967 to 1975. The data are from DaSilva et al. (1994). SSTs are colored and the surface heat fluxes are contored with the convention that positive means a flux from ocean to atmosphere. (b) The equivalent picture for the OGCM-AML model and (c) the difference in modeled ocean heat flux convergence (contours, Wm^{-2}), SSTs and observed winds.

Figure 4. The change in ocean heat flux convergence (Wm^{-2}) between the two epochs due to changes in (a) Ekman, (b) barotropic and (c) baroclinic currents working on the mean mixed layer temperatures.

Figure 5. The change in the depth averaged mixed layer currents between the two epochs.

Figure 6. The change in the (a) Ekman, (b) barotropic and (c) baroclinic currents in the mixed layer between the two epochs.

Figure 7. As for Fig. 2b but for the OML-AML model.

Figure 8. (a) The regression pattern of the curl of surface wind stress ($\times 10^6$) onto the subsequently observed SST in the region $34°N$ to $46°N$ and $139°E$ to $180°E$. The wind stress curl has been subjected to a three year running mean and the pattern shown corresponds to a lag of three years between the winds and the SST. (b) The time series of the SST index and the curl of wind stress, averaged along $40°N$ and subjected to a three year running mean and with the SST lagging the winds by three years. All data are from Da Silva et al. (1994)

Figure 9. Same as for Fig. 8 but using the SST index from the ocean GCM-AML model.

Figure 10. Lag correlation between SSTs in the central Pacific ($27.5°N - 32.5°N, 170°E - 140°W$) and the Kuroshio-Oyashio extension region ($37.5°N - 32.5°N, 150°E - 180°E$) (sold line) and between the SSTs in the Kuroshio-Oyashio extension region and the wind stress curl over the north central Pacific ($37.5°N - 42.5°N, 170°E - 140°W$) (dashed line). SSTs in the central Pacific precede those east of Japan and SSTs east of Japan lag wind stress curl over the North Pacific.