Comparisons of GCM and Observed Surface Wind Fields over the Tropical Indian and Pacific Oceans

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ABSTRACT

Many of the processes that have important effects on both the climatological distribution and interannual variability of sea surface temperatures (SSTs) in the tropical oceans are greatly affected by the surface wind field. For this reason accurate simulation of the surface wind is a key factor governing the success of coupled tropical ocean–atmosphere models. This paper presents the results of two analyses that investigate the quality of wind fields produced by three general circulation models (GCMs) over the tropical Indian and Pacific oceans.

The first analysis concerns the annual cycles of the tropical wind fields simulated by versions of the GCM at the Oregon State University (OSU), European Centre for Medium Range Forecasts (ECMWF), and National Center for Atmospheric Research (NCAR). These models have similar horizontal resolutions but vary widely in vertical resolution. The results show that although there are substantial differences in model performance, apparently related to differences in vertical resolution, there are also clear similarities in their behavior. Each GCM did best in major trade wind regions and somewhat poorly in convectively active areas with light winds. This finding suggests that the formulations governing the interactions between persistent convection and the circulation may limit model performance.

A second analysis examines the response of the NCAR GCM, in terms of tropical Pacific wind stress, to prescribed SST anomalies over the period 1961–1972. It was found that the model response to SST anomalies associated with the El Niño/Southern Oscillation (ENSO) was distinct and in some respects resembled that of the real atmosphere. However, there were important discrepancies in the spatial configuration of the GCM field and in the amplitude of response of the GCM to the SST anomalies. An analysis of these discrepancies suggests that while the trapped equatorial Kelvin wave response of an ocean model coupled to this GCM would be qualitatively correct, differences in the GCM and observed forcing fields would result in large errors away from the equator. Tests with the Florida State University model of the tropical Pacific, to be reported in a later paper, support this conclusion.

Taken together, these findings suggest that while GCMs are capable of reproducing correctly many features of the tropical surface wind fields, discrepancies remain with both the annual cycle and the response to the anomalous SST patterns associated with ENSO. These discrepancies appear to be related, at least in part, to interactions between organized convection and the circulation. To what degree these differences would affect the oceanic component of a coupled model is currently under study.

1. Introduction

The surface wind field has important effects on many of the processes that determine tropical sea surface temperatures (SSTs) including thermocline depth, currents, and surface fluxes. Sea surface temperatures are, in turn, one of the key boundary conditions affecting the atmospheric circulation and the distribution of large-scale convection in the tropics. Thus, the success of coupled ocean–atmosphere models in studying the global climate in general, and such interannual patterns of variability as the El Niño/Southern Oscillation (ENSO) cycle (e.g., Cane and Zebiak 1985; Zebiak and Cane 1987; Schopf and Suarez 1988), depends

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to a major extent on the fidelity with which atmospheric component of such models can simulate real surface wind fields over the tropical oceans. This paper describes the results of two related analyses that examine the performance of three general circulation models (GCMs) from that perspective.

The first analysis focuses on the simulation of the observed annual cycle of wind and wind stress over the Pacific and Indian oceans by versions of the GCM at Oregon State University (OSU), National Center for Atmospheric Research (NCAR), and European Centre for Medium Range Weather Forecasts (ECMWF). These models have comparable horizontal resolutions but range in vertical resolution from 2 to 16 layers. In these simulations the annual cycle of the sea surface temperatures (SSTs) is prescribed; i.e., there are no interannual SST anomalies. Although the time-mean behavior of the tropical and extratropical wind fields produced by these and other models have been described briefly by others (e.g., Fischer 1987; Lau 1985; Schlesinger and Gates 1980; Pitcher et al. 1983), our analysis is more comprehensive and concerns only the tropical wind field. The results show that although the models vary substantially in their performance, each does best in the major trade wind regions and has the least consistent success in convectively active areas with light winds. This suggests that improvements in the model formulations that describe the interactions between persistent convection and the surface circulation may be a logical first step towards better simulations of the tropical wind fields.

The second analysis examines the response of the NCAR GCM surface wind stress over the tropical Pacific to prescribed SST anomalies evolving between 1961 and 1972. Similar analyses have been performed by a number of other workers (e.g., Blackmon et al. 1983; Keshavamurty 1982; Fennessy et al. 1985; Shukla and Wallace 1983) for fixed (non-evolving) SST anomalies, but only Lau (1985) and, very recently, Fennessy and Shukla (1988), have used evolving observed SST anomalies as a boundary condition. In no case have the model and observed responses in terms of surface wind stress been compared side by side as we do here. Our analysis shows that while the response of the model is similar in some respects to that of the real atmosphere, there are also important differences in the amplitude and configuration of the resulting wind field anomalies. In terms of ENSO-related changes in central equatorially Pacific zonal wind stress, the GCM response resembles that observed. There is less agreement away from the equatorial regions. In addition, the overall amplitude of the GCM response to anomalous SSTs is on the order of half that observed in the real atmosphere.

The presentation is organized as follows. The data and methods used in the study are described in section 2. This section includes discussions of the GCMs and the GCM surface wind data, the observed wind data and the statistical procedures applied. Section 3 describes the comparison of the annual cycle of the GCM surface winds with the observed circulation. This is followed by an analysis of the response of the NCAR GCM to prescribed SST anomalies in section 4. A brief summary and concluding remarks are given in section 5.

2. Data and methods

a. Observed wind data

The observed wind velocity dataset was compiled at the Scripps Institution by merging an existing Pacific trade wind set (Wyrtki and Meyers 1976) with ship and station data from the Pacific and Indian oceans using objective analysis (Barnett 1977). The data are monthly averages on a $2^\circ \times 10^\circ$ latitude-longitude grid covering the Indian and Pacific oceans between $30^\circ$N and $30^\circ$S. The period 1950–1984 was used to construct the statistics for this study.

The observed wind stress set was assembled at Florida State University (FSU) from daily hands of tropical Pacific ship and station data on a $2^\circ \times 2^\circ$ grid as described in Goldenberg and O'Brien (1981). The values given in the dataset are referred to as pseudostresses because the air density and drag coefficients were not used in their computation. Pseudostress is referred to simply as wind stress in this paper. The FSU wind stress data have been used extensively to drive tropical ocean circulation models (e.g., Busalacchi et al. 1983; Busalacchi and Cane 1985). In this study we have used the data for 1961 (the beginning of the dataset) through 1972.

The NCAR GCM stresses ($\tau$) were converted to pseudostresses ($\tau^*$) to make the datasets consistent. This conversion was made using the density and drag values used to calculate wind stress in the NCAR model, i.e.,

$$\tau^* = \frac{\tau}{\rho C_d},$$

where $\rho$ is the air density ($1.225 \text{ kg m}^{-3}$) and $C_d$ is the momentum eddy exchange coefficient ($\times 10^{-3}$) used in the GCM.

b. General circulation models and simulated wind data

Three GCMs were used in our comparisons. Each includes the dynamical effects of orography and the hydrologic cycle, and treats such processes as solar and infrared radiation, convection and surface (land and ocean) heat fluxes, although the specific treatments and parameterizations vary from model to model. The GCMs have comparable horizontal resolutions (about 300 km in the tropics) but widely differing vertical resolutions (2 to 16 layers). Each has been shown to simulate realistically many major aspects of the annual cycle of the general circulation.
The OSU model is a finite difference model that operates on a $4^\circ \times 5^\circ$ latitude-longitude grid with two vertical layers. Surface wind data are extrapolated from about 2000 m above the surface (the $\sigma = 0.75$ level). Descriptions of the model and results of annual climate simulations may be found in Ghan et al. (1982) and Schlesinger and Gates (1980). The data used here come from a 10-yr integration using long-term climatological (monthly) SSTs.

The ECMWF GCM is a spectral transform model and the version used here (henceforth ECT21) has 16 vertical layers and uses triangular-21 truncation giving a spatial resolution of about 300 km. Surface winds are for the lowest level of a Monin-Obukhov boundary layer. Computational aspects of the model and the results of annual cycle simulations are described by Fischer (1987). As with the OSU model, the data come from a 10-yr integration using climatological SSTs.

The version of the NCAR GCM referred to here is a nine-layer spectral transform model (rhombooidal-15 truncation) with a resolution of about 350 km in the tropics. Reported surface wind stresses are calculated from the winds at about 100 m above the surface (the $\sigma = 0.991$ level). Pitcher et al. (1983), Blackmon et al. (1983), and Chervin (1986) give computational details of the model and describe the results of climatological- and variable-SST experiments. The wind stress data from two 15-yr integrations are used here. One run (CONTROL) used climatological SSTs and the other (FORCED) used analyzed (observed) SSTs for the period 1958–1972. (The SST data are those described by Oort, 1983.) In our analyses we use only the data for 1961–1972 for which comparative observed wind stress data are available.

c. Comparison methods

To perform the data comparisons on a single grid, all of the datasets were interpolated linearly onto the $2^\circ \times 10^\circ$ mesh of the Scripps data. The OSU and ECT21 model winds are compared with the Scripps dataset for the Indian and Pacific oceans, and the NCAR GCM wind stresses are compared with the FSU wind stress set over the tropical Pacific. For the NCAR GCM comparisons, monthly means for the wind stress data were calculated over the 12-yr period 1961–1972 when the FSU and FORCED data overlap (years 4–15 of the model runs).

In performing the analyses described below, it became clear that the NCAR wind stresses were biased upward in comparison with the observations. This is due, in part, to the fact that the winds from about 100 m above the surface were used to calculate the wind stress. Calculations using a logarithmic wind profile and an oceanic roughness length of 0.1 cm (Roll 1965) give a ratio of the wind speed at 100 m to the usual 10-m reporting height of about 1.25. This would result in model stress magnitudes that are about 1.6 times greater than those that would be experienced at 10 m. Adjustments of the GCM wind stress by this factor greatly reduce the systematic bias. Therefore, in some of the figures and descriptions, the NCAR surface wind stress data have been reduced by a factor of 1.6 to be more representative of the surface values. Where this has been done, it is explicitly mentioned in the text.

A simple method of comparing the annual cycles of modeled ($V'$) and observed ($U'$) fields is based on the representation of the annual cycle of the observed field by its first two empirical orthogonal functions (EOFs). This is done by expressing the long-term monthly means ($U_{it}$) of the wind components as departures ($U'_{it}$) from their respective annual means ($\bar{U}_i$), i.e.,

$$U'_{it} = U_{it} - \bar{U}_i.$$  

Here the subscripts $i = 1, \ldots, nx$ and $t = 1, \ldots, nt$ denote the space and time indices over $nx$ locations and $nt$ realizations, respectively (for example, $nt = 12$ months in the case of the annual cycle); EOFs of these data are then calculated as usual (e.g., Barnett 1977) by decomposing the covariance matrix of $U'$ yielding eigenvectors ($E$) and amplitudes ($\alpha$). Here we define the amplitudes as

$$\alpha = EU'$$

so that the variance of the amplitudes of each mode is equal to the associated eigenvalue. In most climatic data the amplitudes of the first two EOFs of the annual cycle represent the quadrature components of a one-per-annum cycle and account for a large part of the variance of the annual cycle.

Comparisons with model data are made by projecting the GCM annual cycle onto the EOFs of the observed annual cycle. This is done by subtracting the long-term annual mean of the observed data from the monthly means of the model data, i.e.,

$$V'_{it} = V_{it} - \bar{U}_i.$$  

This preserves any bias between the model and observed annual means. The model data are then projected onto the observed EOFs by multiplication,

$$A^* = EV^*,$$

where $A^*$ represents the amplitudes of the projections. It can be shown that nonzero overall means of the projection amplitudes result only from differences between the model and observed long-term annual means projected onto the EOFs, i.e.,

$$E(\tilde{V} - \bar{U}),$$

where $\tilde{V}$ is the annual mean of the simulation.

The wind field EOFs described in this paper were calculated by treating the zonal and meridional components of the wind data as separate scalar variables. This differs from the method outlined by Hardy and Walton (1978) (cf., Legler 1983) in which each wind
observation is treated as a complex variable, \( u + iv \), and the EOFs are calculated from the complex covariance matrix of these data. The resulting complex EOFs have spatial expressions whose real and imaginary components are weights on the magnitude (i.e., speed) and phase (i.e., direction) of the wind at each location. The temporal expressions are also complex variables, the real and imaginary parts representing at each time, a scaling and rotation of the vectors at each spatial location. Debate has arisen concerning the relative merits or validity of these two methods. For the most part, this controversy is unwarranted; the two techniques are simply alternative techniques for orthogonal decomposition of the covariance of a two-dimensional vector field. Either can be used to depict the major patterns of wind field variability and to express the dataset as the superposition of a relatively few modes, and each has advantages and disadvantages. The complex method has the useful attribute that a particular pattern can be expressed by one mode even if the vectors in that pattern are rotated through the same arbitrary angle. In some cases, real vector fields in observed fluids might show such behavior. Frequently, however, the flow at differing locations rotates in opposite directions (or at different rates) through time, so that two modes are required to express a single systematic pattern of evolution through time. The scalar method always requires two modes if rotation is involved (i.e., if the pattern does not simply reverse). Because the temporal expression of the complex modes have two components (i.e., two degrees of freedom), a given fraction of the total variance is usually captured in fewer modes than with the scalar technique. However, one of the drawbacks of the complex method is that relating the behavior of a complex temporal series to some reference variable (e.g., eastern Pacific SST) is not as straightforward as with the scalar technique. Similarly, the use of the complex technique as a pre-conditioning step for regression analysis would be considerably more complicated than with the scalar method. Thus, each technique has benefits and drawbacks that should be recognized and balanced in light of the physical attributes of the field under consideration and the desired results.

A second method of comparing the annual cycles of the modeled and observed fields is the measure

\[
\Delta_x = 1 - \frac{\langle e^2 \rangle^{1/2}}{\langle |U| \rangle},
\]

where

\[
\langle e^2 \rangle = \frac{1}{nt} \sum_{i=1}^{n_t} \left( (V_{it} - U_{it})^2 + (V_{it} - U_{it}^y)^2 \right)
\]

\[
\langle |U| \rangle = \frac{1}{nt} \sum_{i=1}^{n_t} \left( U_{it}^x + U_{it}^y \right)^{1/2}.
\]

Here the superscripts \( x \) and \( y \) indicate the zonal and meridional components of the wind, respectively. Note that \( \langle e^2 \rangle \) is a variance-like quantity; it is the mean of the squared length of the residual vector \( V - U \). The quantity \( \langle |U| \rangle \) is simply the mean wind speed. Although \( \Delta_x \) will be referred to as “skill” here, the reader should note it differs from the more traditional skill scores where the error variance is normalized by the sample variance. By the formulation given here, \( \langle e^2 \rangle^{1/2} \) (hereafter the rms error) is small in comparison with the mean of the magnitude of the observed monthly mean wind \( \langle |U| \rangle \), \( \Delta_x \) gives a skill near unity; rms errors of about the same size as \( \langle |U| \rangle \) give skills near zero and rms errors large in comparison with \( \langle |U| \rangle \) give negative skills.

A third useful comparative statistic is the rms error. This measure provides a convenient comparison with the skill parameter defined above and allows the effects of changes in mean wind speed to be ascertained.

In interpretation of the annual cycle comparisons, it is necessary to remain aware of the fact that the ECT21 and OSU GCM data are surface winds and are compared with observed surface wind climatology over the Pacific and Indian oceans. In contrast, the NCAR GCM data are surface wind stresses and are compared with surface wind stress climatology over the Pacific Ocean only. In addition, the Scripps surface wind climatology was generated in part from objectively interpolated averages of ship winds, while the FSU surface stress climatology is constructed from hand (subjective) analyses based on ship and station data. These differences make one-to-one comparison of the NCAR results with those from the OSU and ECT21 GCMs impossible. Although this fact may limit the generality of our conclusions, the use of the surface stress data from the NCAR model and its comparison to the FSU climatology (which is available only over the Pacific) represents a first step in ocean model simulations in which the NCAR FORCED and CONTROL simulation wind stresses have been used to drive the FSU linear tropical ocean model. The results of these ocean model experiments (referred to occasionally here and to be reported in a later paper) will be related to the comparisons of the wind stress data presented in this paper.

The analysis of the response of the NCAR GCM to prescribed SST anomalies uses both EOF and canonical correlation analysis (CCA) to isolate and quantify the response of the model to changes in SSTs. While EOF analysis has been used frequently in the studies of interannual variability (e.g., Davis 1976; Barnett 1977), CCA has seen application in this regard only recently (e.g., Nicholls 1987; Barnett and Preisendorfer 1987; Graham et al. 1987a,b; Graham and White 1988), despite the potential advantages of the method pointed out by Glahn (1968). In view of its unfamiliarity, a brief description of CCA is given below. Formal derivations and complete discussions of the method may be found in many statistical texts (e.g., Tatsuoka 1971).

Canonical correlation analysis is used to find the linear combinations of two datasets (fields) which are
most highly correlated. Characterized by Tatsuoka (1971) as "double barreled principal component analysis," the CCA procedure decomposes the between-field covariance into separate, orthogonal modes that account for as much of the covariance as possible. Like EOF analysis, each of these modes (or canonical variables) is defined by a spatial and a temporal function, but in CCA there are separate functions for each field. The squared correlation between the temporal functions for each field for a particular mode is the "canonical correlation" and gives a measure of the association between the two spatial functions. Simply put, when the canonical correlation is high the spatial patterns shown for each field tend to occur together.

We apply here a powerful variant of the CCA procedure, in which CCA is combined with EOF analysis. This technique, described in detail by Barnett and Preisendorfer (1987) and Graham et al. (1987a), reduces the computational resources required, lowers the noise level in the data, and aids in relating the within-field to the between-field covariability. Although this approach is somewhat more complicated than the traditional procedure, the mathematical foundation is identical.

The reader will note that we have not attempted to estimate the statistical significance of the findings presented in this paper. Though values could be assigned to such estimates, we feel that the 10–12 yr period of the annual cycle comparisons, and the presence of only four ENSO events in the model response analyses, would result in very large uncertainties in the estimated confidence intervals. Where possible, we have attempted to relate features noted in the discussions to physical processes or other data. However, due to the small number of samples available, the findings presented here should be regarded as tentative. Longer (or additional) simulations will be required to ensure that the relationships described herein are consistent through time.

3. Comparison of the simulated and observed annual cycles

This section presents comparisons between the simulated and observed annual cycles of the surface winds. The section opens with a brief description of the major features characterizing the observed cycle over the Pacific and Indian oceans. Following that, each of the models is compared with its respective observation set. These comparisons examine the differences between the model and observed fields in terms of residual (model — observed) wind fields, skill and rms error maps, and projections of the model data onto the EOFs of the observed annual cycle.

a. The observed annual cycle of surface winds over the Pacific and Indian oceans

The annual cycle of the wind field over the Pacific and Indian oceans has been described in detail by a number of workers (e.g., Horel 1982; Wyrtki and Meyers 1976; Goldenberg and O'Brien 1981; Hastenrath and Lamb 1977, 1979; Fu et al. 1983). In view of these in-depth analyses, we give only a brief summary which will serve to guide the following discussions. This summary is accompanied by maps (Fig. 1) of the mean monthly wind fields for January, April, July, and October for the Scripps trade wind set.

Over the Pacific Ocean, the most pronounced features are the northeast and southeast trade winds which lie north and south of the intertropical convergence zone (ITCZ) located 5° to 10° north of the equator. Each of these wind systems undergoes an annual cycle of equatorward migration and zonal expansion into the western oceans during hemispheric winter and spring followed by poleward migration and contraction into the central and eastern oceans during hemispheric summer and fall. In the southwest Pacific, there is a separate branch of the southeasterly trades. These too are most strongly developed during austral winter and spring when they extend westward into the Indian Ocean. Poleward of the trade wind belts, light winds mark the axis of the subtropical highs. In the January panel, midlatitude westerlies are seen along 30°N in the western and central Pacific.

As noted above, the convergence of the northeast and southeast trades is marked by the ITCZ which extends from off Central America to near the dateline. This feature is most clearly defined and farthest north (7° to 9°N) during boreal late summer and early fall; it is least developed and farthest south (3° to 5°N) during boreal spring. The convergence zone is marked by an increased meridional component relative to the zonal component but with generally lower mean wind speeds. Complementary regions of strongly divergent flow are found in the northeasterly flow north of the ITCZ, in the southeast trades along, and south of, the equator in the eastern and central ocean, and in the southeast Pacific.

A second convergence zone, the South Pacific convergence zone (SPCZ), marks the division between the eastern and western regions of southeasterly trades in the South Pacific. The SPCZ stands out most clearly as a wind field feature in austral summer when it is congruent with the equatorial doldrums and extends southeastward from off New Guinea to near Tahiti (20°S, 160°W). Although the region is less well defined at this time, convergence within the SPCZ is maximized in austral winter when the western South Pacific southeast trades intensify in association with the southwest monsoon.

Near Central America there is a second region of light winds with convergent meridional components. A monsoon circulation is observed in this region with westerly flow towards the Caribbean becoming established in summer and fall. Light easterly flow prevails during the other seasons. The equatorial western Pacific also experiences a pronounced annual cycle with strong
northeasterly winds prevailing in boreal winter which becomes very light (the doldrums) in boreal summer and fall.

The circulation over the northern Indian Ocean during boreal winter is dominated by divergent northeasterly flow associated with the Asian monsoon. In the southern Indian Ocean, moderate southeast trades prevail with a belt of westerlies just south of the equator separating the two regimes. Two convergence zones flank the poleward fringes of the equatorial westerlies.

The flow in the northern Indian Ocean reverses during boreal spring, becoming strong from the southwest in summer across the entire region. The reversal of flow is accompanied by a strengthening and expansion of the southeast trades in the southern ocean. By summer these reach to near the equator where easterly components give way to westerlies farther north. This transition zone is marked by a region of convergence. In fall the phase of the monsoon begins to reverse once again. The equatorial westerlies reappear as the southeast trades in the southern ocean retreat poleward and decline in strength.

The first two EOFs of the annual cycle shown in Fig. 2 summarize the scenario outlined above. The first
mode (77% of the total variance) reaches extremes near each solstice; this mode centers most dramatically on the Asian monsoon over the northern Indian Ocean and related circulation changes in the southern Indian and southwest Pacific oceans. To a smaller degree, this mode also expresses the variability of the trades in the northeast tropical Pacific, the migration and varying intensity of convergence in the ITCZ, and the shifts in position of the subtropical high pressure centers.

The second mode (13% of the variance), which lags the first by 3 months, centers on the variability of the northeast trade winds in the western Pacific and then swings in location and intensity of the ITCZ. This mode appears to reflect the fact that these more oceanic circulation regimes lag the solar cycle more than the continentally driven Asian monsoon, although some monsoon-associated features appear in this mode as well.

The third mode (4% of the total variance, not shown) depicts a 2-cycle per year signal whose spatial expression is dominated by the appearance of an anticyclonic (cyclonic) circulation over the north-central Indian Ocean and decreased (increased) trades north of Australia in boreal winter and summer (spring and fall).

b. The OSU GCM annual cycle of surface winds

The surface winds simulated by the OSU GCM for January, April, July, and October are shown in Fig. 3, along with corresponding residual maps (OSU–Scripps). Inspection of the observed and model wind fields shows that the OSU model reproduces the large-scale features of the tropical circulation such as the Asian monsoon, the northeast and southeast trades in the Pacific and the southeast trades in the Indian Ocean. Important aspects of the annual cycle are also duplicated. These include the changes in phase of the Asian monsoon, the summer (winter) hemisphere contraction (expansion) of the northeast and southeast trades in both oceans, and the meridional migration of the ITCZ.

Although these points demonstrate some success in duplicating the overall circulation features, the residual maps show that there are also some major differences between the simulated and observed fields. Over the North Pacific Ocean the most persistent problem is a residual westerly flow poleward of 20°N during boreal winter, spring and summer. This feature reflects the fact that the simulated North Pacific high is located well south of its observed position. This misplacement results in simulated midlatitude westerlies that reach as far south as 20°N during boreal winter and only light easterlies at that latitude during boreal summer. A similar, less pronounced displacement is evident in the southwest Pacific on the July and October maps where residual westerlies appear in the subtropics. This is in agreement with the findings of Schlesinger and Gates (1980; see their Fig. 3) who noted that the OSU GCM tends to locate subtropical highs equatorward of their observed position.

Another area of consistently large residuals is located in the eastern Pacific between about 20°N and 10°S where the model easterly flow is too strong by 5 m s⁻¹ or more (cf. Schlesinger and Gates 1980; their Fig. 4). This feature is most pronounced during boreal summer and fall (see the July and October maps in Fig. 4), where the model does not reproduce the monsoon-like
Fig. 3. Mean surface winds from the OSU GCM and residual winds (OSU-Scipps) for January, April, July, and October. Dashed lines mark the axes of regions of convergence.
weakening or reversal of the trades during summer and fall. A related point of interest is that the ITCZ is difficult to distinguish in the simulated wind field in comparison to the well-defined feature visible in the observed winds. Similarly, the region of divergent flow near the equator is not well portrayed. The residual maps clearly show a pattern of residual divergence along the ITCZ and convergence in the equatorial region. Maps of simulated and observed long-term monthly mean divergences (not shown) reveal that the simulated convergence in the ITCZ tends to be underestimated by a factor of 3–5, while the divergence south of the equator is typically underestimated by a factor of about 2. These underestimates of convergence do not necessarily relate to the fact that the model winds are extrapolated downward from the $\sigma = 0.75$ level (approximately 2000 m above the surface), since the convergence associated with large-scale tropical convection typically extends to twice that level (Williams and Gray 1972; Riehl 1979). It may also be true that the underestimated convergence may reflect the fact that the latitudinal resolution of the model grid is less than that of the observed data. Whatever the case, it is interesting to note that although the simulated convergence rates appear to be much too low, the OSU GCM does show precipitation rates about three times higher than those observed in the vicinity of the ITCZ (Schlesinger and Gates 1980). These points suggest that the relationships between persistent convective heating and the surface circulation along the ITCZ are not well characterized by the model.

A final point concerning the Pacific circulation is that the OSU model establishes a region of surface westerlies in the western equatorial Pacific during boreal summer and fall; these appear to be extensions of the westerly winds over the northern Indian Ocean associated with the Asian monsoon. While the observed flow clearly does show a trend for diminishing easterly flow in those seasons, prevailing westerlies do not appear.

Over the Indian Ocean the boreal winter and spring flow patterns typified by the January and April maps appear to be well simulated, although the northeast monsoon flow is much too strong in the January. In the southern Indian Ocean, the band of westerlies just south of the equator is well positioned, but too strong, in January. The observed northward migration of this feature in April does not appear in the simulation. The southeast trades in the southern Indian Ocean are well replicated by the GCM; however, the circulation associated with the Australian thermal low does not appear, probably because of the limited vertical resolution of the OSU model. Larger errors ($>10$ m s$^{-1}$) appear in the July and October maps during the southwest monsoon (see the July and October maps in Figs. 1 and 3). These discrepancies reflect misplaced, rather than spurious, circulations, and result from a southward displacement of the core of the monsoon westerlies located between 5° and 10°N in July and just north of the equator in October. These positions contrast with their climatological locations near 15° and 5°N, respectively, and result in large residual easterlies in the northwest Indian Ocean and residual westerlies along the equator. Over the southern Indian Ocean, the austral winter and spring circulation appears to be well replicated by the OSU GCM, although the tendency to locate the winter-hemisphere subtropical high equatorward of their observed position results in residual westerly winds south of 20°S in the western ocean.

The impacts of these points of agreement and disagreement are depicted in Figs. 4 and 5 which show maps of model skill and rms error in reproducing the annual cycle of the surface wind field. Regions where the mean of the monthly wind speeds is less than 2 m s$^{-1}$ are highlighted in Fig. 4 to indicate regions where model errors would have to be relatively small to achieve a high skill score. Figure 4 shows that skills ranging from 0.4 to 0.6 cover much of the tropical Pacific equatorward of 20° latitude in both hemispheres. Comparatively high skills are also found in the southern Indian Ocean north of 20°S. Thus the model does best by this measure in the major trade wind areas.

Poleward of about 20° latitude, skills decrease from about 0.4 to negative values along 30° latitude. This is particularly true in the North Pacific where the southward displacement of the subtropical high produced a large residual flow. Low skills are also found in the northeast tropical Pacific where the monsoon-like annual cycle was not well simulated by the model. A trough of comparatively low skill marks the ITCZ. The regions of low skill in the western equatorial Pacific and the northern Indian Ocean reflect the southward misplacement of the summer monsoon westerlies and their extension into the western Pacific in late summer. A region of low positive skill covers the central portion of the northern Indian Ocean where the model and observed monsoon circulations were in some agreement, particularly during winter and spring.

The features shown in the rms error map (Fig. 5) agree qualitatively with those in the map of model skill, indicating that the patterns shown in the latter do not reflect only the distribution of mean monthly wind speed of the observations. It can be seen that rms errors of 2 to 3 m s$^{-1}$ are found in the high skill areas associated with the trade wind regimes. Regions with values in excess of 4 m s$^{-1}$ cover the central and western North Pacific north of about 20°N latitude where the subtropical high was poorly simulated, and in the northeast tropical Pacific where the monsoon-like annual cycle was not reproduced by the model. Another region of large rms errors covers the western Indian Ocean, reflecting the model's misposition of features during the southwest monsoon.

The temporal aspects of model performance are summarized in Fig. 6 where the projections of the
model annual cycle onto the EOFs of the observed annual cycle are shown. It can be seen that the model performs relatively well between September and April (i.e., during the northeast monsoon); however, the model appears to lead the observed circulation by about one month. During the remainder of the year the agreement with the observed curve is poor, reflecting the model's mislocation of features related to the southwest monsoon.

c. The ECT21 GCM annual cycle of surface winds

The long-term monthly mean surface winds from the ECT21 GCM and the differences with the observed fields for January, April, July, and October are shown in Fig. 7. Comparison with Fig. 1 shows that like the OSU model, the ECT21 GCM reproduces many of the major features of the tropical circulation. It is also apparent that while the latter duplicates some of the more subtle details of the flow as well, there are also some important differences. These points of agreement and disparity are discussed below.

Referring to the residual map for January and April (Fig. 7), one can see that the major region of residual flow over the Pacific is a region of northeasterly residuals (up to 5 m s\(^{-1}\)) over the northwestern ocean. This pattern reflects the northward displacement of the axis of the subtropical high and associated easterlies, and the fact that the circulation is somewhat too strong.
This displacement also results in the residual westerly flow near the equator.

During this time period there are also consistent differences over the southwest Pacific where the western branch of the southeast trades and the SPCZ are weakly developed in the model climatology. The model's depiction of well-developed easterlies just south of the equator and light winds southeast of Australia contrasts with a reversed situation in the observations and results in an anticyclonic circulation in the residuals. This residual pattern is not so clearly defined on the April map but the easterly residuals across the SPCZ remain.

The structure of the northeast monsoon (January and April maps in Fig. 7) is well portrayed by the model, although the flow is too strong by 3 to 4 m s\(^{-1}\) in the prevailing direction in January, and inspection of the individual monthly mean maps for March (not shown) and April indicates that the onset of the southwesterly monsoon apparently occurs about 2 weeks too early in the simulated circulation. Over the southern Indian Ocean the model appears to position the features very well, although the flow is slightly too weak. The thermally driven circulation around western Australia is particularly well depicted. Near the equator, the January residuals are small but an organized area of easterly residuals appears in April when the southeasterly trades extend slightly north of their climatological position.
model gives mean annual convergence rates in the ITCZ that are approximately 25% of the observed values (Horel 1983).

Another point of similarity between the OSU and ECMT21 models is that both establish westerly winds during summer in the equatorial western Pacific, a region climatologically associated with the doldrums. The ECMT21 model amplifies this feature in October, developing a strong cyclonic circulation east of the Philippines. This situation results in northeasterly residuals to the north and large (5–7 m s⁻¹) westerly residuals farther south. It is interesting to note that when portrayed as departures from the overall annual mean, a similar, but weaker, circulation appears, as shown in the second wind field EOF (Fig. 2). These features qualitatively resemble the results obtained analytically by Gill (1980) and numerically by Webster (1981) for an off-equatorial heat source. Inspection of maps of the divergent component of the model and observed 200 mb wind components for Northern Hemisphere summer (Fischer 1987) show that simulated upper-level divergences are about 50% higher than those observed. This suggests that the model circulation is an exaggeration of the climatological tendency to develop a convectively driven thermal circulation in this region during boreal summer and fall.

In the Indian Ocean, easterly residuals of up to 5 m s⁻¹ are present along the equator with westerly residual flow through the Bay of Bengal. This pattern results from a 2° to 4° northward shift of the model circulation in this region. In October the westerly residuals decline, indicating that the change in phase of the monsoon circulation occurs at approximately the correct time in the model. The equatorial easterly residuals persist, now located about 20° west of their July position, as the northward displacement of the model circulation remains clear. Over the southern Indian Ocean, the model July and October circulation closely resembles the climatological pattern.

The skills for the ECMT21 model are mapped in Fig. 4. Comparison with the corresponding map for the OSU model shows that while the patterns are broadly similar with maximum skill in the major trade wind areas, the values are considerably higher for ECMT21 in most regions. The map shows broad regions of skills in excess of 0.6 covering the core regions of the northeast trade winds in the North Pacific and the southeast trades in the eastern and western South Pacific. Lower skills are found along the poleward limits of the study area. This is particularly true in the northwest Pacific, reflecting the large residual flow found there in boreal summer and fall. Low skills are also present in a region extending southeastward from near New Guinea in rough agreement with the location of the SPCZ. The ITCZ and equatorial divergence region are also marked by a trough in the skill field.

Over the northern Indian Ocean, skills are mostly positive and generally range from 0.4 to 0.6, although
Fig. 7. Mean surface winds from the ECT21 GCM and residual winds (ECT21-Scripps) for January, April, July, and October. Dashed lines mark the axes of regions of convergence.
an area of lower skill appears near the tip of the subcontinent. This is in agreement with the comparisons above, where it was found that many of the observed features of the monsoon circulation were duplicated by the model. Skills were uniformly negative in the equatorial Indian Ocean reflecting the residual easterly flow present in that region much of the year. High skills (generally above 0.7) are found across the southern Indian Ocean between 5° and 25°S, with lower values farther poleward.

On the whole, although the skill map for the ECT21 model is similar in character to that for the OSU model, the skill levels of the former are generally 15% to 30% higher in regions where both models show their highest skills. Major improvements in skill are seen in the eastern tropical North Pacific where the ITCZ and Central American monsoon are better depicted, along 30°N and 30°S where the subtropical highs are more correctly positioned, and over the northern Indian Ocean where the ECT21 model simulates the southwest monsoon more successfully than the OSU GCM.

Figure 5 shows the rms errors of the ECT21 model annual cycle. It can be seen that the errors are less than 2 m s⁻¹ over most of the Pacific and Indian oceans. The ITCZ and SPCZ both show up clearly as areas of relatively large errors, as does the region east of the Philippines and the northern Indian Ocean. Inspection of the EOFs in Fig. 2 shows that these are regions of comparatively large annual cycles.

The projection of the ECT21 model annual cycle onto the observed annual cycle EOFs are plotted in Fig. 6. As would be expected, the agreement with the climatological cycle is better than that shown by the OSU model, particularly in boreal summer and fall when the OSU model displaced major features relating to the southwest monsoon. The fact that the model curve is centered quite close to the origin indicates that there is little consistent bias in the model winds, as is suggested by the residual maps. This contrasts with results presented in Fischer (1987) indicating that the wind stresses from the ECT21 GCM are too low by a factor of about 2 when compared with the climatology of Han and Lee (1981); such a discrepancy is of major importance with respect to experiments in which the ECT21 model surface wind stress data are used to drive an ocean circulation model. This contradiction arises from the fact that the surface winds are calculated from the surface wind stress by a postprocessing procedure that apparently adjusts for the errors in the surface stresses data (von Storch, private communication). Thus, some of the success of the ECT21 model in replicating the observed surface winds rests not with the GCM itself, but rather with the postprocessing calculations.

The major region of discrepancy occurs in March and April where low amplitudes for the second mode result from the persistence of strong northeasterly flow off Asia and the too early expansion of southeasterly trades into the region of the SPCZ. The slight phase shifts in April and March reflect the fact that the onset of the simulated southwest monsoon takes place about 2 weeks earlier than in the climatology. A similar shift in August and September, but with the model lagging, apparently results from the model's persistence in maintaining strong southwesternly winds over the northern Indian Ocean, while the climatology shows slowly decreasing winds.

d. The NCAR GCM annual cycle of surface wind stress

This section describes the comparison between the NCAR model FORCED (prescribed observed SSTs) integration with the FSU Pacific wind stress set. Although the annual cycle of the CONTROL (prescribed long-term monthly mean SSTs) run differs in some interesting ways from that of the FORCED run, the differences are small from the perspective of comparison with climatology. Moreover, comparisons of the model results using actual, rather than average, SSTs would seem to give a more meaningful comparison with the observations.

Maps of the observed data for January, April, July, and October are presented in Fig. 8. These data show the same general features as the Scripps Indo-Pacific wind set described in the previous section, so they will not be discussed here except in comparison with the model fields.

The first two EOFs of the annual cycle of the FSU wind stress data are presented in Fig. 9. These EOFs are quite similar to those from the Indo-Pacific set and account for 69% and 12% of the variance, respectively. Although the first mode looks very much like that from the Scripps data (Fig. 2) over the Pacific, the percentage of variance on the first mode is somewhat less, probably because the monsoon-dominated Indian Ocean wind field is not included in this domain. Comparison of the amplitudes of the first EOF of the FSU wind stress data (Fig. 9) with those of the Scripps data (Fig. 6) shows that the first EOF of the FSU wind stress set reaches extreme values in February and August, rather than January and June as found in the corresponding EOF of the Scripps climatology. It seems likely this lag reflects the phasing of this mode with the extremes of the Asian monsoon which, because it is driven mostly by continental temperature changes, is nearly in phase with the solar cycle (Fu et al. 1983). In contrast, the Pacific wind field probably responds to a greater degree to the annual cycle of oceanic surface temperatures which, poleward of 15° latitude, increasingly lag the solar cycle proceeding eastward from Asia (Horel 1982).

More prominent differences with the Indo-Pacific EOFs appear in the second mode where the FSU EOF shows a smaller increase (decrease) in the northwest Pacific trade winds during boreal spring (fall) than the Scripps data. This difference may reflect the different
character of the datasets (monthly average stresses calculated from daily hand-analyzed wind fields rather than the overall monthly means of ship wind velocities), or it may result from differences in the character of the wind fields over the periods of record for the datasets.

The monthly mean and residual maps from the NCAR FORCED integration for January, April, July, and October are presented in Fig. 10. As was found with the other GCMs, it can be seen that although the major features of the circulation are reproduced by the model, there are differences in detail, placement and magnitude. In this case, the most obvious difference between the simulated and observed fields is that the stresses in the northeast and southeast trades are too strong by a factor of 1.5 to 3 (50–100 m² s⁻²). As discussed in section 2, this bias results in part from the fact that the winds used to calculate the surface wind...
stress are from about 100 m above the surface rather than from the commonly used observation height of 10 m. Adjustment for this effect results in a large reduction in the bias seen in the residual maps.

The largest errors in this respect occur during boreal winter as shown on the January map where major errors are depicted between 30° and 10°N in a region between 125°E and 160°W. These residuals result in part from the bias mentioned above, but also from the fact that the simulated North Pacific high is centered farther north and has a tighter pressure gradient than shown by the observations (Pitcher et al. 1983; their Fig. 6), thereby placing strong easterlies over the region of relatively light observed surface flow along the axis of the high.

Large easterly residuals of up to 100 m² s⁻² also appear between 15° and 20°N along the coast of North America where the model simulates strong easterlies over this area of climatologically light winds. A third region of residual flow that stands out on the January map is the belt of southeasterly residuals extending east-southeastward from New Guinea. Here the NCAR GCM depicts strong southeast trades across the doldrum are marking the SPCZ. Similar behavior can be noted in the ECT21 model (Fig. 7). Over the rest of the region the model flow is in general agreement with the observations except for the pervasive overestimation of the magnitude of the stress.

The April map shows better agreement between the model and the observations, particularly in the North Pacific where the major areas of January residuals are much reduced. It can be seen that the model tracks the southward movement of the ITCZ and reduced stresses in the extreme western North Pacific as the northeast trades begin to decline and contract. The residual pattern in the South Pacific is much the same as in January, although larger overestimates of the stresses are seen north of the South Pacific subtropical high.

During July a major region of residual easterly flow (60–80 m² s⁻²) appears along the equator off South America. In this case the model pushes the Southern Hemisphere southeast trades about 5° too far north into the eastern equatorial Pacific, a region of climatologically light winds. Similar behavior is apparent in the July 950 mb winds from the GFDL GCM (Lau 1985; see his Fig. 3). Experiments with the FSU wind-driven model of the tropical Pacific have shown that this spurious zonal flow on the equator produces major differences in the ocean model annual cycle as opposed to forcing with observed wind stress. In the northwest
NCAR FORCED GCM
- 100 m² s⁻²

Fig. 10. Mean surface wind stress from the NCAR GCM (FORCED) and residual wind stress (NCAR-FSU) for January, April, July, and October (top to bottom). Data are for 1961–72. Dashed lines mark the axes of regions of convergence.

Pacific, northeast residual flow reflects the extension of model easterlies too far west and south proceeding west of the dateline where the observations show the flow becoming light and southeasterly. This error results from the fact that the simulated North Pacific high is too strong and extends too far east in this season (Pitcher et al. 1983; their Fig. 6). As with the other GCMs, the region of the ITCZ shows up as an area where the flow is too strong and not sufficiently convergent; simulated stress convergence rates are about half that observed in July (and would be further reduced if the bias correction were made). Convergent residuals appear along the equator. This is particularly noticeable in the western Pacific where these residuals reflect the spuriously strong east-northeasterly flow in the simulation mentioned above. Over the rest of the ocean the errors appear to be dominated by the bias in magnitude.

The residual pattern during October is similar to that noted during July, except that the large residuals in the eastern equatorial Pacific are no longer present. This improvement results from the better simulation of the Central American monsoon than in July. Large residuals (50 m² s⁻²) remain over the SPCZ where well-developed southeasterly flow is simulated rather than the light winds observed. In the Northern Hemisphere, the model persists in placing the northeast trades and the ITCZ somewhat south of their climatological position, resulting in large easterly residuals.

The annual cycle skill for the NCAR FORCED integration surface wind stress is mapped in Fig. 4. For these calculations the magnitude of the stress has been lowered by a factor of 1.6 to give values more representative of conditions at 10 m. This correction reduces the rms error substantially. Note that because wind stress and velocity represent somewhat different
expressions of the wind field, the skills of the ECT21 and OSU GCMs should not be directly compared with those from the NCAR model. It can be seen that the pattern for the NCAR model is similar to that for the OSU and ECT21 GCMs. Like the other models, the NCAR GCM performs most skillfully in the major trade wind regions. Another region of relatively high skill lies off the Asian mainland where the model correctly simulates the circulation reversal associated with the monsoon. Areas of low skill are found near the poleward edges of the study area, in the ITCZ and SPCZ. As was true for the OSU GCM, the NCAR model performed poorly off the coast of North America and in the equatorial eastern Pacific.

The map of rms error for the NCAR FORCED integration appears in Fig. 5. The pattern is much like that for the skill, indicating that the features on the skill maps generally relate to model errors rather than to changes in the magnitude of the mean wind stress. In Fig. 5, the core areas of the trade winds show up clearly as regions of comparatively small errors, as does the area adjacent to the Asian coast. The tropical east Pacific, ITCZ and SPCZ show larger errors, as does the central North Pacific where, as noted earlier, the simulated subtropical high is not correctly positioned.

The projections of the FORCED run annual cycle onto the EOFs of the FSU stress set are plotted in Fig. 11. The projection of the CONTROL run climatology is also plotted and suggests the similarity between the two runs. No altitude correction has been made for these data. It can be seen that the major difference between the simulated and observed annual cycle is that the model curve is offset towards positive values of the second mode. This bias indicates that the residual map showing the difference between the annual mean of the GCM and observed circulations bears some resemblance to the pattern shown by the second EOF of the FSU wind stress. We prepared such a map (not shown) and, as expected, it is dominated by marked easterly residuals at nearly all locations and are largest in both hemispheres between 5° and 15° latitude. The model bias is reflected most by the second mode because that mode is dominated by generally positive weights on easterly departures from the overall annual mean in that region, while the first mode projects onto zonal anomalies of opposite sign in either hemisphere.

4. The NCAR GCM response to prescribed SST anomalies

One of the primary goals of our study was to evaluate the response of the NCAR GCM surface wind stress field to prescribed SST anomalies. To do this we compare the FORCED run wind stress patterns associated with ENSO SST anomalies with those found in the observations. This analysis, carried out using CCA and EOF analyses, shows that although the GCM response is similar in some respects to that of the real atmosphere, there are also some major differences. To indicate the effect of prescribed SST anomalies on the model circulation, the FORCED run is compared with the CONTROL integration in terms of interannual variability and rms error of the simulated wind stress data. These measures show large increases in interannual variability of the model wind stress equatorward of 20° latitude and overall improvement in model performance.

As an initial effort to characterize the dominant modes of interannual variability in the FSU wind stress field and to isolate the response of the observed surface circulations to anomalous SSTs, we calculated the EOFs of those data over the 1961–72 comparison period. The first EOF mode, which accounted for 7.8% of the total variance, represents variability in the strength of the circulation around the North Pacific high, with the largest weights being on the zonal flow in the region poleward of 20°N between 160°E and 140°W. Between 20°N and the equator, the weights become smaller but are still dominated by the zonal...
component; there is almost no signal in the Southern Hemisphere. The time series for this mode is comparatively noisy with a tendency for the largest (absolute) values to occur in boreal winter. This high frequency signal is superimposed on a steady linear trend for increasingly positive amplitudes on this mode, indicating a slow increase in the northeastnerly trades south of the North Pacific high during the 1961–72 period. The second EOF mode (5.7% of the total variance) is very nearly the austral reflection of the first mode, and focuses on variability in the subtropical Southern Hemisphere zonal flow, with the maximum weights tending to occur near 20°S, and extending further equatorward (to near 10°S) in the eastern South Pacific than in the western ocean (approximately 18°S). Near the equator and throughout the Northern Hemisphere portion of the domain, the weights are quite small. The time series of this mode has maximum variability during austral winter.

The third EOF of the FSU wind stress (Fig. 12, 4.4% of the total variance) data resembles patterns shown to be associated with ENSO in previously published composites and statistical analyses of tropical Pacific wind field variability (e.g., Rasmusson and Carpenter 1982; Graham et al. 1987a). These analyses show that the dominant feature accompanying warm SSTs in the central and eastern Pacific is the presence of westerly anomalies over the western and central equatorial ocean with suggestions of flanking cyclonic circulations. In Fig. 12 the characteristic jet of anomalous westerlies along the equator is clear, as are cyclonic circulations near 20°S, 140°W and, less distinctly, near 15°N, 160°W. The connection between this EOF mode and ENSO SST anomalies is corroborated by the agreement between the time series of eastern Pacific SST anomalies during the analysis period and the EOF amplitudes. These are plotted in Fig. 13a and b. The SST data are the mean of monthly anomalies for the region 8°N to 8°S and from 86° to 122°W on a 4° by 4° grid taken from the Comprehensive Oceanographic Atmospheric Data Set (COADS) (Slutz et al. 1985; Bourbour 1986). These are not the same SST data used to drive the GCM, but differences between the COADS and Oort (1983) datasets are probably small. The EOF amplitudes have been smoothed (cosine filter, half-width 3 months, 0.1 power point at 5 months) to remove high-frequency variability; the wind stress data were not smoothed prior to calculating the EOFs. The agreement between the two curves is clear (correlation = 0.67), with distinct peaks in each in 1963, 1965, 1969 and 1972 associated with warm water ENSO events ranging in strength from weak to strong. The fact that this EOF mode captures a small fraction of the total variance [and is probably not significant by the measure of Preisendorfer et al. (1981)], and yet is well correlated with eastern tropical Pacific SSTs, emphasizes the importance of relating EOFs to other physical processes when possible, rather than immediately discarding modes judged insignificant by purely statistical tests.

That the association between the third EOF and eastern tropical Pacific SSTs represents the dominant mode of covariability between the SST and surface wind stress fields is confirmed by canonical correlation analysis. For this analysis the SST data cover the region 16°N to 16°S from 122°E to 86°W. In this diagnostic application, CCA shows the dominant contemporaneous patterns linking the SST and observed wind stress fields. The first five EOFs of each field (50.7% and 25.5% of the variance, respectively) served as input to the canonical analysis. The results show that the first canonical mode centers on the variability represented by the first and fourth SST EOFs (not shown) and the third EOF mode of the FSU wind stress data, and the squared canonical correlation (0.64) indicates the association is a close one. The filtered time series for the

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**FSU STRESS INTERANNUAL EOF**

![Diagram](image)

**Fig. 12.** Third EOF of the FSU interannual wind stress anomalies from monthly means. Vectors are resultants of the \( u \) and \( v \) components of the EOF eigenvectors.
wind field expression of this canonical mode is plotted in Fig. 13d and clearly resembles that of the amplitudes of the third EOF mode identified with ENSO variability.

The spatial patterns of the first canonical mode for each field are combined in Fig. 14. As would be expected, the wind pattern depicted is quite similar to the third EOF of the observed wind stress data (the projection of this wind field canonical pattern onto the observed data accounts for 3.6% of the total variance, compared with 4.4% for the third EOF mode). The SST pattern is not one that is generally associated with the height of the warm phase of the ENSO cycle when the warm SST anomalies extend west from the eastern boundary; rather it is reminiscent of conditions somewhat later (around February–March of the year following the onset of El Niño) with the maximum anomalous SSTs detached from the coast of South America and the major center of warm SSTs located near 165°W. A region of cooler SSTs is indicated in the northwest Pacific. The canonical correlation, (i.e., the squared correlation of the temporal amplitudes of the patterns shown in Fig. 14) is 0.64, indicating a close agreement between the fields. As noted above, this SST pattern shown represents primarily the combination of two EOF modes. In order to clarify the interpretation of the canonical pattern, we give a brief discussion of the temporal and spatial aspects of these SST EOFs below.

The first SST EOF (34% of the total variance) resembles the pattern associated with the height of El Niño with the largest weights located along the equator in the eastern Pacific between approximately 95° and 140°W, in qualitative agreement with the composite pattern shown by Rasmusson and Carpenter (1982, their Fig. 20; cf., Weare et al. 1976) for the period August–October of the El Niño year. The fourth SST EOF (3% of the total variance) is similar in some respects to the pattern shown by Rasmusson and Carpenter (1982, their Fig. 23) for the period May–July of the year following an El Niño, with negative SST anomalies now at the eastern equatorial boundary, and residual positive SST anomalies in the central equatorial ocean. The main difference between Rasmusson and Carpenter’s figure and our EOF 4 is that in the former the warm pool is centered near the dateline on the equator, while in the latter the major equatorial anomaly is centered near 160°W. Analysis of the lagged correlations between the amplitudes of the first and fourth EOFs supports the idea that the former mode leads the latter. These correlations show relatively high values (0.6–0.7) at leads of 3–5 months, indicating that high values on the first mode between September and December precede high values on the fourth mode in January–April.

In assembling the canonical pattern seen in Fig. 14, the main effect of the fourth SST EOF (when combined with the first EOF mode) is to nullify the positive weights on SST anomalies in the extreme eastern equatorial Pacific and to increase the weights on positive SST anomalies in the central equatorial ocean. It is interesting to note that this results in maximum SST weights at the approximate location of the mean 27.5°C SST isotherm which vacillates between extreme positions of 170°W (February and March) to 150°W (May), but is located between 160° and 170°W from August through March. Sea surface temperatures of approximately this value appear to be a required condition for the organized deep convection in the tropical Pacific (Graham and Barnett 1987). A count of the number of occurrences when total SST exceeded 27.5°C and mean monthly SSTs were less than 27.5°C (i.e., months when interannual variability, not the annual cycle, pushed SSTs to above 27.5°C) produces a distribution (not shown) in many ways similar to that shown in Fig. 14. Thus, the importance of central equatorial Pacific SST anomalies indicated by the canonical analysis may reflect the fact that in this region ENSO-related SST anomalies, though frequently smaller than those further east, can result in a change...
from an environment of little precipitation to one associated with organized convection. The consequent diabatic heating of the atmosphere act in concert with the sensible heat anomalies produced by the SSTs, thereby enhancing the effects on the circulation (cf. Fennessy and Shukla 1988) above those produced by sensible heating alone.

It can also be seen in Fig. 14 that the maximum zonal wind anomalies along the equator are displaced about 10° east of the major thermal anomaly in qualitative agreement with aspects of the analytic results of Gill (1980) and numerical experiments (e.g., Webster 1981; Lau 1985). The anomalous flow along the equator between 160°E and 160°W appears to be down the thermal gradient, although westerlies do cross the region of maximum SSTs dying out about 10° farther east. Proceeding off the equator, the flow becomes directed roughly along the anomalous thermal gradient in both hemispheres by 8° to 12° latitude, suggesting geostrophic flow around a warm-cored low associated with the warm SSTs and related convective heating. Enhanced easterly flow in the eastern North Pacific appears to be associated with the anomalously warm SSTs located at 5°N, 120°W. In order to ensure that the relationships described above have been consistent through time, a similar canonical analysis of Indo-Pacific COADS SSTs and the Scripps wind data for the period 1950–84 was carried out. The results were essentially identical to those shown in Fig. 14.

An identical set of EOF and canonical analyses was carried out using the NCAR FORCEN integration wind stress data. The first three EOFs do not appear to be associated with ENSO variability and are described briefly here. Like the first EOF of the observations, the first EOF of the FORCEN wind stress data (9.5% of the total variance) is dominated by variability in the strength of the Northern Hemisphere trade winds, in this case between 5° and 20°N. The time series for this mode is characterized by high frequency variability with time scales of 2 to 4 months superimposed on a longer period signal, suggesting a decrease in the simulated easterlies from 1961–66 followed by an increase from 1967–72. The second EOF of the FORCEN wind stress data (7.5% of the variance) focuses on variability in the zonal circulation poleward of 20°N between 140°E and 150°W. Like the first mode, the amplitudes contain much high-frequency variability with time scales of 2–4 months. The third FORCEN wind stress EOF (6.4% of the variance) centers on variability in the zonal flow across the South Pacific between 15°S and 30°S. A second major area of variability is the suggestion of a cyclonic (anticyclonic) circulation near 20°N, 160°E, when anomalous easterlies (westerlies) are present in the Southern Hemisphere. The circulation around this cyclonic circulation shows anomalous west-southwesterly flow from the equatorial western Pacific to near 20°N and the dateline. The temporal amplitudes for this mode show less high frequency variability than the first two modes, with a tendency for episodic behavior with periods of 1 to 2 yr.

For the FORCEN data the fourth EOF (Fig. 15; 5% of the variance) appears to be most strongly associated with ENSO SST variability. The association of this mode with ENSO SST anomalies is suggested by the time series of smoothed EOF amplitudes plotted in Fig. 13. The agreement with both the SST index (correlation = 0.57) and with the observed EOF mode (correlation = 0.52) is apparent, although there are some phase shifts. It can be seen that this pattern shows several features in common with the EOF of the observed winds (Fig. 12). These include westerlies along...
the equator in the central equatorial ocean; a well-developed cyclonic circulation near 20°S, 160°W; indications of anticyclonic circulation north of New Guinea; and anomalous westerlies in the eastern North Pacific along 30°N. There are also important differences between the patterns. For example, the EOF of the model wind stress shows a region of strong anomalous easterlies near 10°N, 100°W that is not present in the EOF of the observed data. Also, the model EOF does not show the increased Northern Hemisphere trade winds in the central ocean; however, some increase is seen in the Southern Hemisphere. The simulated equatorial westerly anomalies extend about 10° north and south of the equator, much farther than the roughly 5° shown for the observed stresses, and are located slightly farther east.

It is of interest to examine the results of similar experiments with other atmospheric models to see if these show similar discrepancies. For example, the simple 1½ layer models of Gill (1980) and Zebiak (1986) both produce large areas of spurious easterlies in the eastern tropical Pacific, a result that Zebiak (1986) attributed to eastward propagating atmospheric Kelvin wave activity excited by convective activity further west. Similar patterns appear in results from the more complex two-level models of Webster (1981) and in some of the results of Schopf and Suarez (1988), although the latter is a “dry” model, without an explicit parameterization of convective heating. Going to much more complex models, inspection of the results from the GFDL GCM (Lau 1985) and the Goddard Laboratory for Atmospheric Sciences (GLAS) GCM (Shukla and Wallace 1983; Fennessy et al. 1985) suggest the following conclusions. Both of these models appear to produce equatorial westerly wind regions that extend farther from the equator than observed in the real atmosphere. The GFDL model does show some enhancement of the off-equatorial easterlies in qualitative agreement with the observations; however, it also mistakenly simulates easterlies in the eastern tropical Pacific as does the NCAR model. In contrast, the GLAS GCM shows neither the desired central Pacific trade wind enhancement nor the spurious eastern Pacific easterly anomalies. In addition, many of the models, including the NCAR GCM, show smaller meridional components than are seen in the observed wind fields. This problem is probably related to the problems with the model boundary layers and may also reflect the limited spatial resolution.

One of the important parameters affecting tropical SSTs in the central and eastern equatorial ocean is the thickness of the warm upper layer (others include advection and surface fluxes). Changes in upper layer thickness (ULT) are associated with wind-forced baroclinic wave activity which act to change the depth of the warm upper layer of the ocean. The discrepancies between the FORCED and observed wind stress responses the ENSO SST anomalies can be discussed with respect to the generation of such waves. Near the equator the dominant forcing parameter for baroclinic waves is the zonal wind stress, changes in which generate equatorial Kelvin waves. As noted above, in some ways the NCAR GCM and other models do produce a qualitatively correct response with respect to equatorial zonal wind stress; i.e., warm (cool) SST anomalies in the eastern and central equatorial ocean produce anomalous westerly (easterly) wind stress in the central equatorial ocean. This point is highlighted in Fig. 16 which shows the smoothed time series of zonal wind anomalies (from the long-term monthly means) of the FSU and FORCED run data (adjusted to 10 m height) from the region 3°N to 3°S and 155°E to 165°W. Although the coherence between the curves is apparent, the amplitude of the GCM stress anomalies is less than that observed. In fact, the variability of zonal stress is less in the GCM than in the observed
data over most of the Pacific, as shown in the map of ratios of interannual standard deviations of zonal stress shown in Fig. 17. It can be seen that along the equator the model variability is generally 50% to 60% of the observed. Elsewhere, values range from 40% to 80% over most of the Pacific, although model variability is larger by a factor of 2 in the eastern equatorial and North Pacific where the GCM produces spurious easterlies in association with ENSO SST anomalies. Experiments with the FSU model of the tropical Pacific show that the qualitative agreement between the FORCED and observed central equatorial Pacific zonal wind stress results in some similarity in the remotely forced changes in eastern tropical Pacific pycnocline depth for simulations forced by the two datasets. Because the interannual variability in SST in that region is closely associated with changes in pycnocline depth (e.g., Busalacchi et al. 1983; Hickey 1975; Graham and White 1988), this is an encouraging result. However, other deficiencies in GCM performance would limit the degree to which actual SST anomalies could be specified if an SST parameterization was included in these simulations.

Away from the equator, wind-driven vertical motion in the ocean at long time scales is related to the Ekman pumping (which in turn is largely a function of the curl of the wind stress), i.e.,

$$\frac{\partial h}{\partial t} = \nabla \times \frac{\tau}{f},$$

where $\partial h/\partial t$ is the wind-driven vertical motion, $\tau$ the wind stress vector and $f$ the Coriolis parameter. In reference to the comparison of the GCM and observed wind field response to SST anomalies, it appears that the model fields do not produce as much curl in the
off-equatorial regions as found in the observations. This idea is supported by the pair of curves (labeled c and d) in Fig. 16 which show the smoothed time series of anomalies of $\nabla \times (\tau/f)$ from the FORCED run (adjusted to 10 m height) and from the observations for the region bounded by 9° to 13°S and 155°E to 175°W. Here it can be seen that although there is some agreement between the curves, the amplitude of forcing produced by the GCM is muted in comparison with the observations. This lower variability in $\nabla \times (\tau/f)$ is evident over much of the ocean (Fig. 18) which shows ratios of interannual standard deviations of $\nabla \times (\tau/f)$ between the FORCED GCM and FSU observations.

It can be seen that the variability in the model is less than half that in the observations over most of the tropical Pacific south of 20°N. In view of the evidence that Rossby wave activity plays a role in timing the phase changes between warm and cool water regimes of the ENSO cycle (e.g., White et al. 1987; Pazan et al. 1986; Graham and White 1988; Schopf and Suarez 1988), these discrepancies would have important effects on the behavior of coupled models of the tropical Pacific.

Application of CCA to the FORCED run winds and observed SST field shows that the first canonical mode centers on the simulated wind field EOF mode dis-
ncussed above and on the first EOF of the SST data. The squared canonical correlation of 0.26 for this mode indicates that the response of the GCM to prescribed SSTs is not as strong as in the real atmosphere. This idea is in agreement with the fact that the correlation between the simulated wind stress EOF and the eastern Pacific SST index was lower than for the observed wind stress EOF. The filtered time series for this canonical variable is plotted in Fig. 13 and, as expected, resembles the time series of the simulated wind stress EOF amplitudes.

The spatial expressions of this canonical mode for the SST and wind stress fields are depicted together in Fig. 19. It can be seen that the wind field pattern is nearly identical to the EOF shown in Fig. 15. The SST pattern shows the major anomalies in the eastern and east-central equatorial ocean still contiguous with the coast of South America, with cool SSTs in the western ocean, most prominently south of the equator. This distribution is different from that depicted for the observed wind canonical mode in Fig. 14 and suggests conditions typically found a few months earlier (extended EOFs of SSTs, not shown here, suggest a time around November–December) in an ENSO event. The lack of maximum weights on central equatorial Pacific SSTs seen for the observed case (Fig. 14) and the larger weights in the eastern ocean suggest that the GCM circulation is responding more to sensible heat anomalies and less to the convective heating anomalies associated with SSTs greater than 27.5°–28°C than the real atmosphere. Although the simulated circulation and SSTs appear to be qualitatively similar to the observations near the equator, with the flow being down the thermal gradient, the off-equatorial flow appears to be following the thermal gradient in the opposite sense from that seen in the observations (and from what would be expected for a thermally forced circulation), suggesting that the response of the simulated off-equatorial pressure field to the equatorial SST anomalies is incorrect. It is interesting to note that if the GCM wind stress pattern in the Northern Hemisphere was compressed towards the equator by approximately 10°, the agreement between the model and observed fields distribution would be somewhat better. The differences in the canonical SST and wind stress patterns suggest problems with the vertical and horizontal distributions of diabatic heating in the model.

The suggestion that the pattern representing the maximum GCM variability in conjunction with ENSO SST anomalies occurs somewhat earlier than in the real atmosphere is supported by lagged correlations between the time series shown in Fig. 13. These are given in Table 1 which presents lagged correlations between the amplitudes of the observed and model wind stress EOFs, between each EOF and the eastern Pacific SST index, and between the time series of the wind field canonical variables. Although not entirely consistent, these data suggest that the maximum response in the observed wind field occurs about 2 months after SSTs peak in the index region, or about 1 month after the maximum response in the simulated wind field. Since these statistics are based on only four events, and the behavior seen in the various plots in Fig. 13 is somewhat ambiguous with respect to leads and lags, this interpretation is tentative.

It is also of interest to examine the response of the GCM to anomalous SSTs by comparing the FORCED and CONTROL run results. To do this we have pre-
pared maps of 1) the ratios of standard deviations of the zonal wind stress, and 2) the ratio of rms error (model versus observation) between the two simulations. These statistics were calculated from the departures from the overall monthly means over the 144 months that the FORCED GCM simulation and the FSU data overlap. The first of these is presented in Fig. 20 and shows increases in interannual variability of more than 20% over a large portion of the Pacific equatorward of 20° latitude in the FORCED run. The areas of greatest increase in variability agree qualitatively with the areas of high variability indicated by the EOF of the simulated wind stress data shown in Fig. 15, with centers on the equator near the dateline and in the tropical northeastern Pacific. Surprisingly, a large region of reduced variability appear north of 20°N. Why this should be the case is unclear.

A map of the ratio of rms error between the GCM and observed winds appears in Fig. 21. It can be seen that the model response to prescribed SST anomalies has resulted in improvements in model performance of more than 5% in some areas, although in other areas it has been reduced. Major areas of improvement appear in the northeast Pacific, in the equatorial central Pacific and in the south central ocean. For the most part, these do not correspond to the regions where the largest increases in model variability due to SST anomalies were found. Indeed, the region of reduced rms error north of 20°N is a region where model variability was reduced in the FORCED integration. Overall, 81 (of 407) locations showed decreases of 5% or more in rms error while 39 showed corresponding increases. The major regions where rms error were larger in the FORCED run than in the CONTROL run correspond well with areas where the GCM response to SST anomalies was not like that shown by the observations. In particular, the areas of increased error along about 10° latitude agree with the region where the flow in the FORCED EOF was westerly, while the EOF of the observed winds showed easterly anomalies. Similarly, the region off Mexico and Central America corresponds to the area where the FORCED run showed spurious strong northeasterly winds during warm ENSO events.

5. Concluding remarks

The correct simulation of the wind stress field and its response to changes in SSTs is a key factor governing

![Fig. 20. Ratio of standard deviation of zonal component of pseudostress for the NCAR FORCED to NCAR CONTROL run. Areas with ratios higher than 1.2 are stippled; contour interval is 0.1, 1.0 contour is heavy.](image-url)
the usefulness of coupled tropical ocean–atmosphere models. In this paper we have presented the results of two analyses of the accuracy of GCM-produced wind fields over the tropical oceans. Because the periods of comparison for these analyses are relatively short, we have not attempted to place statistical confidence limits on our findings so they should be regarded as tentative. However, for the most part we point out only large-scale features and, where possible, note relationships with other data and physical processes to make these results more useful.

The first analysis concerned the annual cycles of the tropical wind fields simulated by the OSU, ECT21, and NCAR GCMs. The results show that the 16-layer ECT21 model appeared to perform somewhat better than the 9-layer NCAR GCM and considerably better than the 2-layer OSU model. However, postprocessing of the ECT21 data may have contributed to its better results. Although there are substantial differences in the performance of the models, there are also clear similarities in their behavior. Each did best in major trade wind regions and somewhat poorly in convectively active areas with light winds. These findings suggest that the formulations governing the interactions between persistent convection and the circulation may limit model performance. Possible reasons for such behavior include limited model resolution, problems with boundary-layer parameterizations, and incorrect specification of the vertical distribution of diabatic heating associated with deep convection.

The response of the NCAR GCM to prescribed SST anomalies, in terms of tropical Pacific wind stress, was examined in the second analysis. It was found that the model response to ENSO SST anomalies was clear and in some respects qualitatively resembled that of the real atmosphere. However, there were important discrepancies including 1) the latitudinal extent of the simulated equatorial westerly anomalies associated with ENSO events was roughly twice that seen in the observed wind, 2) differences in the location of the Northern Hemisphere off-equatorial easterly anomalies, and 3) the coupling between the simulated wind field and SST variability is less than half as strong as it appears to be in the real atmosphere.

A comparison of the NCAR FORCED and CONTROL runs showed that the response to SST anomalies produced large (20% to 30%) increases in interannual variability over much of the Pacific equatorward of 20° latitude. Surprisingly, a large area of reduced variability appeared north of that latitude in the Northern Hemisphere. In terms of rms error, the FORCED run showed some improvement over the CONTROL integration with twice as many locations showing reductions in error of 5% or more than showed increases of the same amount. Where large increases in rms error occurred in the FORCED run, they can generally be attributed to the differences between the model and real atmosphere responses to ENSO SST anomalies noted above.

Taken together, these findings suggest that although GCMs are capable of reproducing correctly many features of the tropical surface wind fields, major discrepancies remain with respect to both the annual cycle and to the response to the anomalous SST patterns associated with ENSO. In part, at least, these appear to be related to interactions between large-scale convection and the circulation. From the perspective of wind-forced baroclinic waves in coupled tropical ocean–atmosphere models, our results (and others) show that with respect to zonal wind stress anomalies in the equatorial central Pacific, the response to ENSO SST anomalies is simulated relatively well. Experiments
with the FSU model of the tropical Pacific (to be described in a later paper) show that the FORCERD case wind stress data does produce baroclinic Kelvin wave activity at the eastern boundary at approximately the correct times. However, because the GCM’s response to ENSO SST anomalies is too weak, the amplitude of these anomalies is lower than that observed. Away from the equator over the tropical Pacific, the GCM did not replicate the observed response in terms of Ekman pumping. To the degree that wind-forced baroclinic Rossby wave activity contributes to the later changes in eastern equatorial Pacific upper layer thickness (e.g., White et al. 1987; Schopf and Suarez 1988; Graham and White 1988), the poorly simulated response in these regions would degrade coupled model performance. In view of these discrepancies, the differences between the observed and simulated annual cycles, and the fact that other processes affecting SSTs must also be correctly simulated, it is not clear that a fully coupled ocean–atmosphere model using the version of the NCAR GCM referred to in this study would reproduce ENSO-like oscillations in coupled setting. Nevertheless, it is of value to examine the sensitivity of a simple ocean model to the differences between the observed and model wind stress fields we have noted. A manuscript describing such an experiment is currently in preparation.

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