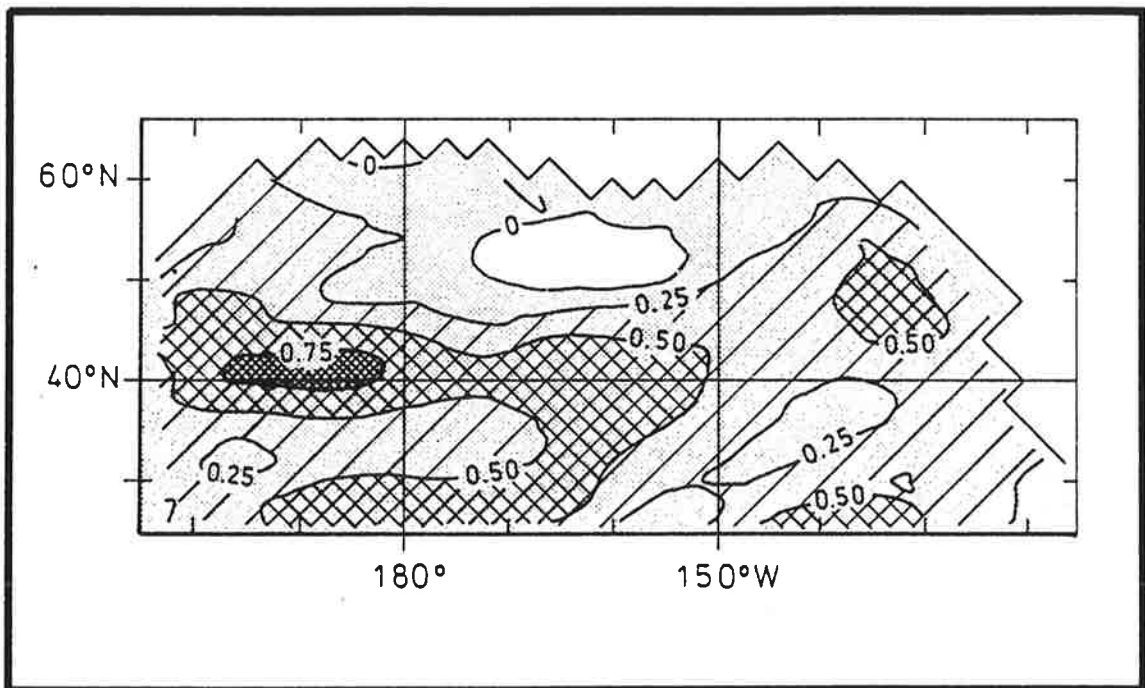




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MODELING NORTH PACIFIC SST ANOMALIES AS A RESPONSE TO ANOMALOUS ATMOSPHERIC FORCING

by

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MODELING NORTH PACIFIC SST ANOMALIES
AS A RESPONSE TO ANOMALOUS ATMOSPHERIC FORCING

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Abstract

Large-scale sea surface temperature anomalies (SSTA) in the North Pacific ocean often persist for several months during wintertime. There is observational evidence that these patterns are forced by anomalous atmospheric circulation. Since the latter is in part related to the tropical El Niño/Southern Oscillation (ENSO) phenomenon it is hypothesized that part of the North Pacific SSTAs may be interpreted as remote oceanic response to anomalous equatorial Pacific SSTAs.

Two experiments with a multi-level primitive equation model of the North Pacific have been conducted to study the influence of such anomalous atmospheric circulation on the SST. In both experiments anomalous wind stress as derived from the 1950-1979 COADS subset is specified as anomalous forcing. In Experiment 1 no anomalous heat flux is introduced whereas in experiment 2 anomalous heat fluxes are estimated from anomalous surface winds and a simple advective atmosphere.

In both experiments the GCM SSTA responses are able to reproduce the main features of the time series of observed SSTA, in particular in winter. In experiment 1, however, the magnitudes are systematically too low. The addition of anomalous heat fluxes in experiment 2 significantly improves the simulation. The ENSO signal is clearly present in both simulations.

1. INTRODUCTION

North Pacific SSTAs are characterized by large spatial patterns (Namias, 1969; Weare et al. 1976). The persistence of these patterns, which often last for more than 5 months, is considerably greater in winter than in summer (Namias et al., 1988). One reason for this seasonal modulation is the shallowing of the upper mixed layer from 100-200m in winter to 10-30m in summer (Bathen, 1972). The lower heat storage of the summer mixed layer causes the SST to be sensitive to high frequency processes like clouds and fronts. In winter such high frequency processes are considerably less effective.

The SST is influenced by horizontal and vertical advection, turbulent mixing and by heat exchange with the atmosphere. Oceanic processes alone seem to be too slow to establish the observed persistent large-scale SST anomalies: the advective oceanic velocities in the Central North Pacific are generally less than 20cm/s (i.e., 5° /month; White et al., 1985). It is therefore reasonable to hypothesize that the anomalous atmospheric forcing is mainly responsible to the evolution of anomalous SST. Several studies have been published supporting this notion:

1. According to statistical analysis do atmospheric circulation anomalies lead ocean anomalies in winter (Davis, 1976, 1978; Wallace and Jiang, 1987).
2. Drifting buoys measurements in the midlatitude North Pacific indicate highly coherent variations of wintertime SST and surface wind on time scales of 1 to 16 days (McNally et al., 1989).
3. The low frequency variability of the North Pacific atmospheric circulation is dominated by planetary scale standing oscillations with time scales of about one month (Wallace and Gutzler, 1981, Barnston and Livezey, 1987). The most prominent pattern is the so-called "Pacific - North American" pattern (PNA).

The heat fluxes seem to be the most important forcing for the development of the observed SST. Using a mixed layer model for the ocean Frankignoul and Reynolds (1983) found the oceanic response to anomalous heat fluxes (derived from observations) to be almost four times larger than the response to wind stress anomalies. Numerical experiments with a primitive equation model without anomalous heat fluxes resulted in SST anomalies three times weaker than observed (Haney et al., 1985).

Namias et al. (1988) found the most persistent pattern occurred in periods for

which SST was colder than normal in the central part of the North Pacific and warmer than normal in the eastern part. The results of EOF analysis and correlation studies of the observed SST obtained by various authors (e.g., Weare et al., 1976; Frankignoul and Reynolds, 1983; Wright et al., 1985) indicate, that the SST along the west coast of North America varies in phase with the central equatorial Pacific SST whereas the SST in the central North Pacific is 180° out of phase. The magnitude of the North Pacific pattern may conveniently be measured by an index, defined as the SST difference between the northeast Pacific and the north-central Pacific. A comparison of this index with the (tropical) Southern Oscillation Index (SOI) indicates that the midlatitude SST pattern is closely associated with the Southern Oscillation with a slight tendency for the tropical index to lead the extratropical index (Wright, 1979, 1983). Because of the short time lag of less than 2 months, it appears that the remote connection between the North Pacific SSTA and the tropical ENSO phenomenon can be established only by means of atmospheric teleconnections - as argued above.

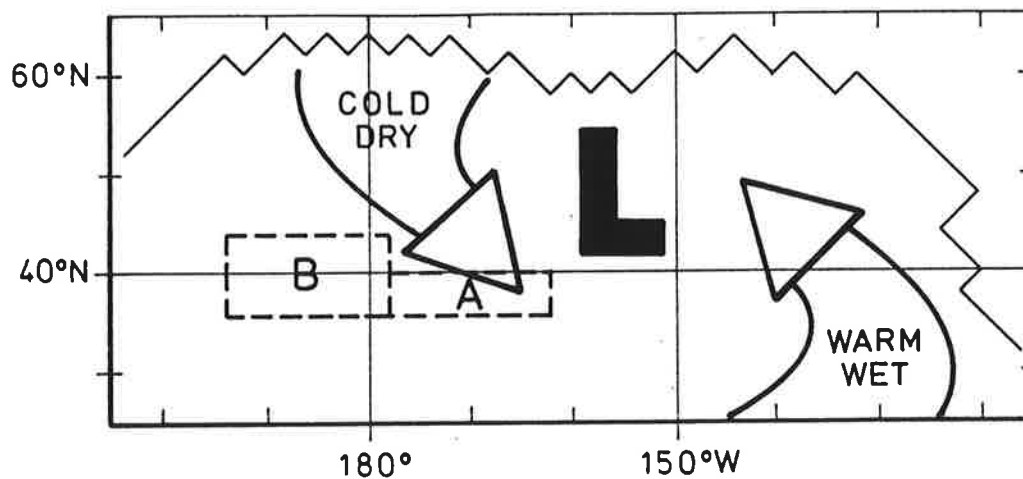
The existence of such teleconnections was first hypothesized by Bjerknes (1969, 1972). His hypothesis was later verified by various empirical studies and GCM experiments. In a composite analysis, van Loon and Rogers (1981) calculated the averaged difference of the monthly mean 700mb height at the extreme states of SO. In accordance with Wright's correlation pattern, their composite map (a schematic of which is given in Fig. 1) indicates cooling of the central North Pacific and warming of the eastern North Pacific. The anomalous flow advecting cold and (continental) dry air from the north to the central North Pacific cools the central North Pacific; while the warm and humid air blowing from the tropical Pacific warms the eastern North Pacific. The cyclonic shear in the central part of the ocean causes upwelling of cold water in that area. Along the American coast the anomalous northward flow is associated with coastal downwelling. Various AGCM experiments support the concept that a positive SSTA in the tropical eastern Pacific generates lowered temperature over most of the North Pacific and a warming over the western North America (e.g., Blackmon et al., 1983; Cubasch, 1985).

In the present study we report OGCM experiments on the influence of wind stress and of heat fluxes on the evolution of North Pacific SSTA. In Section 2 the ocean GCM is described, and its skill in reproducing the observed mean state is discussed. Two major experiments were conducted (Section 3). In both experiments the ocean was forced with anomalous wind conditions as observed

from 1950 to 1979 (COADS). In the first experiment, the only anomalous forcing is the wind stress whereas in the second experiment, anomalous thermal (atmospheric) conditions are imposed in addition. In Section 4, the model's SST response in these two experiments is compared to the observed SST anomalies. Finally, the remote El Niño signal in the model runs are examined (Section 5).

Figure 1:

Schematic of the anomalous wintertime flow during a warm SO event (after a 700 mb height composite given by van Loon and Rogers, 1981). Time series of observed and simulated SST anomalies averaged in area A and B are shown in Section 4.3.



2. THE OCEAN GCM

2.1 Model structure

Latif et al. (1985) used a similar version of the model for tropical simulations. The full (primitive) equations are simplified by the usual hydrostatic and Boussinesq approximations. The basic set of prognostic equations is given by the equations of motion, the conservation of temperature and salt and the kinematic boundary condition. The components of the basic system state vector are given by the horizontal velocity, sea level, salinity and temperature. The diagnostic set of equations consists of the hydrostatic equation, the equation of state of sea water and the equation of continuity.

The barotropic and the baroclinic components of the extratropical ocean dynamics have very different time scales. They are therefore subjected to different numerical methods. For the rapidly changing barotropic component, an implicit numerical scheme is applied. This strongly reduces the energy and phase speed of gravitational waves, but only slightly damps the Rossby waves. The fast coastal Kelvin waves therefore cannot be reproduced by this model. The slowly varying baroclinic component is discretized with an explicit time differencing scheme.

The analysis domain is the North Pacific from 28°N to 60°N. To avoid boundary problems, and to model realistic ocean mean currents, the model domain is expanded southward to 30°S. The horizontal resolution is uniformly 3°×3°. A realistic topography is used. There are 13 layers in the vertical with a higher resolution in the upper 200m. The horizontal turbulent eddy transport coefficients are set at the constant values: $A_h = 2 \cdot 10^5 \text{ m}^2/\text{s}$, $D_h = 5 \cdot 10^2 \text{ m}^2/\text{s}$, while the vertical mixing is dependent on the local Richardson number, essentially in occurrence with Pacanowski and Philander (1981).

2.2 Model climatology

The model was initialized with horizontally homogeneous salt and temperature distributions and was then spun up for 50 years. The forcing was the climatological wind stress from Hellerman and Rosenstein (1983) and surface fluxes of heat and salt specified by a Newtonian formulation using climatological distributions of surface salinity (yearly mean after Levitus,

1982) and SST (monthly mean after COADS: Woodruff et al., 1987).

In the last 10 years of the spin up experiment, the model was almost (cyclo-) stationary in its upper 200 m. The observed mean state in the upper ocean of the analysis domain, i.e. from 28°N to 60°N , is reproduced satisfactorily by the model. Fig. 2 shows two north-south sections of the temperature, one near the dateline and the other near the west coast of America. The agreement is generally good, but the gradients are too weak. The simulation of too high temperatures north of 40°N might be related to the "wall" at 60°N in the model which inhibits any exchange of heat with the polar ocean.

In the analysis domain the modeled sea level is similar to the observed dynamic topography (Wyrтки, 1975) (Fig. 3). The simulated difference of 1.40 m between the extremes in the western North Pacific is in good agreement with the observed 1.20 m difference. The meanders in the Kuroshio region are not resolved in the model. Outside the analysis domain (south to 20°N) the agreement between model and observations is not so good. On the Southern Hemisphere the model results are influenced by the boundary conditions. However, these discrepancies may be expected to have no impact on the quality of the simulation of SST anomalies by local atmospheric forcing in the analysis domain.

Figure 2:

Latitude/depth cross-sections of simulated and observed ocean temperature in January. Units: °C. Simulated data: at the end of the 50 year spin-up experiment. Observed data: after Levitus (1982).

- a. Model data along 178°E.
- b. Observed data along 178°E.
- c. Model data along 122°W.
- d. Observed data along 122°W.

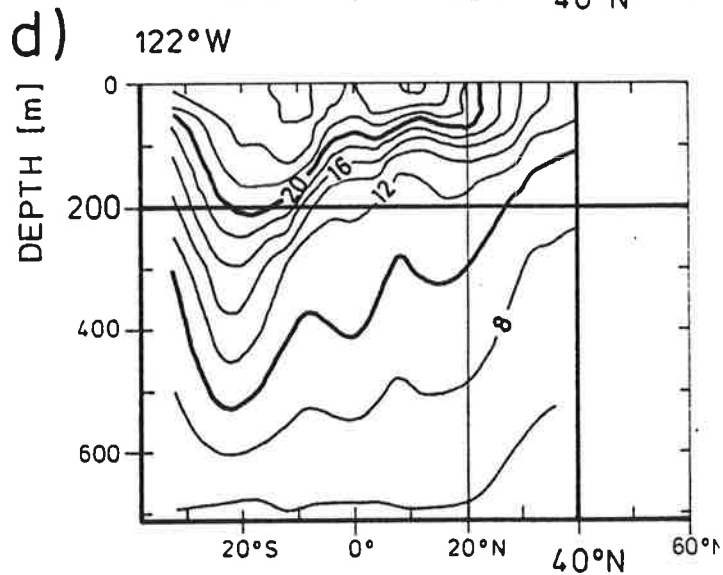
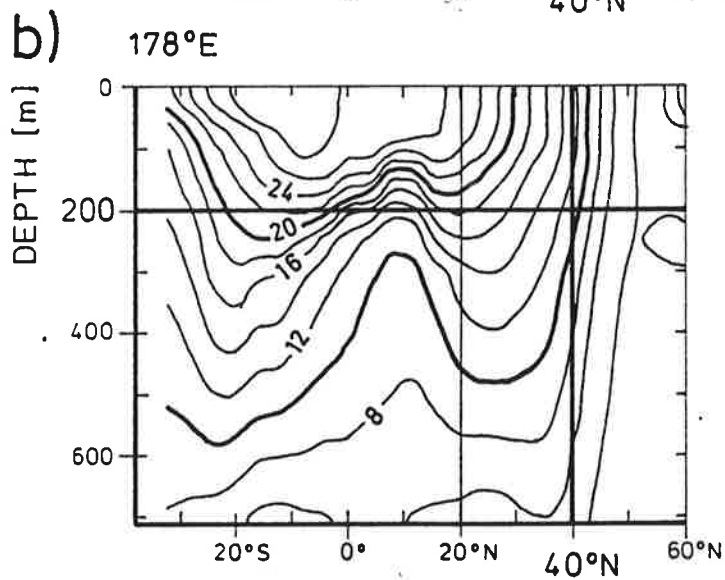
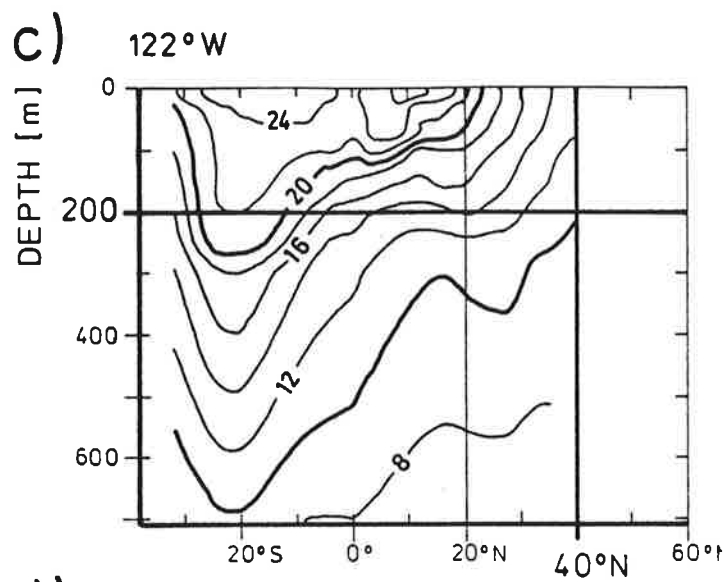
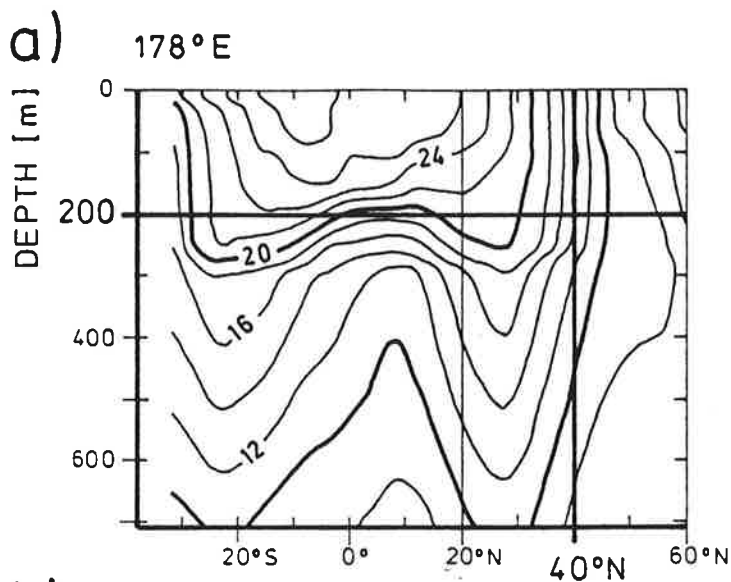
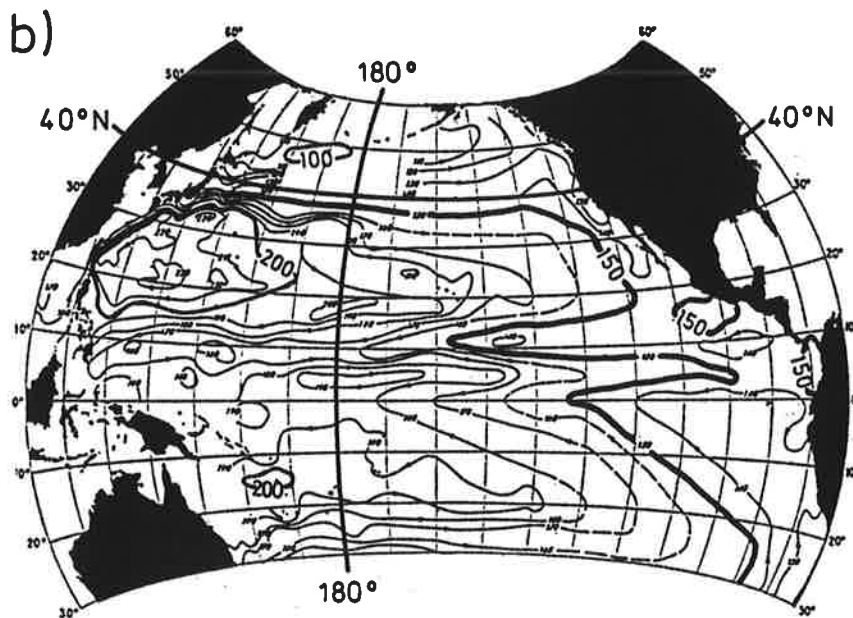
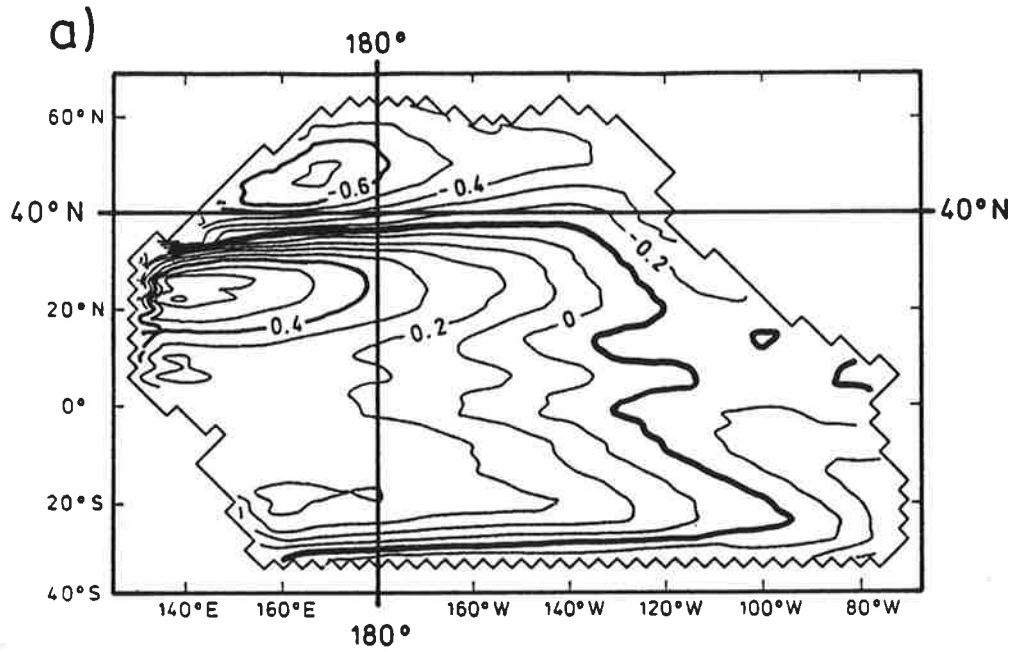


Figure 3:

Modeled sea level and observed dynamical topography. Note the agreement in the experimental domain, 28°N to 60°N . The difference between the extremes in the western North Pacific is 1.40m in the simulation and 1.20m in the observation.

- a. Modeled January sea level at the end of the spin-up experiment. Units: m.
- b. Dynamic topography (sea surface relative to 1000 db) in November / December after Wyrтки (1975). Units: 10^{-2}m .



3. THE EXPERIMENTS

To study the impact of anomalous atmospheric circulation on SST in the North Pacific, two experiments were carried out in which the ocean GCM was forced by anomalous atmospheric conditions. Both were initialized with the realistic state at the end of the spin-up experiment (Figs. 2 and 3). The experiments were designed as anomaly experiments, i.e., the forcing \mathcal{F} was represented as $\mathcal{F} = \mathcal{F}_c + \mathcal{F}_a$ with a climatological forcing \mathcal{F}_c and an anomalous forcing \mathcal{F}_a . The climatological wind stress \mathcal{F}_c^τ was specified according to Hellerman and Rosenstein (1983) and the climatological heat and fresh water fluxes, \mathcal{F}_c^H and \mathcal{F}_c^S , were derived from the last 10 years of the spin-up experiment. With this set-up the model maintains the equilibrium reached in the spin-up experiment if no anomalous forcing, \mathcal{F}_a , is introduced. In contrast to the spin-up experiment (Section 2), the model is no longer fixed to the observed values.

For **experiment I** no anomalous heat and salt fluxes were introduced, $\mathcal{F}_a^H = \mathcal{F}_a^S = 0$, but only anomalous (observed) wind stress, \mathcal{F}_a^τ , which was computed from the 1950-1979 COADS using a constant drag coefficient ($c_d = 1.5 \cdot 10^{-3}$).

In **experiment II**, in addition to the anomalous wind stress forcing of experiment I an anomalous heat flux, \mathcal{F}_a^H , was also introduced. The most straightforward way to do this would have been to use the observed COADS air temperature. However, we decided against this approach as it would have introduced a strong relationship between the observed air temperature and the SST which is just what we wish to test with the model.

As an alternative method we introduced a simple advective model for the surface air temperature. The basic idea is summarized in Fig. 1: Wind blowing from Siberia in wintertime is generally dry and cool. In this case we may expect an enhanced flux of energy into the atmosphere. Wind blowing northward from the tropical and subtropical Pacific, on the other hand, advects warm and moist air, and the turbulent energy fluxes into the atmosphere will generally be reduced. The details of the model are given in the Appendix.

4. RESULTS: SST ANOMALIES SIMULATED IN EXPERIMENTS I AND II

4.1 EOF analysis

Empirical Orthogonal Function (EOF) analysis is known to be a powerful method to derive coherent and energetic anomaly patterns from a noisy environment. We applied this technique to the observed SST fields and to the simulated SST in both experiments. Each data set contained 360 monthly mean SST fields.

The first three EOFs derived from the 360 months of observations are shown in Fig. 4. The three patterns explain more than 50% of the total variance of North Pacific SST. Frankignoul and Reynolds (1983) made an EOF analysis of SST of the North Pacific and the tropical Pacific. Their dominant mode had its center of action in the El Niño region. In the North Pacific, it is almost identical to our first EOF.

Experiment I

Fig. 5 shows the first three EOFs of SST simulated in experiment I. The associated coefficient time series of the first EOFs (observed and experiment I) are displayed in Fig. 6.

The first two EOF of experiment I are similar to the observed ones. The agreement between the third EOFs (Fig. 4c and 5c) is less good. The first three EOFs of the simulation explain only 40% of the variance, as compared to more than 50% for the observations. The observed and simulated coefficient time series of the first two EOFs (Fig. 5a,b) agrees well in phase, but the simulated amplitudes are only one third of the observed amplitudes.

Experiment II

The first three EOFs of SST simulated in experiment II are shown in Fig. 7. The associated coefficient time series (observed and simulated) are displayed in Fig. 8.

The first simulated EOF (Fig. 7a) has no similarity with the first two observed EOFs. It is the dominant mode explaining 38% of the variance, which is about as much as the first two observed EOFs together. Its maximum lies in eastern subtropical part of the ocean. However, it shows some general

Figure 4:

EOF analysis of the observed SST anomalies derived from 1950-1979 COADS (360 months).

- a. first EOF, explained variance: 23%
- b. second EOF, explained variance: 19%
- c. third EOF, explained variance: 11%

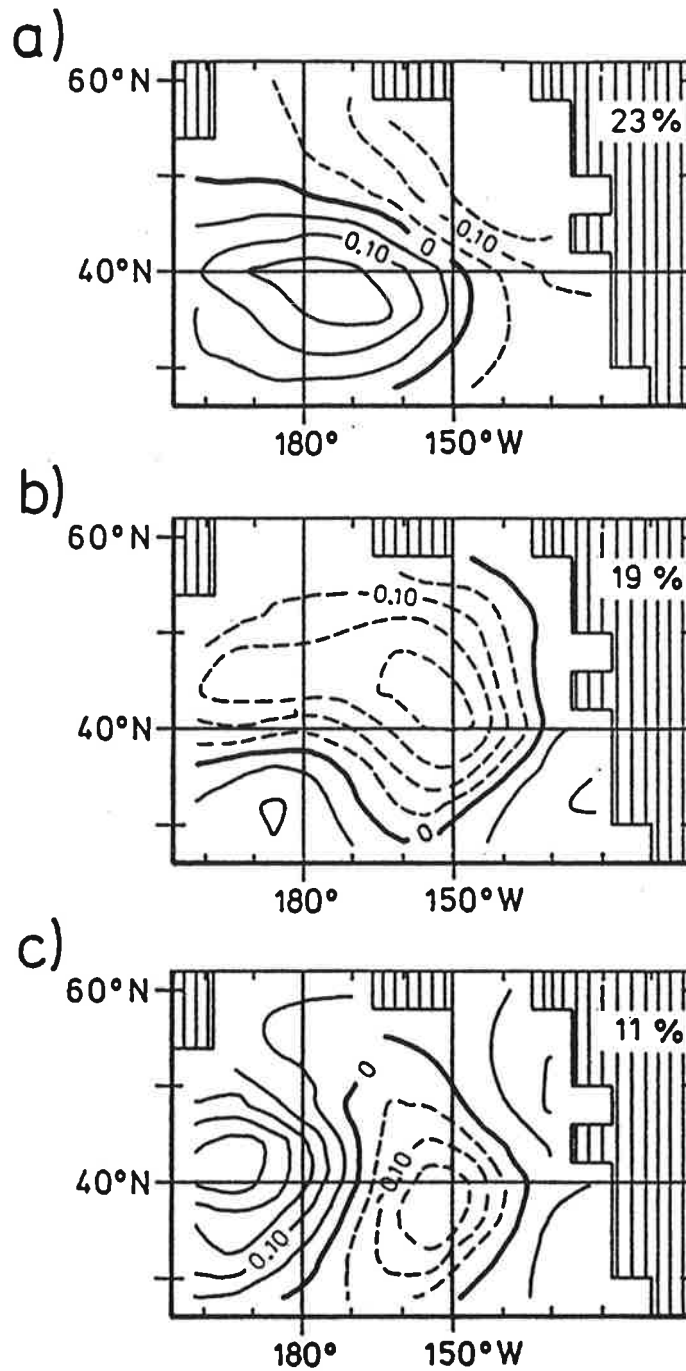


Figure 5

EOF analysis of the SST anomalies simulated in experiment I (no anomalous heat fluxes) based on 360 monthly means.

- a. first EOF, explained variance: 17%
- b. second EOF, explained variance: 12%
- c. third EOF, explained variance: 11%

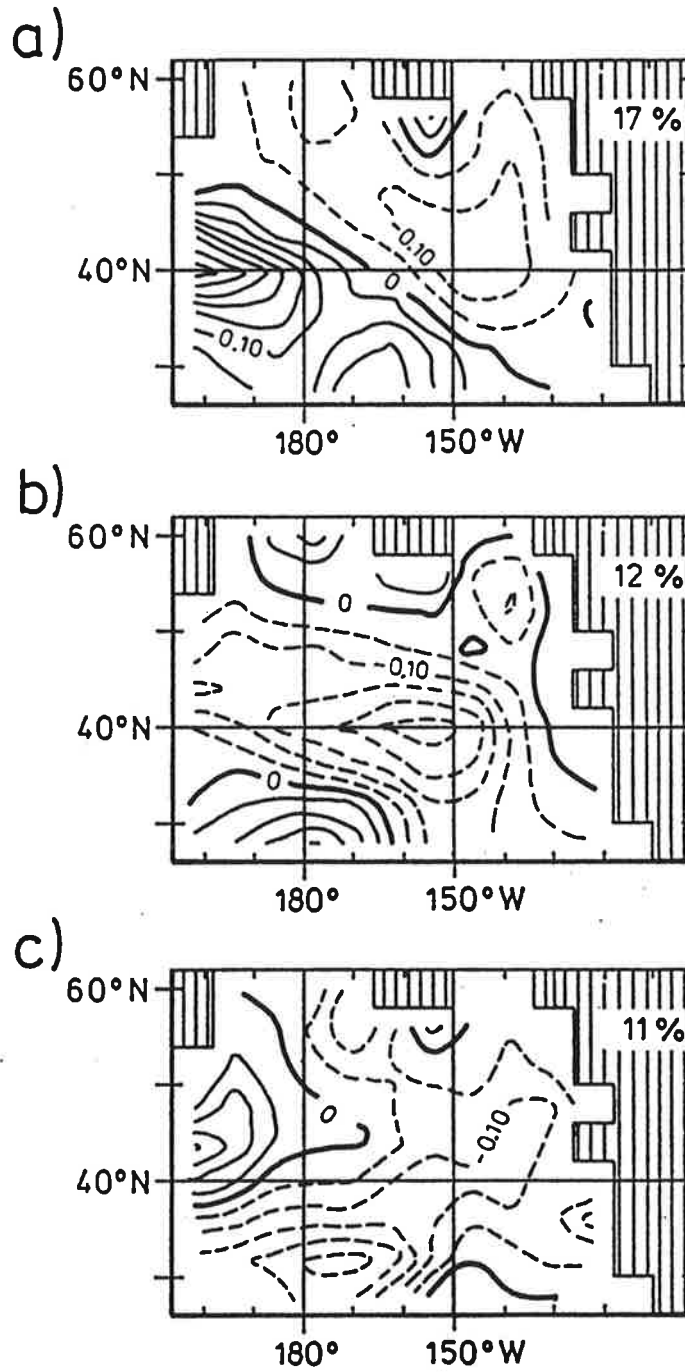


Figure 6:

Coefficient time series of the first two observed (continuous line) and experiment I (dotted line) EOF. The data are smoothed by a 5-months running mean filter.

- a. first EOFs
- b. second EOFs.

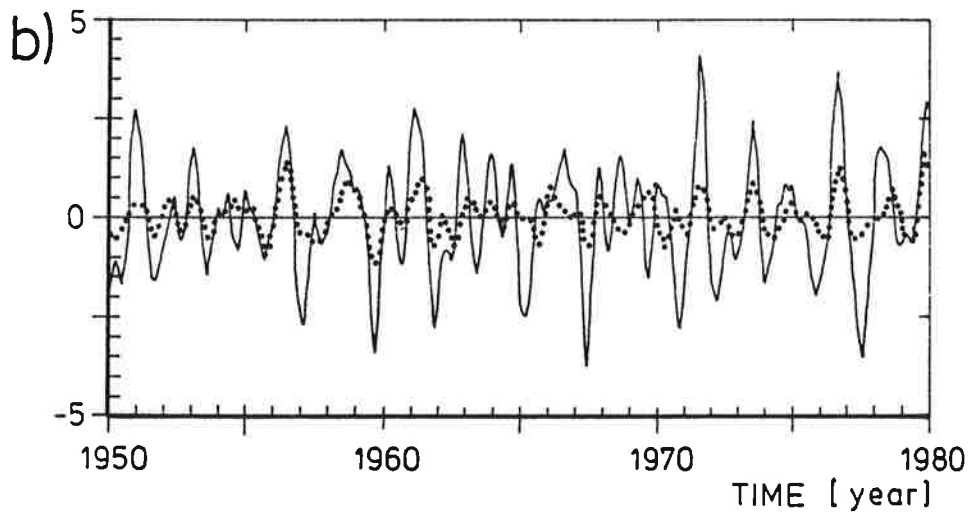
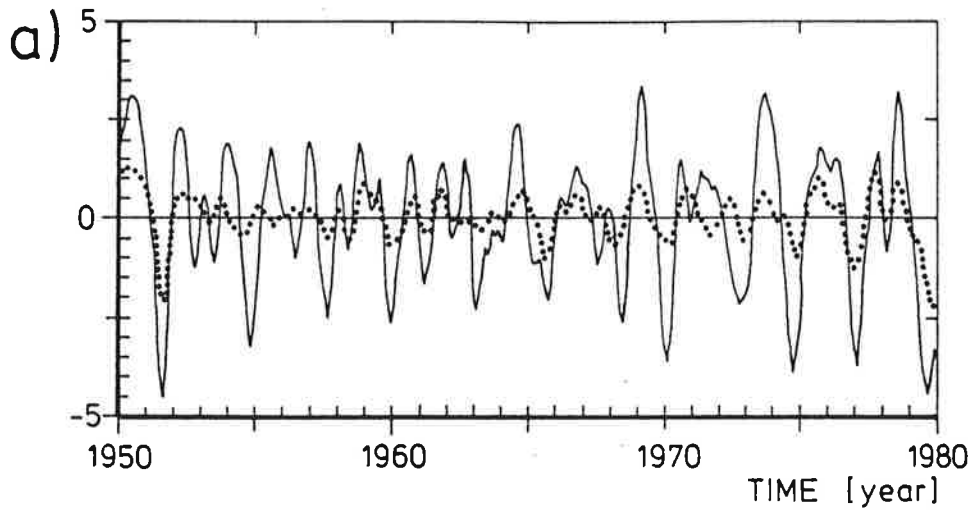


Figure 7:

EOF analysis of the SST anomalies simulated in experiment II (anomalous heat fluxes derived from the temperature advection model) based on 360 monthly means.

- a. first EOF, explained variance: 38%
- b. second EOF, explained variance: 15%
- c. third EOF, explained variance: 11%

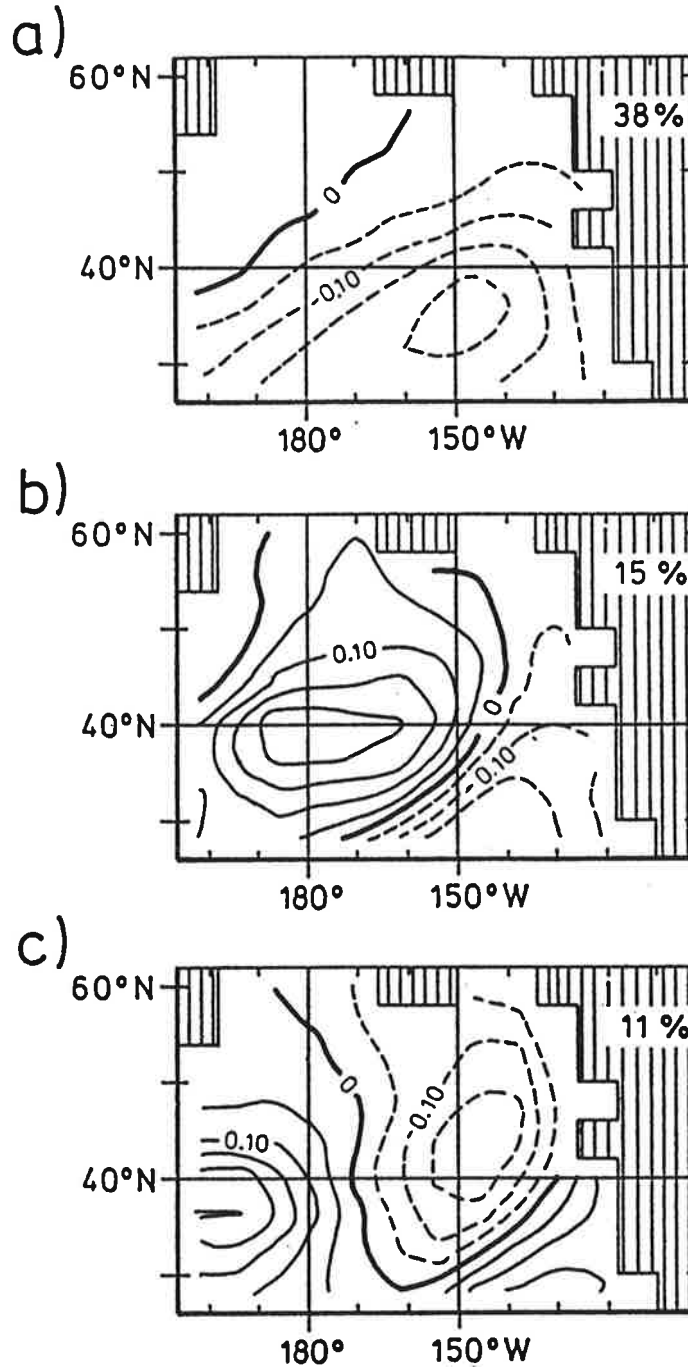
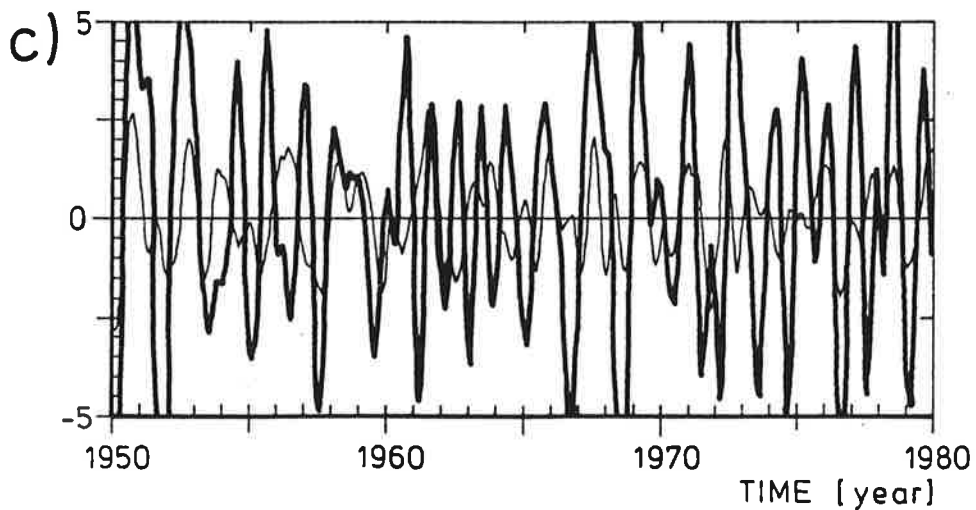
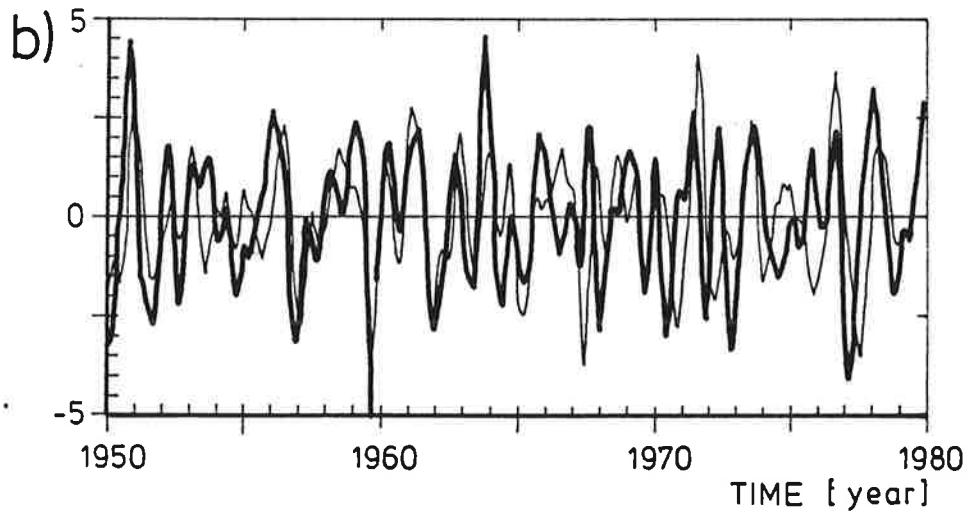
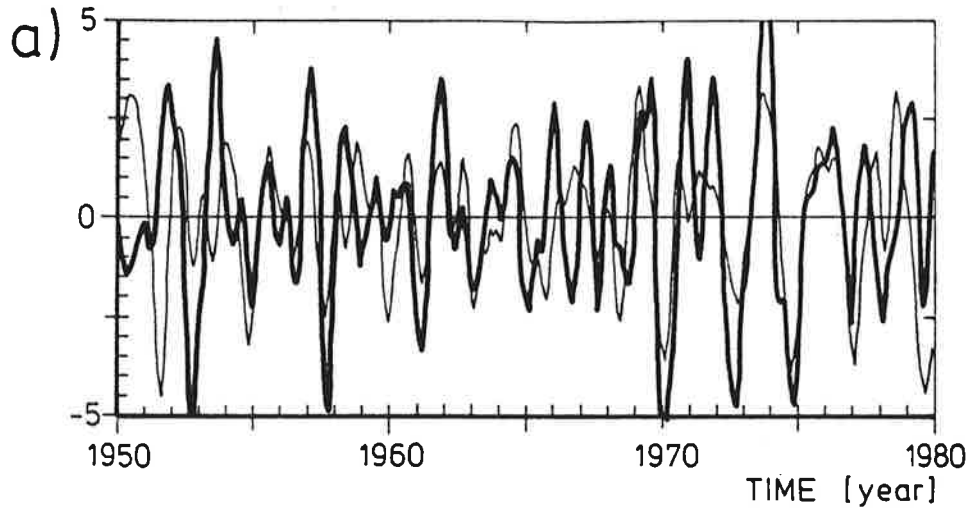


Figure 8:

Coefficient time series of the first three observed (continuous line) and experiment II (heavy line) EOF. The data are smoothed by a 5-months running mean filter.

- a. first observed EOF, second simulated EOF
- b. second observed EOF, third simulated EOF
- c. third observed EOF, first simulated EOF



similarity with the third observed EOF noting that the variability in the south eastern part of the analysis domain is highly overestimated. Consistent with this interpretation of a reordering of the EOF sequence, we note that the simulated EOFs 2 and 3 are similar to the first and second observed EOFs, and the coefficient times series are also similar both in phase and amplitude.

If the bulk transfer coefficients for the parameterization of the sensible and latent heat flux depends on wind speed and stability (which is not given in the presented experiments) the variance of the first EOF decreases and the variances for the next two EOF pattern increases. Nevertheless the overestimation of the variability in the subtropical area is still there.

4.2 Dependence on the seasonal cycle: correlation between simulated and observed SST

During summer the energy of the low-frequency modes of variability over the North Pacific weakens considerably (Horel and Wallace, 1981; Barnston and Livezey, 1987) and the upper mixed layer is shallow (Section 1). We therefore expect the ocean model to be more successful in reproducing the observed SST anomalies in winter than in summer. To test this hypothesis the correlation between simulated and observed 1950-79 SST was calculated separately for January and July.

The statistical stability of the results was tested using a univariate local significance analysis. As the SST for a particular January is approximately independent of the SST in the following or the previous January, the number of independent samples was taken as 30. This yields a rejection of the null-hypothesis of no local relationship between the observed and simulated SST 95% (99%) confidence if the estimated correlation coefficient is greater than 0.31 (0.42).

Experiment I

In January the correlations are mostly positive (Fig. 9a). The simulated winter SST along 40°N is significantly correlated with the observations. In these regions the anomalies should be mainly produced by advection and turbulent mixing. The negative correlations in the region north of 50°N could be related to the "wall" at 60N , which inhibits all heat exchange with the polar region (see Section 2.2).

Figure 9:

Correlation coefficients between observed and simulated SST (experiment I) in (a) January, and (b) in July. Areas with positive correlation coefficients are stippled and coefficients greater than 0.25 (0.50) are hatched (cross hatched).

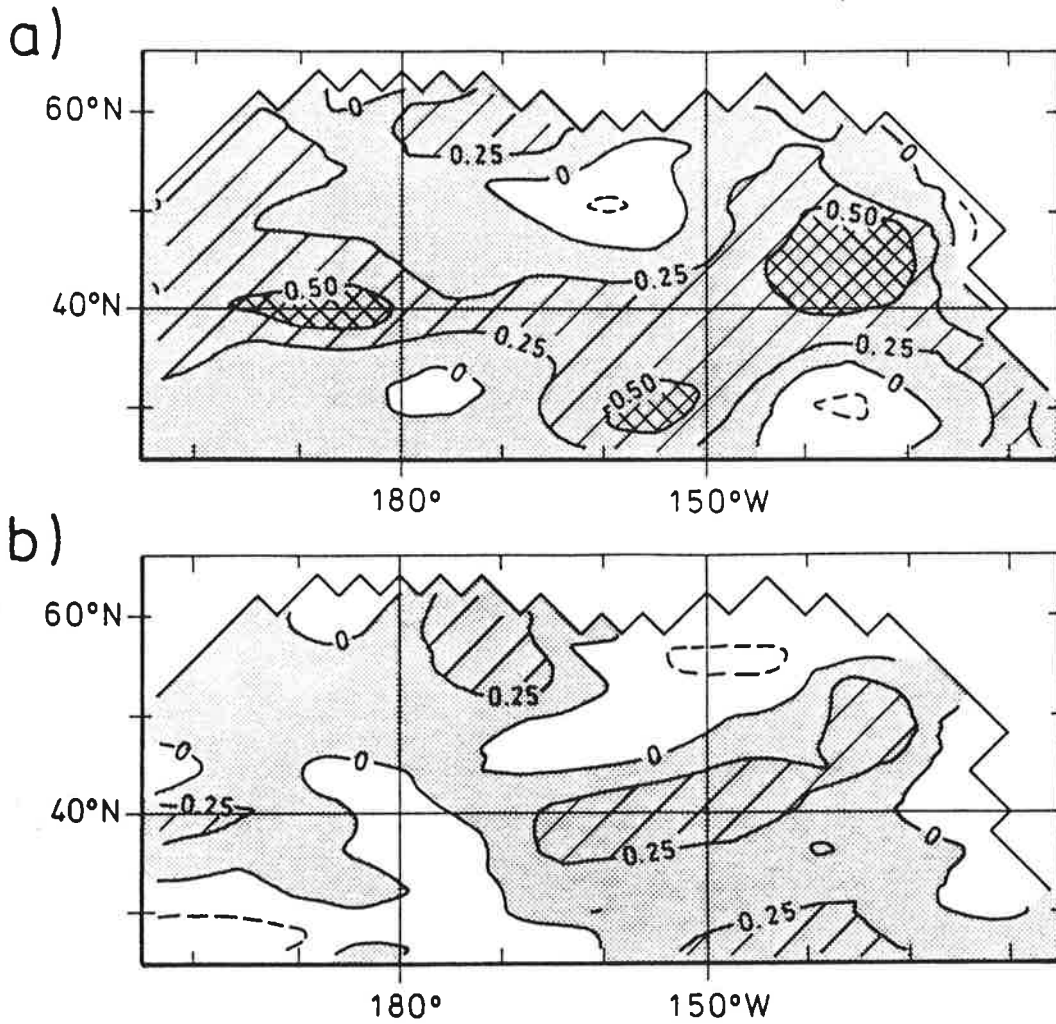


Figure 10:

Correlation coefficients between observed and simulated SST (experiment II) in (a) January, and (b) in July. Areas with positive correlation coefficients are stippled and coefficients greater than 0.25 (0.50) are hatched (cross hatched).

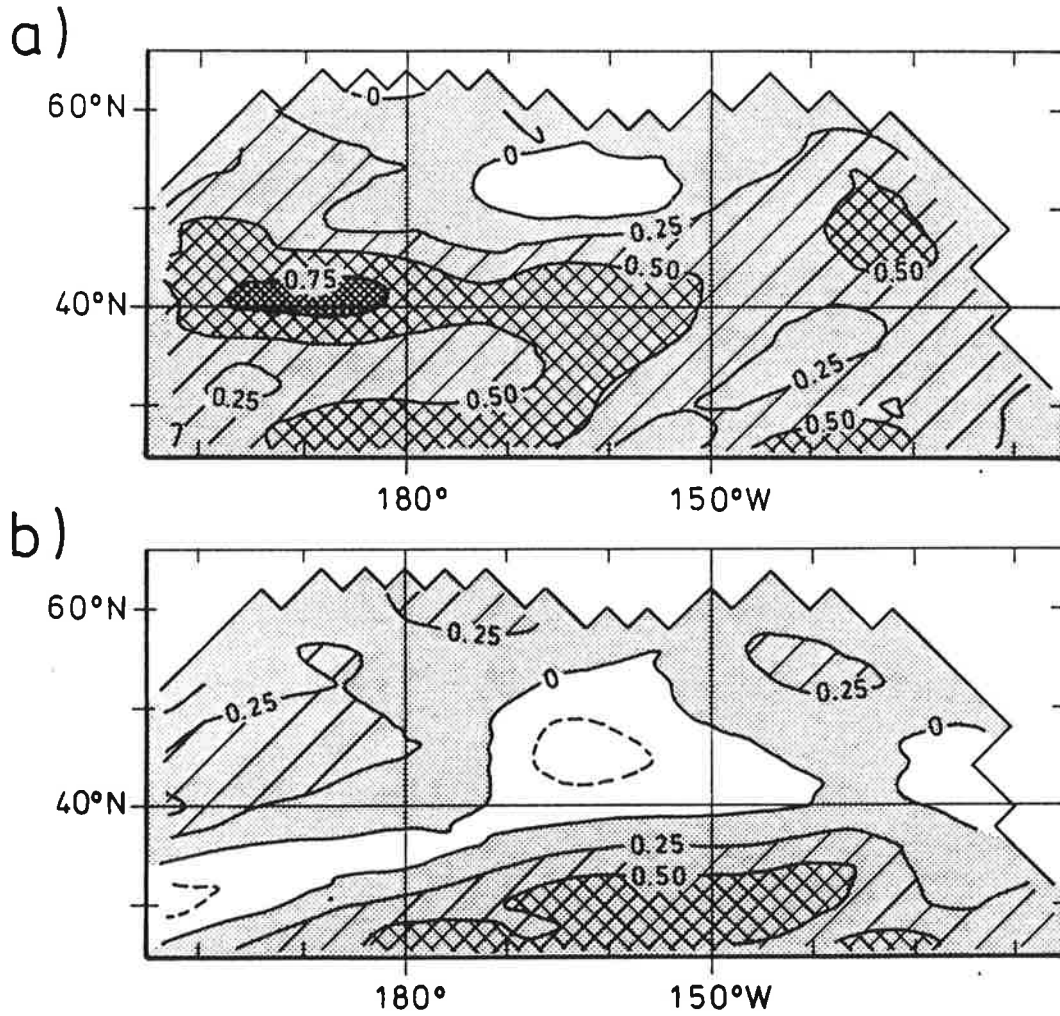
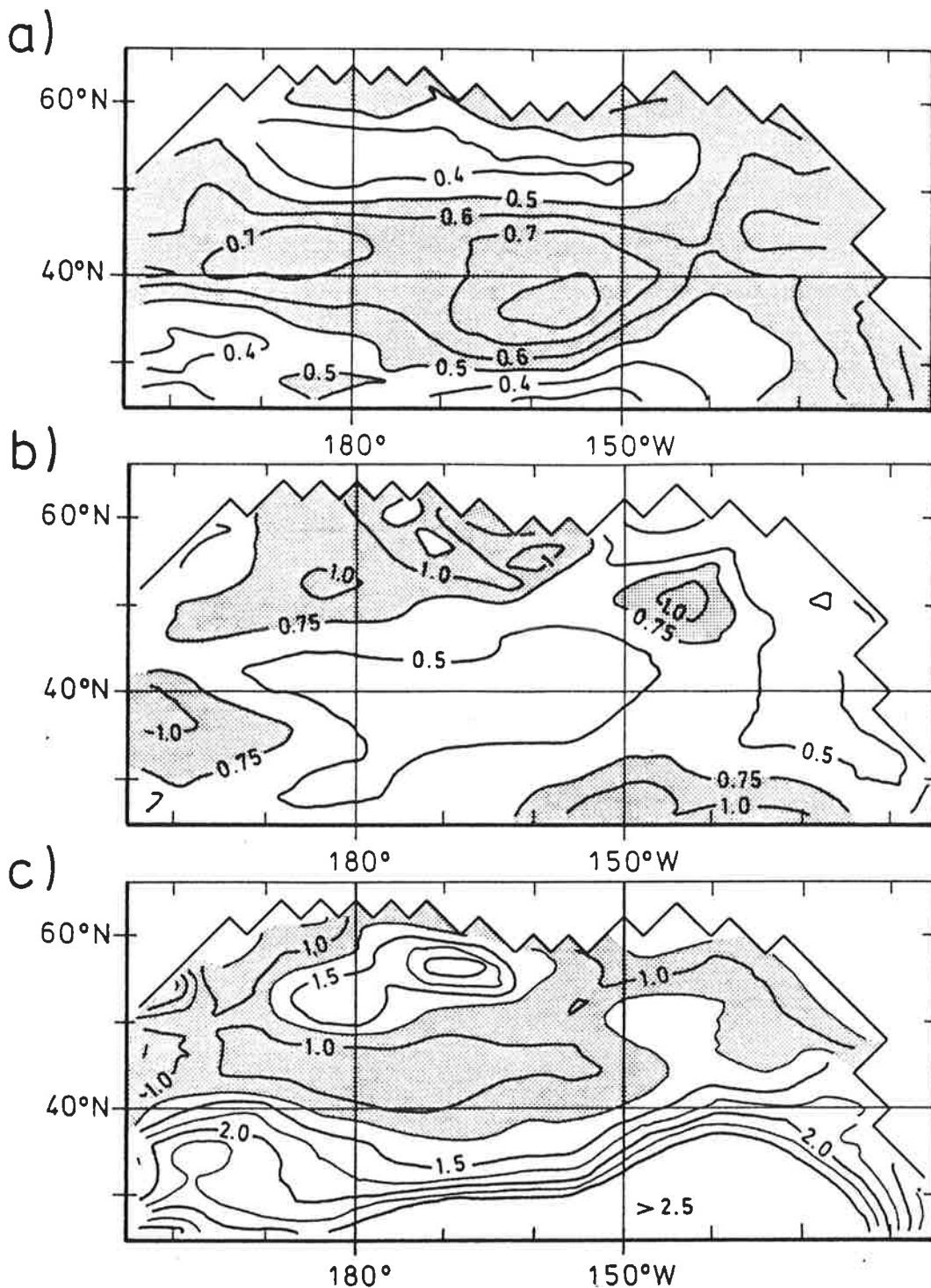


Figure 11: Month-to-month variability of observed and simulated January SST

- a. Standard deviation of the observed SST anomalies (COADS, 1950-79). Areas with standard deviations greater than 0.5°C are stippled. Contour interval 0.1C.
- b. Ratio of standard deviations of experiment I and observed January SST. Ratios less than 1 indicate an underestimation by the model. Areas with ratios of $1 \pm 25\%$ are stippled. Contour interval 0.25.
- c. Ratio of standard deviations of experiment II and observed January SST. Ratios greater than 1 indicate an overestimation by the model. Areas with ratios of $1 \pm 25\%$ are stippled. Contour interval 0.25.



In July, the correlations are considerably smaller. In large areas, the correlations are even negative, and the area of 95%-significance is small (Fig. 9b).

Experiment II

The correlations between simulations and observations are generally higher in this experiment, in which anomalous heat fluxes are introduced as additional forcing (Fig. 10). In January the correlation coefficients in almost the entire North Pacific are significantly positive. The major exception is the area of negative correlations in the northern part near 160°W , 50°N . The same deficiency was found in experiment I (Fig. 9a).

In July (Fig. 10b), the correlations are mostly positive. They are considerably smaller, however, than in January particularly in midlatitudes. In the subtropics, there is a large area in which the correlations are significantly positive.

We find again that the connection between atmospheric low-frequency variability and SST (north of the subtropical region) is significantly stronger in winter than in summer. The following discussion is therefore limited to the winter season.

4.3 Further Analysis of the simulated SST in January

The EOF analysis indicated that the overall level of the simulated SST variability is too low in experiment I and fairly realistic in experiment II. To examine the simulated SST variability in more detail, the month-to-month standard deviation of observed SST and the ratios of the standard deviations of the simulated and observed SST were calculated (Fig. 11). Maximum standard deviations of the observed SST of more than 0.7°C are found in the 40°N -belt of maximum westerly wind and along the American coast (Fig. 11a). Minimum values of less than 0.4°C are located in the trade wind region and in the northern part of the ocean.

In experiment I, the modeled variability is underestimated almost everywhere, as expected (Fig. 11b). In the central part of the ocean the modeled standard deviation is less than half of the observed standard deviation, which is maximal in this area. This result is in agreement with the results of

Frankignoul and Reynolds (1983) and Haney et al. (1985) mentioned in Section 1.

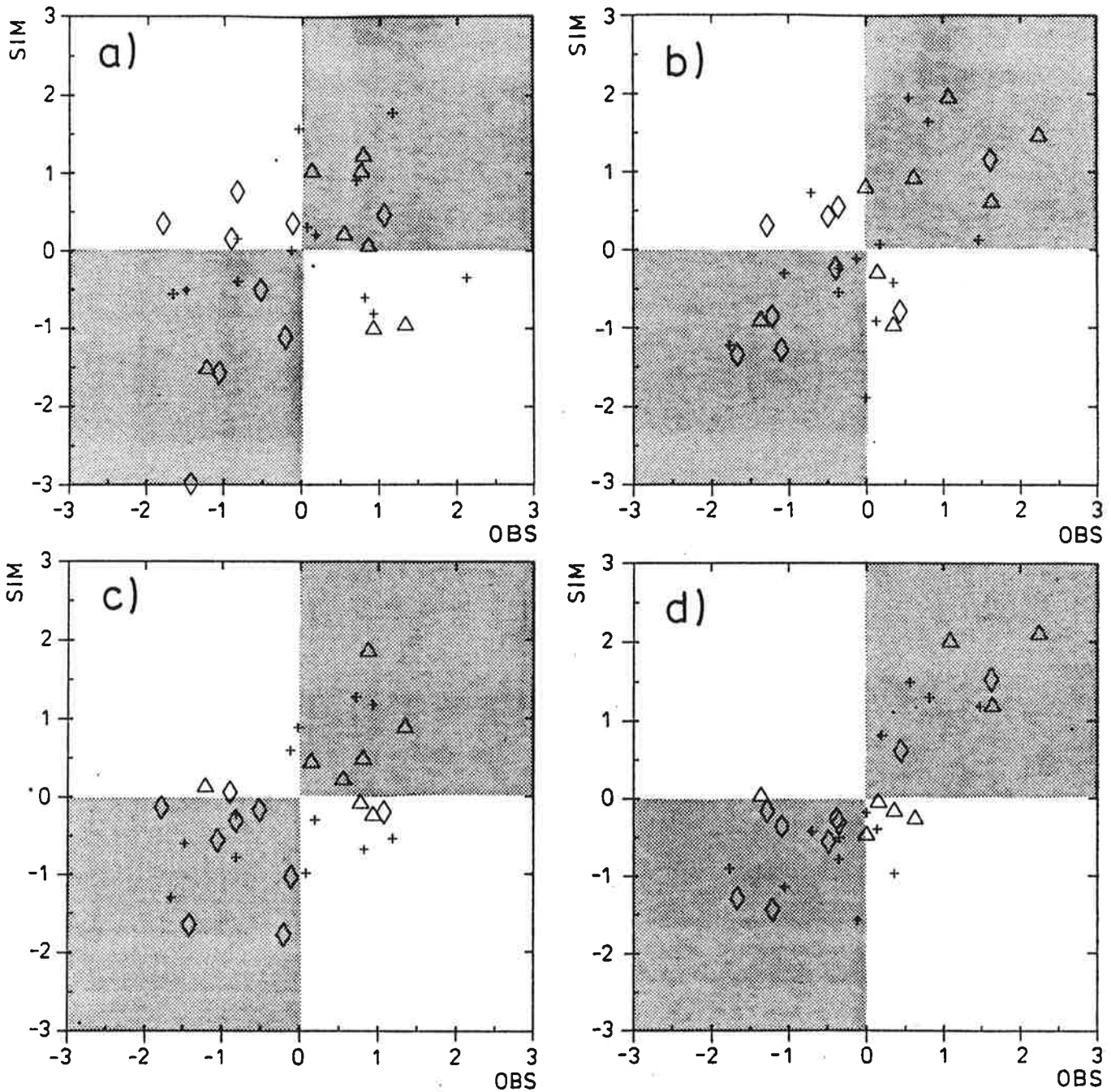
In Experiment II, on the other hand, the modeled variability is almost everywhere larger than observed (Fig. 11c). The overestimation is strongest south of about 35°N . The maximum ratio is found at the southern boundary of the experimental domain, where the modeled standard deviation is up to 450% of the observed (in this area the observed variability is small). The severe overestimation in the subtropics is presumably related to the introduction of a strong advective term in the atmospheric forcing (see also Section 4.1).

Fig. 12 shows scatter diagrams of observed and simulated area averaged January SSTA time series. Two areas (indicated in Fig. 1) were chosen: area A, 162°W - 178°W , 36° - 40°N , covers the maximum of the first observed EOF (Fig. 4a) and area B, 166°E - 178°W , 36° - 44°N , is located west of the dateline, where the correlations in both experiments are large (Figs. 9a and 10a). The modeled and observed area averaged SSTAs exhibit a fair degree of similarity, which is reflected in the correlation coefficients r : 0.32 (experiment I) and 0.58 (experiment II) in the first area (A), and r =0.60 (experiment I) and 0.78 (experiment II) in the second area (B). These numbers are significantly nonzero (at the critical level of 0.31 and 0.42 correspond to 95% (99%) confidence; as discussed in Section 4.2). In both areas, the variations of SST in experiment I are drastically underestimated. This cannot be seen in Fig. 12, because the anomalies are normalized by the standard deviation. The addition of anomalous heat fluxes related to the anomalous atmospheric circulation clearly improves the simulation in both areas (Fig. 12 c,d). The agreement between observed and modeled data is best in the 1970's and worst in the mid-1960's (not shown).

Figure 12:

Scatter diagram of area averaged time series of modeled and observed January normalized SSTA. Area A: 162°W - 178°W , 36° - 40°N ; and area B: 166°E - 178°W , 36° - 44°N ; see Fig. 1. Triangles indicate Warm Events, and diamonds Cold Events following the classification by van Loon and Madden (1981). Units: standard deviations of observed or modeled SST.

- observations vs. experiment I in area A
- observations vs. experiment I in area B
- observations vs. experiment II in area A
- observations vs. experiment II in area B



5. REMOTE RESPONSE TO ANOMALOUS TROPICAL CONDITIONS: ENSO

The anomalous atmospheric flow in midlatitudes during extremes of the SO represents a superposition of the atmospheric response to remote anomalous tropical conditions, to weather noise, and, possibly, to local extratropical SST anomalies. It is therefore not surprising that the anomalous atmospheric flow, and also the associated oceanic response, exhibit significant variability between individual ENSO years (Hamilton, 1988). Nevertheless, the persistent anomaly flow pattern in the Northern Hemispheric winter atmosphere that tends to appear during extreme phases of ENSO (as discussed above, see also Mo and Livezey, 1986; Livezey and Mo, 1987) can be expected to be connected with a strong large-scale oceanic response.

To measure quantitatively the relation between North Pacific SST and the state of the SO, we correlated the standard SO index (air pressure difference between Darwin and Tahiti) and SST at each gridpoint. The resulting maps for the observations and the two experiments are given in Fig. 13. In the observations (Fig. 13a), negative correlations are found over most of the central and western North Pacific, and positive correlations over the northern and eastern part. This result is completely in accordance with previous studies by, e.g., Weare et al. (1976) or Wright et al. (1985).

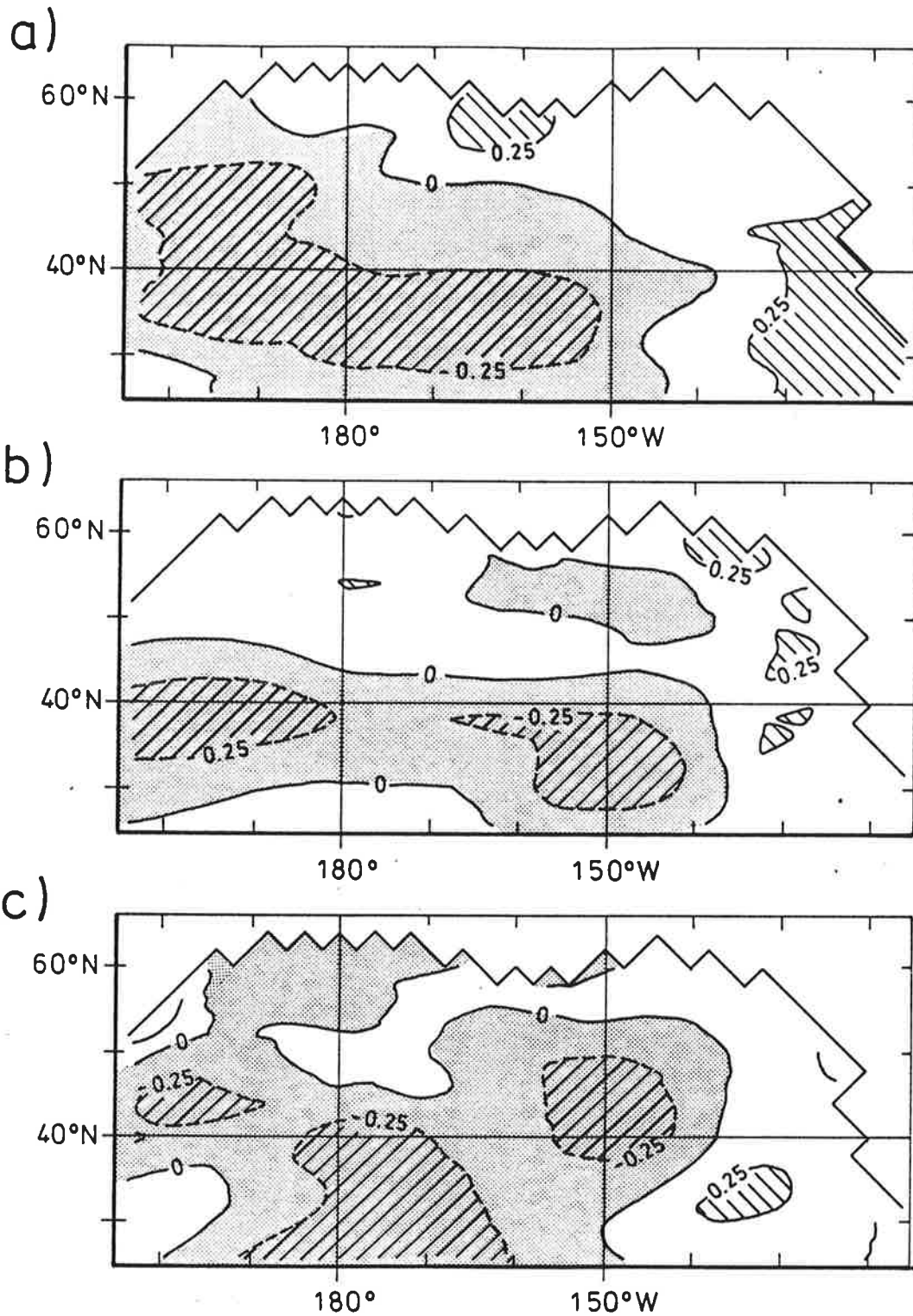
In both experiments (Fig. 13b,c) the main observed correlation pattern is reproduced, but the correlations are generally underestimated. In the eastern part of the ocean, this underestimation might partly be due to the strong damping of coastal Kelvin waves (Section 2; Pares-Sierra and O'Brien, 1989; Johnson and O'Brien, 1989). In the central North Pacific, experiment II seems to yield better results than experiment I.

In Fig. 12, the years of a Warm or a Cold Event, according to the classification of van Loon and Madden (1981) are marked by triangles and diamonds, respectively. When forced with anomalous stress and heat fluxes (as result to the anomalous atmospheric circulation; experiment II) the model reproduces the strong North Pacific SST anomalies associated with the extremes of the SO fairly well. In most Warm and Cold Events, the model simulates the correct sign of anomalies in both areas A and B.

Figure 13:

Correlation coefficients between observed SOI (air pressure difference Darwin and Tahiti) and SST in January. Areas with negative correlation coefficients are stippled and areas with coefficients smaller than -0.25 and greater than $+0.25$ are hatched.

- a. observed North Pacific SST
- b. simulated SST, experiment I
- c. simulated SST, experiment II



6. CONCLUSION

The main conclusions we draw from our study are the following:

- (1) It is possible to simulate the gross aspects of the observed anomalous North Pacific surface temperature with an OGCM associated with the observed anomalous motion of the atmosphere. The model is much more successful in winter, when the mixed layer is deep, than in summer, when the mixed layer is shallow.
- (2) The main patterns of the SST variability already are reproduced when the forcing is restricted to anomalous wind stress, without anomalous heat fluxes. In this case, however, the strength of the modeled SST anomalies are too weak by a factor of about 3. A similar result was obtained by Haney et al.(1985).
- (3) The inclusion of anomalous heat fluxes requires a model of atmospheric "surface" temperature as a function of SST and, say, prescribed temperature over land. The optimal way to solve this problem would be to use a coupled atmosphere-ocean GCM. As no such model was available for this study, however, we used a highly simplified model that considered only the horizontal temperature advection and heat exchange between atmosphere and ocean. Despite this simplicity, however, the results were clearly improved by inclusion of the heat fluxes compared to the case without anomalous heat fluxes. In the subtropics, however, the variability of SST is highly overestimated.
- (4) Observed and simulated SST anomalies are correlated with the state of ENSO. The model, particular with inclusion of forcing by both anomalous heat fluxes and anomalous stress, reproduces the observed correlation pattern.

The processes responsible for changes of SST are advection, turbulent mixing and heat fluxes. Our results indicate that the processes sensitive to wind stress anomalies, namely advection and turbulent mixing, produce response patterns which are similar to those produced by anomalous heat fluxes.

Our study reveals two cases of deficiencies:

- (1) The massive overestimation of the variance in the subtropical region in experiment II. We suggest that this may be due to the strong advection term in the atmospheric forcing intensified by the constant bulk transfer coefficients, which was used in the heat flux parametrization.
- (2) The present experiments do not provide a properly filtered SO signal in the forcing of the North Pacific SST. A major part of the anomalous

atmospheric flow may be seen as remote response to anomalous tropical conditions. However, the remaining part, which is unrelated to the SO, is certainly not small. To filter out the SO related North Pacific SSTA characteristics, we plan to perform experiments in which the anomalous forcing is restricted to the remote response to the SO.

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Appendix

The simple model to estimate anomalous heat fluxes used in **experiment II** is designed to describe the advection of climatological surface air temperature T_A by the observed time dependent wind field $\vec{u}_A(t)$ and its cooling or heating due to sensible heat fluxes into the ocean, Q_H :

$$(1) \quad \frac{d T_A}{dt} = \vec{u}_A \cdot \nabla T_A + Q_H$$

Over land, the surface air temperature T_A is held constant at its climatological value. The wind field $\vec{u}_A(t)$ is taken from the 1950-79 COADS. The sensible heat flux is estimated from a standard bulk formula using a constant bulk transfer coefficient of $c_H = 1.3 \cdot 10^{-3}$ (Hasselmann and Dörsner, 1986):

$$(2) \quad Q_H = \rho_A c_p c_H \cdot (SST - T_A) \cdot |\vec{u}_A|$$

SST is taken from the ocean GCM run, ρ_A is the density of the air, and c_p is the specific heat of the air at constant pressure. The equations (1) and (2) form a closed predictive system for the surface air temperature T_A that may be integrated parallel to the ocean model. The anomalous heat flux \mathcal{F}_a^H is calculated as the sum of estimated sensible and latent heat fluxes subtracted by the climatological values for the sensible and latent heat fluxes ($\mathcal{F}_c^S, \mathcal{F}_c^L$) derived from the control run:

$$\mathcal{F}_a^H = \rho_A \cdot |\vec{u}_A| \left[c_p c_H (SST - T_A) + c_E L (q - q_A) \right] - (\mathcal{F}_c^S + \mathcal{F}_c^L)$$

L is the latent heat of vaporization, q is the specific humidity at saturation near the water surface and q_A is the specific humidity of the air. To derive the specific humidity of the air the simulated air temperature T_A and the actual relative humidity after COADS is used (1950-1979; we get similar results for climatological values of the relative humidity). We assumed:

$$c_E = c_H.$$

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