Decadal upper ocean temperature variability in the tropical Pacific

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Abstract. Decadal variability in upper ocean temperature in the Pacific is studied using observations and results from model experiments. Especially propagation of upper ocean thermal anomalies from the midlatitudes to the tropics is studied as a possible source for decadal equatorial thermocline variability. In the observations propagation along the subtropical gyre of the North Pacific is clear. However, no propagation into the equatorial region is found. Model experiments with an ocean model forced with observed monthly wind and wind stress anomalies are performed to study the apparent propagation. Distinct propagation of thermal anomalies in the subtropics is found in the model, although the amplitude of the anomalies is small. The anomalies clearly propagate into the tropics but they do not reach the equatorial region. The small response at the equator to extratropical variability consists of a change in the mean depth of the thermocline. It appears that most variability in the subtropics and tropics is generated by local wind stress anomalies. The results are discussed using results from a linear shallow water model in which similar features are found.

1. Introduction

The variability in the upper tropical Pacific has predominately an interannual time scale and is associated with El Niño Southern Oscillation (ENSO). This phenomenon has been extensively studied and has been explained as a coupled ocean-atmosphere mechanism [*Neelin et al.*, 1998]. The tropical Pacific also exhibits substantial decadal to interdecadal variability [*Zhang et al.*, 1998]. The spatial pattern of this low-frequency variability is similar to the pattern of interannual variability. Along the equator a large sea surface temperature (SST) anomaly stretches from the eastern boundary to the west. At the eastern side of the basin the anomaly extends from the equator towards the midlatitudes. In the western North Pacific an anomaly of the opposite sign is found. The main difference with the classical interannual ENSO pattern is that it extends further away from the equator and has a larger amplitude in the midlatitudes. Around 1942-43 and 1976-77 major shifts have occurred associated with this pattern in the Pacific [*Trenberth and Hurrel*, 1994; *Zhang et al.*, 1998].

Different physical mechanisms have been proposed to explain the origin of low-frequency variability in the tropical Pacific. *Gu and Philander* [1997] suggested that oceanic advection of subducted anomalies from the extratropics to the tropics sets the time scale of a decadal oscillation. In their conceptual model anomalies are formed in the winter mixed layer in the midlatitudes. When the mixed layer restratifies the anomalies subduct and propagate along the westward return flow of the subtropical gyre. The anomalies may reach the tropics where they can affect the tropical thermocline and generate an unstable air-sea feedback. The atmospheric winds in the midlatitudes respond to the tropical SST variations. These wind anomalies generate anomalies in the oceanic mixed layer that subsequently subduct and the second half of the cycle starts. A critical link in this mechanism is the advection of the thermal anomalies in the ocean from the midlatitudes towards the tropics which sets the time scale of the oscillation.

The existence of water mass pathways between extratropics and tropics is undisputed. Tritium observations show a connection between the equatorial region and subtropics [*Fine et al.,* 1987]. Also, different model studies indicate watermass pathways from the extratropics to the tropics [*Rothstein et al.,* 1998]. Whether heat or haline anomalies propagate into the tropics is less apparent. The propagation of subducted thermal anomalies along the westward return flow of the subtropical gyre is clear from the data [*Deser et al.,* 1996; *Zhang et al.,* 1998], but upon arrival at the western boundary the fate of the anomalies is not clear anymore [*Tourre et al.,* 1999]. Recent model studies indicate that propagation by advection by the mean flow of thermal anomalies towards the equator hardly takes place [*Schneider et al.,* 1999; *Pierce et al.,* 2000].

An oceanic connection between extratropics and tropics may also be accomplished by propagating planetary waves as suggested by theoretical and modelling studies [*Cane and Sarachik,* 1977; *Lysne et al.,* 1997; *Liu et al.,* 1999]. Rossby waves can be excited by a variable wind stress forcing at the midlatitudes, propagate to the western boundary and transfer to the tropics by coastal Kelvin waves. This mechanism will result in a shorter time lag between variability in the extratropics and tropics compared to the lag resulting from the subduction and advection mechanism. Upper ocean temperature anomalies in the 1980s suggest such a small lag.

Other authors stress the importance of an atmospheric "bridge". Decadal variability in the extratropical winds could influence the trades and therefore affect the equatorial thermocline almost instantaneously [*Wang and Weisberg,* 1998; *Barnett et al.,* 1999]. *Pierce et al.* [2000] used a coupled ocean-atmosphere model and found that the impact of the winds dominate, and speculate that changes in the midlatitudes may be important. Finally, low-frequency variability in the tropical Pacific can be locally coupled phenomenon as indicated by some coupled models [*Cane et al.,* 1995].

The different mechanisms that are proposed for low-frequency variability in the tropics indicate that there is a lot of uncertainty regarding the origin of that variability. Here we focus on the possible role of the ocean by studying propagation of thermal anomalies from the midlatitudes towards the equator. First, observed heat content variations will be used to see whether thermal anomalies from the midlatitudes propagate into the tropics. These observations include both surface and subsurface thermal variations. Then we will use a numerical ocean model coupled to an atmospheric mixed layer model forced with observed winds to investigate the mechanisms of the observed low-frequency variability. By performing different sensitivity experiments we deduce by what mechanism thermal anomalies are formed in the extratropics and whether they can reach the tropics. Unlike studies that have been performed with coupled ocean-atmosphere models, we can compare our model results directly with observations. On the other hand, we will not be able to infere whether the simulated variability is part of a coupled ocean-atmosphere phenomenon.

This paper is set up as follows. In the following section we analyze upper ocean temperature in the 1970s when propagation from the midlatitudes to the tropics is apparent (section 2). Then, in section 3 we present the model and motivate the sensitivity experiments. The results of the different experiments are shown and compared with observations in section 4, the results are discussed in section 5 using an simple linear shallow water model and we conclude in section 6.

2. Observed upper ocean variability

Thermal anomalies that are created at the sea surface by anomalous heat fluxes in the midlatitudes can subduct and propagate southwestward along the gyre circulation while shielded from the sea surface [*Deser et al.,* 1996]. Thus, SST and thermocline variations can be decoupled in the westward return flow of the subtropical gyre. Therefore surface and subsurface data must be used to study propagation of thermal anomalies propagate from the tropics to the extratropics. Here we analyze upper ocean heat content (the integral of temperature with depth, here 0-400 m) derived from XBT data from 1955 to 1998 [*White,* 1995]. This data has been used by *Tourre et al.,* [1999] to analyze interdecadal variability in the North Pacific. Using the complex EOF technique they found an interdecadal mode of variability with a time scale of 20-25 years. The mode consists of a growing phase during which anomalies are formed in the western and central North Pacific (along 40^{o}N). After about 4 years the anomaly turns southwestward at 150^{o}W and propagates along the westward return flow of the subtropical gyre. When arriving at the western boundary around 15^{o}N the anomaly neither propagates to the north or to the south, but dissipates away at the western boundary. Similar propagation of subsurface temperature anomalies is presented by *Zhang and Liu* [1999]. *Zhang et al.* [1998] and *Zhang and Liu* [1999] argue that these anomalies propagate into the equatorial region and affect the equatorial thermocline.

The statistical analysis of *Tourre et al.* [1999] shows one full period of the interdecadal cycle. Here we took a different approach and present snapshots of the low-pass filtered anomalies themselves during half a period of the cycle rather than the reconstructed anomalies from the statistical analysis.

The clearest event starts in the late sixties. At that time an anomalously warm anomaly forms in the upper layers of the central North Pacific (Figure 1a). This anomaly is accompanied by a decrease in the westerly winds that created the warm anomaly due to a decrease in the outgoing surface heat fluxes. The anomaly strengthens in the early seventies and two maxima are found (Figure 1b). One maximum covers the Kuroshio region. The other maximum is found in the center of the basin at 40^{o}N, 150^{o}W. In the early seventies this anomaly turns southwestward and follows the gyre circulation, propagating to the west in 1973, 1974 and 1975 (Figure 1c,d). In the following years, the anomaly halts around 20^{o}N, 180^{o}W-140^{o}W and it slowly fades away (Figure 1e,f). Another positive anomaly starts to form in 1974-1975 at 150^{o}E, 15^{o}N near the western boundary (Figure 1d). This anomaly seems to be related to the anomaly propagating around the gyre, but the anomalies never connect. The statistical analysis in *Tourre et al.* [1999] lumps the 2 positive anomalies together, possibly due to the stronger low-pass filter that they used. *Zhang et al.* [1998] and *Zhang and Liu* [1999] interpret the apparent connection between the anomalies as a verification of the advective path of midlatitude anomalies to the equator. However, the snapshots indicate that the anomalies in the central subtropical Pacific and western tropical Pacific were formed independently. Propagation of the midlatitude anomalies is halted in the central subtropical Pacific where they fade away. Also, in the second, less clear part of the interdecadal cycle that started around 1985, the subduction and initial propagation is clear, but the advection towards the western boundary is hardly visible (not shown).

As shown here, the evidence for an oceanic bridge for thermal anomalies from the midlatitudes to the equatorial region from data is weak, but the mechanism cannot be ruled out owing to the sparsity of the data and the relatively short data record. Therefore, we have performed a number of experiments with a numerical ocean model forced with observed winds to clarify whether propagation towards the tropics can take place. Furthermore, we intend to find out what mechanisms are responsible for generating the observed anomalies in the (sub)tropical Pacific.

3. Experimental setup

3.1. Models

For most experiments a primitive equation ocean model is used (see [*Visbeck et al.,* 1998; *Rodgers et al.,* 1999; *Hazeleger et al.,* 2000]). The model has a resolution of 2.5^o by 2.5^o in the midlatitudes. In the tropics the resolution in the meridional direction increases towards 0.5^o at the equator. The model has 20 levels in the vertical. The domain spans from 62^oN to 62^oS and from 90^oE to 70^oW. The model includes the Indonesian Throughflow of which the barotropic transport is set at 10 Sv. Furthermore, the model includes a Kraus-Turner mixed layer scheme and a one and a half layer thermodynamic ice model. An isopycnal thickness mixing scheme is incorporated with a spatially varying eddy diffusivity [*Visbeck et al.,* 1997]. The diffusivity coefficient varies from 150 m^2s^{-1} in the interior to 1500 m^2s^{-1} in the jet regions. The vertical mixing is Richardson number dependent. The model uses a TVD scheme for the horizontal advection terms. For the time-differencing a second-order Lorenz N-cycle is used. At the southern boundary of the basin the temperature and salinity are restored towards observations [*Levitus et al.,* 1994].

The ocean model is coupled to an atmospheric mixed layer model [*Seager et al.,* 1995] to create the so called LOAM model. In the atmospheric mixed layer the virtual potential temperature and the humidity are computed from a balance between advection, surface fluxes, fluxes at the top of the mixed layer, and radiative fluxes. The surface heat fluxes are determined by familiar bulk transfer formulae. The fluxes at the top of the atmospheric mixed layer are parameterized according to *Seager et al.,* [1995]. Cooling by midlatitude storms is parameterized according to *Hazeleger et al.,* [2000]. Only the wind, wind speed, cloud cover, and short wave radiation are prescribed. The advantage of this model is that the surface latent, sensible and long wave fluxes are internally generated. This implies that SST signals are not already captured in the heat fluxes, which is the case in hindcasts with prescribed heat fluxes or prescribed air temperatures. The salinity at the surface is restored to climatology [*Levitus et al.,* 1994]. Furthermore fractional cloudiness and the solar radiation at the sea surface is prescribed (ISCCP, [*Bishop and Rossow,* 1991]). At the lateral boundaries the potential temperature and the specific humidity at the surface are prescribed [*Da Silva et al.,* 1994].

The LOAM model has been spun up for 40 years with climatological winds, wind speed and wind stresses [*Da Silva et al.,* 1994]. The mean state of the model solution is realistic. For details on the model and a comparison of model climatology with observations we refer to *Hazeleger et al.,* [2000].

To identify some basic mechanisms of exchange of mass between the midlatitudes and the tropics we performed an experiment with a linear shallow water model The non-dimensional equations that are solved are:

$$\frac{\partial u}{\partial t} - fv + \frac{\partial h}{\partial x} = \tau_x \tag{1}$$

$$\frac{\partial v}{\partial t} + fu + \frac{\partial h}{\partial y} = \tau_y \tag{2}$$

$$\frac{\partial h}{\partial t} + \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} = 0 \tag{3}$$

Where u and v are the zonal and meridional velocity perturbations, h is the layer depth perturbation, and f the coriolis parameter. τ_x and τ_y are the zonal and meridional wind stress anomalies. The equations have been nondimensionalized with length scale $L = (c/\beta)^{1/2}$and time scale $T = (c\beta)^{-1/2}$, where $c = (gH)^{1/2}$ is the wave speed, H the equivalent depth, and $\beta = df/dy$.

The shallow water system is solved for two vertical standing modes, with speeds c_1 = 2.98 m/s and c_2 = 1.84 m/s. These wave speeds are calculated from the observed vertical density profile along the date line [*Levitus et al.,* 1994] and are assumed to be uniform over the entire domain. The domain of the model captures the Pacific Ocean from 60^oN to 60^oS and has realistic coast lines. The linear model calculates monthly thermocline depth anomalies at a resolution of 2^o in the zonal direction and 0.5^o in the meridional direction.The model is solved using the numerical scheme that is described by *Isreali et al.,* [2000]. The results of this model are compared with those of the LOAM model.

3.2. Experiments

After the spin up, the LOAM model has been forced with observed monthly vector wind and windspeed anomalies from COADS (1945-1993, [*Da Silva et al.,* 1994]) on top of the climatological monthly mean winds. This is the control experiment (see Table 1).

In the first sensitivity experiment the winds were held at their climatological monthly mean values in the tropics and subtropics. In a strip between 30^{o}N and 30^{o}S (tapered gradually between 28^o and 32^o, see Figure 2) no anomalies in the surface forcing are applied (experiment *MID*). This setup ensures that all low-frequency variability in the tropics is of midlatitudinal origin and is generated by oceanic processes alone. With this experiment we can test whether the oceanic "bridge" can be accomplished by advection of anomalies from the midlatitudes towards the tropics.

In a second experiment we applied observed anomalies in the atmospheric forcing in a tropical strip between 16^{o}N and 16^{o}S (gradually tapered between 14^o and 18^o). In this experiment all extratropical influences on the tropics by oceanic processes are excluded (experiment *TRP*).

In a third sensitivity experiment only anomalies in the wind stress were used (experiment *WST*). Climatological monthly mean wind speed and vector wind were applied. The surface heat fluxes are determined by a bulk transfer relation. So, anomalous surface fluxes can arise from anomalous wind speeds and anomalous air temperatures or humidity. By keeping the wind speeds and vector winds constant the air-sea temperature and humidity differences remain the only free variables in the bulk transfer relation. In that case, anomalous heat fluxes can only damp SST anomalies. The main way for the atmospheric forcing to create SST anomalies is by Ekman divergence and convergence. With this experiment we investigate by which surface forcing mechanism upper ocean thermal anomalies are formed.

In the last experiment we used the linear model. The model was forced with observed wind stress anomalies (experiment *LIN*). Just as in experiment *MID*, no anomalies were applied in a strip between 30^{o}N and 30^{o}S (tapered to 25^o).

Some of the model experiments are similar to those performed by *Schneider et al.* [1999], but here we use a different model configuration. The differences can potentially have a large effect. Schneider et al. used the HOPE model, which is a z-coordinate ocean-only model with a horizontal eddy mixing scheme and a closed

Indonesian Throughflow. Their model was forced with observed surface fluxes and a relaxation to observed SST and sea surface salinity. The LOAM model incorporates an isopycnal thickness mixing scheme with a varying eddy diffusivity for parameterizing the subgridscale eddies. This scheme leads to less spurious mixing than the horizontal mixing scheme which may lead to more long-lived anomalies in the model. Furthermore we opened the Indonesian Throughflow that regulates the distribution at the equator of water originating from the South Pacific and North Pacific. This is essential for simulating a proper distribution of temperature and salt in the equatorial thermocline [*Rodgers et al.,* 1999]. Tracer release experiments in the model revealed that twice as much water is transported from the South Pacific to the equator than from the North Pacific. This is in accordance with estimates from observations [*Johnson and McPhaden,* 1999]. Finally, our surface heat fluxes and outgoing long-wave radiation are free to evolve. This means that the simulated SST is not overly constrained, which would be the case if air temperature and humidity were specified.

4. Simulated thermocline variability

The wind speed from the COADS data has an upward trend in the North Pacific that seems not realistic (see *Miller et al.* [1994]). The trend caused an unrealistic, almost linear downward trend in the SST in the North Pacific. Therefore, we removed the linear trend from the model results. The trend is possibly caused by the switch from Beaufort estimates to anemometers measurements on ships (e.g. *Cardone et al.,* [1990]). By removing the linear trend we might get unrealistic results in the late 1980s and 1990s as the switch was largely completed by the early 1980s.

In Figure 3a we show the observed and modelled SST anomaly in the NINO3 region (region A in Figure 2). Big El Niño events such as those during 1972/73 and 1983/84 are well simulated. In Figure 3b,c we show the observed and simulated upper ocean thermal anomalies (averaged temperature anomaly over the upper 440 m) in the off-equatorial tropics and in the midlatitudes (region B and C in Figure 2). Here, we focus on the decadal time scales by low-pass filtering the data. In the off-equatorial tropics, the peak values around 1976 are well simulated (Figure 3b). These peaks correspond to the event discussed in section 2. The peak at 1985 is too large in the model. Also in the beginning of the run the model results differ from the observations. This discrepancy may be owing to the sparsity of the observations before 1970 [*Tourre et al.,* 1999] and to the detrending of the data. Most decadal swings are well simulated. This is true for the variations in the midlatitudes as well (Figure 3c). Here as well, most discrepancies are found at the beginning of the simulation and at the very end. In general the model captures the tropical interannual variability and extratropical decadal variations that are observed in nature. This makes the model suitable for studying the effect of decadal extratropical variability on the tropical thermocline.

4.1. Equatorial upper ocean variability

In Figure 4 we show the low-frequency upper ocean temperature anomalies along the equator. In the control experiment the signature of ENSO-like decadal variability is clear (Figure 4a). The variability consists of decadal swings in the tilt of the thermocline. For instance, from 1950 to 1960, the eastern equatorial Pacific was anomalously warm, while the western equatorial Pacific was cold. In the early seventies we find the opposite pattern. The standard deviation of these low-frequency upper ocean temperature anomalies in the warm pool of the equatorial Pacific is 0.16 K (region D in Figure 2; Table 2). The simulated amplitude of the low-frequency variability in the warm pool is in excellent agreement with observations. When no filtering is applied the modeled variability is larger than the observed variability. This is probably due to the underrated variability in the first 15 years in the observations owing to the sparsity of the data. When only data after 1970 are used a higher value is found in the observations (Table 2).

Experiment *MID* was designed to investigate whether low-frequency tropical variability can be of extratropical oceanic origin. Because the surface forcing in the tropics does not vary (except for a constant seasonal cycle), any low-frequency variability in the tropics must be of extratropical origin and transfered to the tropics by the ocean. In experiment *MID* low-frequency equatorial thermocline variability is clearly simulated (Figure 4b). Unlike the decadal variations in the observations, the decadal swings consist of deepening and shallowing of the thermocline rather than changes in the tilt of the thermocline. This is indicated by the single-signed upper ocean temperature anomalies along the equator. This distinct response of the equatorial thermocline to extratropical variability will be discussed further in section 5.

Although an impact of the midlatitudes on the equtorial thermocline is found, the amplitude is very small. Only a sixth of the observed decadal equatorial variability is simulated, that is, the standard deviation of upper ocean temperature in the warm pool is only 0.03 K (Table 2). Interestingly, this variability occurs almost entirely at low frequencies. The variations along the equator in experiment *MID* are quite unlikely related to the observed variations and the variations in the control run.

A small impact of the midlatitudes on the equatorial region was also found by *Lysne et al.* [1997] and *Schneider et al.* [1999]. Whether the coupled ENSO phenomenon is affected by the small low-frequency variations could not be simulated with the present model.

4.2. Origin of the thermal anomalies in the tropics

In the previous section we showed that the impact of midlatitude oceanic anomalies on the equatorial thermocline is small, but not negligible. As this may still be important we focus on the origin of the anomalies in this section. Here we study whether propagation of anomalies from the midlatitudes to the western boundary in the tropics induces this equatorial low-frequency variability. As shown in section 2 this is not evident from observed upper ocean temperature anomalies.

Figure 5a shows the upper ocean temperature anomalies in experiment *control* along the path that is depicted in Figure 2. This pathway mimics the advective pathway along which thermal anomalies propagate from the midlatitudes to the tropics. The pathway was constructed using the apparent propagation in the observations. No clear propagation takes place in experiment *control*. Furthermore, the equatorial low-frequency variability is uncorrelated with the off-equatorial variability.

In experiment *MID* anomalies of upper ocean temperature are found well south of 30^{o}N where no anomalous surface forcing is applied (Figure 5b). These anomalies must be generated by oceanic heat transport. However, the amplitude of the anomalies is small (note the different contour interval compared to Figure 5a). Propagation is clearest between 25^{o}N and 15^{o}N. South of 15^{o}N no propagation is found. This implies that the anomalies stay north of the North Equatorial Current where the upward doming isopycnals act as a potential vorticity barrier for the anomalies.

Propagation into the equatorial region of thermal anomalies from the South Pacific seems more likely because the transport in the upper ocean from the subtropics to the tropics is larger from the South Pacific than from the North Pacific. The observations are too sparse to deduce whether propagation of anomalies takes place. In the model we do not find propagation of anomalies from the South Pacific (not shown).

This experiment shows that thermal anomalies generated in the midlatitudes do not connect with the equatorial Pacific by an advective path along the subtropical gyre. However, the sign of the anomalies at the equator is almost identical to that in the western off-equatorial tropics. The absence of a lag of several years between the anomalies at 15^{o}N and the anomalies at the equator suggests that anomalies propagate fast from the western tropics into the equatorial region along the western boundary. We will come back to this potential mechanism of subtropics-tropics interaction in section 5.

Along 15^{o}N and 20^{o}N two and a half cycles of an interdecadal cycle is present

(Figure 5b). This regular cycle seems to correspond to the cycle found by *Tourre et al.* [1999]. They used observations of upper ocean temperature after 1970 and show only 1 cycle. Our model results suggest that indeed, a regular interdecadal cycle is present in the upper ocean from 1945 on. Furthermore, this experiment shows that oceanic heat transport is responsible for generating this interdecadal mode of variability in the off-equatorial tropics.

The anomalies in the tropics in experiment *MID* are small in comparison to the anomalies in experiment *control* (Figure 5a,b). Since advection is not a dominant mechanism for generating these thermal anomalies, the question arises what alternative mechanism generates the anomalies in this region. The atmospheric forcing is the main candidate. The atmospheric forcing creates heat anomalies in the ocean by either the surface heat fluxes or by Ekman divergence and convergence due to varying wind stresses. Also changes in turbulent mixing in the mixed layer due to varying surface fluxes can create anomalies.

A run with only varying wind stress (experiment *WST*) was performed to distinguish the effect of the wind-driven variability from the variability induced by surface forcing. In LOAM the surface heat fluxes are determined by a bulk transfer relation. By keeping the wind speeds and the winds that advect air temperature and humidity to their climatological value we largely remove the heat flux forcing. However, changes in the Ekman transport divergence can introduce significant upper ocean heat content anomalies. In Figure 5c we show the upper ocean temperature anomalies along the advective path for this experiment. When compared to Figure 5a it is evident that most tropical variability is captured in this experiment. Since advection from midlatitudes is small, this implies that most variability is induced by local anomalous divergence/convergence in the upper ocean due to the varying wind stress or by propagating Rossby waves from tropical origin. The difference between the control experiment and experiment *WST* shows the effect of the surface heat fluxes.

Indeed, the surface heat fluxes make a very small contribution and act to damp the anomalies (Figure 5d).

To confirm that anomalies in the tropics are almost entirely generated by local tropical processes we performed experiment *TRP*. Only very small differences could be found between experiment *TRP* and the control experiment (not shown).

4.3. An example of a locally forced thermal anomaly

In section 2 we showed an apparent advective path of thermal anomalies from the midlatitudes towards the tropics along the southern rim of the subtropical gyre. However, the connection between the anomalies in the central subtropical Pacific and the western tropical Pacific was not clear. The key question is whether the anomalies in the western off-equatorial tropics of the North Pacific are formed by advection of heat anomalies or by another mechanism. Therefore, we examine the mechanisms that generate upper ocean temperature variability in the western tropics in more detail.

In Figure 6 we show time series of upper ocean temperature anomalies in the western tropics derived from the different model experiments (see also Figure 3c where observed anomalies from the same area [region C in Figure 2] are shown). The observed time series is dominated by a number of peaks. The largest has its maximum around 1978. Figure 1 shows the spatial distribution of the anomalies at that time. It is evident that the very same peak is simulated in the model with tropical forcing only (experiment *TRP*, Figure 6). Also the peaks in 1985 and 1990 are well simulated in this experiment. Since the forcing north of 16^{o}N is constant in this experiment, the anomalies must be locally generated. We conclude that this anomaly cannot be from extratropical oceanic origin such was suggested by e.g. *Zhang et al.* [1998]. This is also confirmed by experiment *MID* which shows hardly low-frequency variability in the western tropics (Figure 6). This example confirms the suggestion made in section 2 and 4.2. Based on the observations and the model experiments, the thermal anomalies in the western tropical North Pacific are locally generated. Also, in experiment *WST* the observed anomalies are well simulated (not shown, but already evident from Figure 5c). This indicates that local Ekman divergence and convergence is responsible for creating the anomalies. Although advection may have preceded the strong anomaly and an unstable air/sea feedback could have enhanced the anomalies, the above results show that is not a very likely mechanism.

5. Discussion

In the previous sections we have shown that the impact of variability in the midlatitudes on the tropical thermocline by oceanic heat transport is small, but not entirely negligible. Experiment *MID* has shown that advection by the mean flow was not responsible for the small response in the equatorial region. The absence of a time lag between variability in the western off-equatorial tropics and the equatorial region suggests another mechanism. Anomalies could be generated in the midlatitudes and propagate as coastal Kelvin waves anticlockwise to the equator such as envisaged by e.g. *Lysne et al.* [1997] and *Liu et al.* [1999]. If this mechanism generates low-frequency variability in the equatorial thermocline, the response found in experiment *MID* should be consistent with linear theory. Therefore, to test the aforementioned hypothesis we forced a linear shallow water model (see section 3.1) with observed wind stress anomalies north of 30^{o}N and south of 30^{o}S (experiment *LIN*), equivalent to experiment *MID*.

The linear model calculates thermocline depth disturbances with respect to a mean thermocline depth of 175 m. The depth anomalies can be translated into upper ocean temperature anomalies using the mean vertical temperature distribution. The standard deviation of the upper ocean thermal anomalies in the tropical warm pool in the linear model is 0.02 K (Table 2). This is close to the variability in experiment *MID* taking into account the idealizations in the linear model. Most variability takes place at low frequencies, which is in accordance with experiment *MID*.

In Figure 7a we show the thermal anomalies along the equator of experiment *LIN*. Just as in experiment *MID*, there are slow decadal variations along the equator. Also, there is no zonal variation, that is, the anomalies correspond to changes in the mean depth of the thermocline. Qualitatively, this result is consistent with the result found in experiment *MID* using the LOAM model (see Figure 5b).

So, the amplitude and the spatial structure compare favorably, but there are large differences between results of both experiments in the time domain as shown in Figure 8. Here the upper ocean temperature in the eastern and western equatorial Pacific for the different experiments are plotted. The time series in the east and in the west are in-phase in experiment *LIN* and experiment *MID*, but the time of occurences of maxima and minima don't correspond. In experiment *control* the decadal variation in the tilt of the thermocline associated with the decadal ENSO signal is visible as indicated by the out-of-phase behavior.

Differences between the results of the linear model and the LOAM model can have been caused by, for instance, the assumption of a uniform vertical density gradient over the entire Pacific in the linear model and the associated lack of mean flow. Anomalies generated by wind stress anomalies in the North Pacific propagate westward as Rossby waves. Upon arrival at the boundary they propagate fast into the tropics as Kelvin waves. This can be seen in Figure 9 where standard deviation of the upper ocean temperature anomalies are shown for both experiments. Large anomalies are found north of 30^oN in experiment *LIN*, but south of 30^oN anomalies are only found along the western boundary (tapering of the forcing towards zero causes anomalies to be found south of 30^oN). The anomalies along the western boundary are the signature of the Kelvin waves. In the LOAM model anomalies propagate along a path along the subtropical gyre well south of 30^oN before arriving at the western boundary. Then they propagate fast into the equatorial Pacific as well, although this propagation is less clear than in the linear model. The lack of propagation along this path due to the absence of a mean flow in the linear model is also clear in Figure 7b and can explain differences in the time domain.

A common feature of the experiment *MID* and experiment *LIN* is the zonally uniform response along the equator. The zonally uniform response occurs only at low frequencies. At higher frequencies, variations in the tilt of the thermocline are found. This can be seen in Figure 10 where spectra are presented of the difference between the phase angle of time series of upper ocean temperature in the NINO3 region and in the warm pool region (see Figure 2, region A and D). A value near π means that the time series in the eastern and western Pacific are out-of-phase. When the value is zero the timeseries are in-phase. For experiment *control* out-of-phase relationships are found at nearly all frequencies. This reflects the ENSO mode at high as well as at low frequencies. This is in accordance with observations that show interannual ENSO signals and ENSO-like (inter)decadal signals in the equatorial Pacific [*Zhang et al.,* 1997]. The mode is associated with variability in the wind stress in the equatorial Pacific. The phase angle differences in experiment *MID* and *LIN* show a different behavior. At high frequencies the time series are out-of-phase. The phase angle drops to zero after a few years. For experiment *LIN* this is most clear. At a period of 6 years the angle has become zero, which corresponds to a zonally uniform response. For experiment *MID* the picture is somewhat less clear, but the phase angle clearly drops for decreasing frequencies. At a period of about 8 years the angle becomes zero. Although there are quantitative differences between the results of the LOAM model and the linear model for reasons that have been stated above, the similarities in the frequency domain (Figure 10b and 10c) leads to the conclusion that the same physical mechanisms of equatorial thermocline variability are acting in both models. That is, at high frequencies free equatorial modes are found as reflected by out-of-phase relationships between the eastern and western Pacific. These are equatorial Kelvin and Rossby waves that build up to an equatorial basin mode at low frequencies [*Cane and Moore,* 1981]. The basin mode consists of a zonally uniform deepening and shallowing of the thermocline (see also Figure 4b and 7a). It is worth noting that the analytical solution of *Cane and Moore* [1981] (their equation 12) also has a break from zero phase at these low frequencies.

In experiment *control* there are free and forced modes of equatorial variability at high and low frequencies. The large amplitude differences between the upper ocean temperature variations in experiment *control* and experiment *MID* implies that the forced response dominates over the effect of the free basin modes. Unlike the low-frequency basin modes, the response to wind stress anomalies consists of a changes in the tilt of thermocline. This is in accordance with the findings in the previous sections.

6. Conclusions

Observations of heat content suggest propagation of thermal anomalies along the westward return flow of the subtropical gyre. Previous studies have suggested that these anomalies connect with the tropical thermocline and affect the equatorial thermocline [*Gu and Philander,* 1997]. The propagating anomalies would be a part a cycle of a decadal extratropical-tropical climate oscillation. In the current data record 2 events are present when propagation in the off-equatorial tropics takes place, but we have shown that the connection to the equatorial region is not convincingly clear. Anomalies formed by subduction in the eastern North Pacific seem to fade away in the subtropical and tropical Pacific rather than propagating into the equatorial region.

Model experiments have been performed to investigate the apparent connection between the extratropical Pacific and the tropical Pacific. We found that anomalies propagate from the midlatitudes into the tropics. Advection of anomalies takes place from the east towards the west, following the westward return flow of the subtropical gyre. Interdecadal variability induced by midlatitude variability could be observed in the off-equatorial tropics. The amplitude, however, is small. The anomalies hardly propagate into the equatorial region from either hemisphere. Based on the model experiments, a sixth of the variability in the equatorial thermocline can be attributed to oceanic variability in the midlatitudes The remaining low-frequency variability in the tropics is mainly of local origin, forced by wind stress variations.

Results from an experiment with a linear shallow water model showed similar features as the experiments with the primitive equation model. The most distinct feature is that the response to extratropical wind stress variability consists of a change in the mean depth of the equatorial thermocline. This is in contrast to the response to local tropical wind stress anomalies. In that case the tilt of the thermocline would be affected. At high frequencies however, when no anomalies in the forcing were applied, changes in the tilt were found. These correspond to the free equatorial basin modes. These modes build up to a basin mode at low frequencies. The low-frequency mode consists of a zonally uniform deepening and shallowing of the thermocline. The linear model showed that extratropical oceanic thermocline disturbances can reach the tropics by propagation along the western boundary as coastal Kelvin waves. This explains the small time lag between variability in the western off-equatorial tropics and the equatorial region.

Our results confirm findings from previous studies. *Schneider et al.* [1999] used an ocean-only model to show that the impact of oceanic advection of thermal anomalies from the midlatitudes to the equator is small. In a coupled ocean-atmosphere model *Pierce et el.* [2000] could not find evidence for an advective path as well. Using a different model with a different configuration we come to the conclusion that at the impact of the midlatitudes on the tropics by oceanic processes is small. The analysis of observed upper ocean heat content confirms our findings. In addition we have shown that the observed low-frequency tropical variability is almost entirely forced by local wind stress variations. This confirms the results of the coupled model study by *Pierce et al.* [2000]. Finally we have shown that the response of the equatorial thermocline to midlatitudinal oceanic variations is consistent with linear wave dynamics, but it is modified by the mean flow.

Although the impact of extratropical wind variations on the equatorial thermocline is small, it is not entirely negligible. We plan to study how the changes in the mean depth of the thermocline affects ENSO in a coupled ocean-atmosphere model in the future.

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Figure 1. Snapshots of low-pass filtered upper ocean temperature anomalies (averaged temperature anomaly over upper 400m) in 1969, 1971, 1973 1975, 1977, and 1979 (a,b,c,d,e,f). The anomalies have been low-pass filtered with a Parzen window with a width of 4 years.

Figure 2. Schematic drawing showing the positions of the mask used in experiment *MID* (dashed), *TRP* (dotted) and *LIN* (dashed). The continuous line shows the path along which the time-latitude diagrams in Figure 4 is made. The boxes are the areas over which is averaged (see text).

Figure 3. Observed (COADS, fat) and modelled (control exp, thin) time series of (a) SST anomalies in the NINO3 region (150W-90W; 5N-5S, region A in Figure 2), (b) low-pass filtered (4yr running mean with Parzen window) upper ocean temperature anomalies in the off-equatorial tropics (130E-150E; 18N-11N, region C in Figure 2), and (c) low-pass filtered (4yr running mean with Parzen window) upper ocean temperature anomalies in the midlatitudes (160E-170W; 45N-35N, region B in Figure 2).

Figure 4. Low-pass filtered (4yr running mean with Parzen window) upper ocean temperature anomalies along the equator in (a) experiment *control* (contour interval: 0.1 K) and (b) experiment *MID* (contour interval: 0.02K).

Figure 5. Low-pass filtered (4yr running mean with Parzen window) upper ocean temperature anomalies along the ventilation pathway depicted in Figure 2. (a) Experiment *control* (contour interval: 0.1 K), (b) experiment *MID* (contour interval: 0.05 K), (c) experiment *WST* (contour interval: 0.1 K) and (d) experiment *control* minus *WST* (contour interval: 0.05 K).

Figure 6. Low-pass filtered (4yr running mean with Parzen window) upper ocean temperature anomalies in the off-equatorial tropics (130E-150E; 18N-11N, region C in Figure 2, see also Figure 3c) in the observations (fat), experiment *MID* (thin) and *TRP* (dashed).

Figure 7. Low-pass filtered (4yr running mean with Parzen window) upper ocean temperature anomalies (K) in experiment *LIN*. (a) Along the equator (contour interval: 0.01 K), and (b) along the ventilation pathway (contour interval: 0.1 K).

Figure 8. Time series of low-pass filtered (2yr running mean with Parzen window) upper ocean temperature anomalies (K) in (a) experiment *control*, (b) experiment *MID* and (c) experiment *LIN*.

Figure 9. Standard deviation of upper ocean temperature (K) in (a) experiment *LIN* and (b) experiment *MID*.

Figure 10. Spectrum of phase angle differences between time series of upper ocean temperature in the eastern tropical Pacific (region A in Figure 2) and the western tropical Pacific (region D in Figure 2). (a) Experiment *control*, (b) experiment *MID* and (c) experiment *LIN*.

Table 1. Experiments. The first column states the name of the experiment. The second column states the model that has been used (LOAM: primitive equation ocean model coupled to atmospheric mixed layer, LOAM2: as LOAM, but only variable forcing in the windstress was applied. LIN: linear shallow water model forced by wind stress anomalies). The third column states the region where the anomalous forcing is applied. See text for further details.

Experiment	Model	Mask
control	LOAM	-
MID	LOAM	30^oS-30^oN
TRP	LOAM	$< 16^o$S and $>16^o$N
WST	LOAM2	-
LIN	LIN	30^oS-30^oN

Table 2. Standard deviation of upper ocean temperature (K) in the warm pool of the tropical Pacific (5^{o}S to 5^{o}N, 120^{o}E to 180^{o}E, region D see Figure 2). See text for further details.

data	low-pass filtered (> 4 year)	no filter
obs	0.16	0.31 (0.37 after 1970)
control	0.16	0.35
MID	0.03	0.03
TRP	0.17	0.36
LIN	0.02	0.02