Diagnostic Studies of Pacific Surface Winds

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ABSTRACT

Analyses are performed using surface winds derived from the FSU Pacific pseudo-stress fields. Vorticity budget calculations reveal the relative contributions of various terms, and allow simplification of the full momentum equations. Based on the simplified equations, a procedure is developed for estimating surface pressure, and then adjusting the winds. Finally, the winds and pressure are used to infer boundary layer and upper level forcing, within the context of particular model formulations. The results reveal deficiencies in these models, and suggest a convective heating structure different from what is often assumed.

1. Introduction

The importance of surface winds has long been appreciated in the context of the general circulation of the world's oceans, but recently has been underscored by developments in the theory and modeling of tropical ocean-atmosphere interaction. As the primary agent of communication between atmosphere and ocean, surface winds must be considered an essential component of all long-term climate variability. Yet, until recently, the surface wind field over the oceans has received relatively little attention. Fortunately, recent efforts have resulted in substantial improvements in operational surface wind analyses (Trenberth and Olsen,1988), and also in the reworking of historical ship winds using more complete data sets and new analysis procedures. The problems are still many: evidence of artificial trends (Ramage,1987; Wright,1988; Posmentier et al.,1989; Cardone et al.,1989), and large data gaps in both space and time, even in current analyses. Despite such difficulties, many of these products have been used to drive ocean model simulations with notable success (e.g.,Philander and Seigel,1985; Seager,1989; Harrison et al,1989), implying that they are indeed capturing salient features of the real variability.

We might then expect that the same products, even with their limitations, could be useful in evaluating different conceptual scenarios represented in recent atmospheric modeling studies. We could ask, for example, whether some of the simplifications of these models are justified in the context of "observed" surface winds. If so, then perhaps something can be learned about systematic model errors in terms of parameterized forcing functions. These were the motivations for the present study. The data source is the FSU tropical Pacific analysis product (Goldenberg and O'Brien, 1981), which spans the period from 1961 to present. The choice is arbitrary; other products could (and ultimately should) be used for the same purpose. Since most of the results appear to depend on very general properties of the wind fields, and not details, it seems likely that similar findings would be obtained with other products.

The study consists of three parts. First, a vorticity budget calculation is done, as a means to evaluate the relative importance of various terms, especially those that are neglected in most simple models. This calculation also affords the opportunity to make an indirect assessment of the

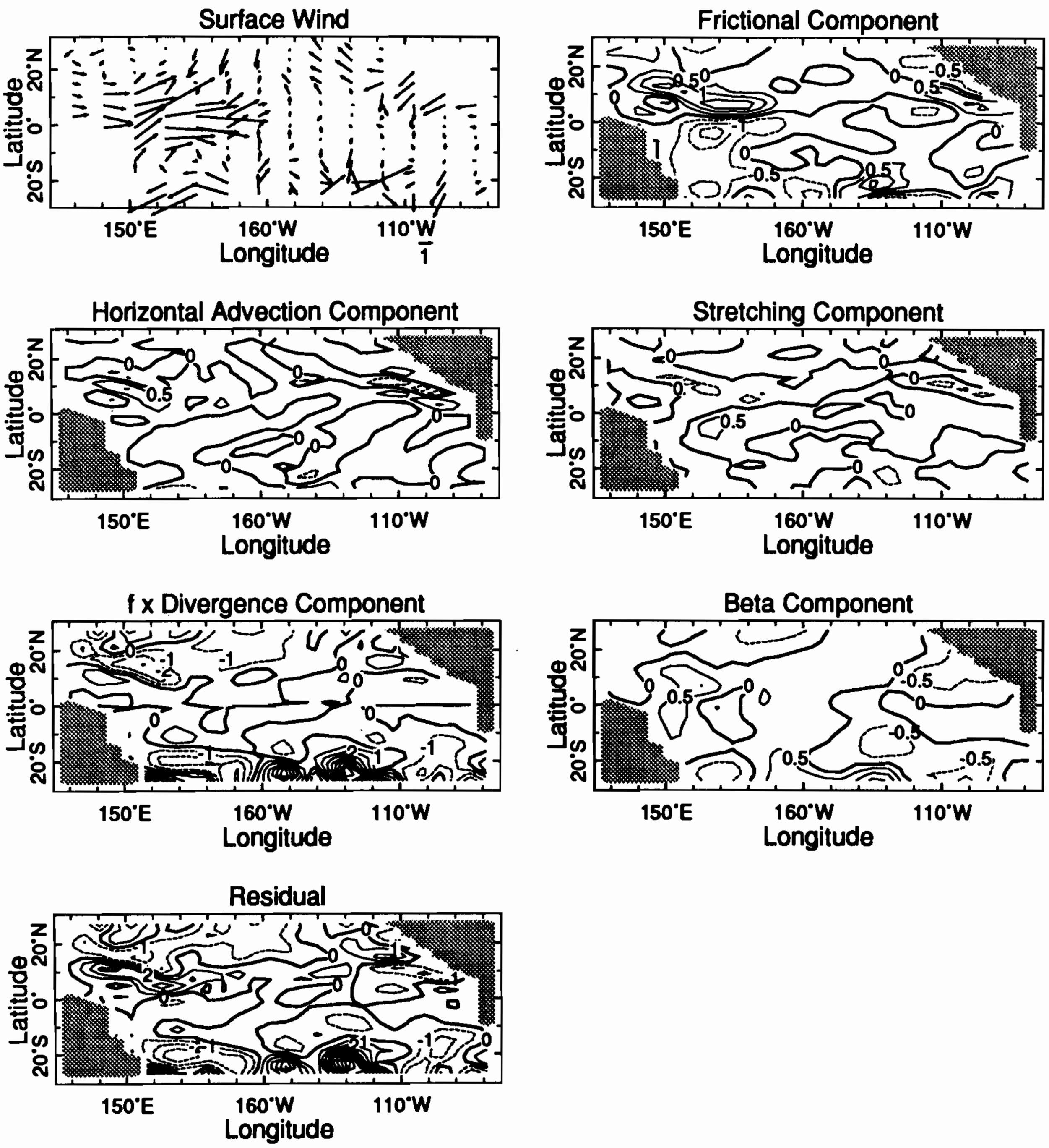

Fig. 1. Surface wind anomalies, and the component terms of the vorticity budget (Eq. 1), for July 1982.

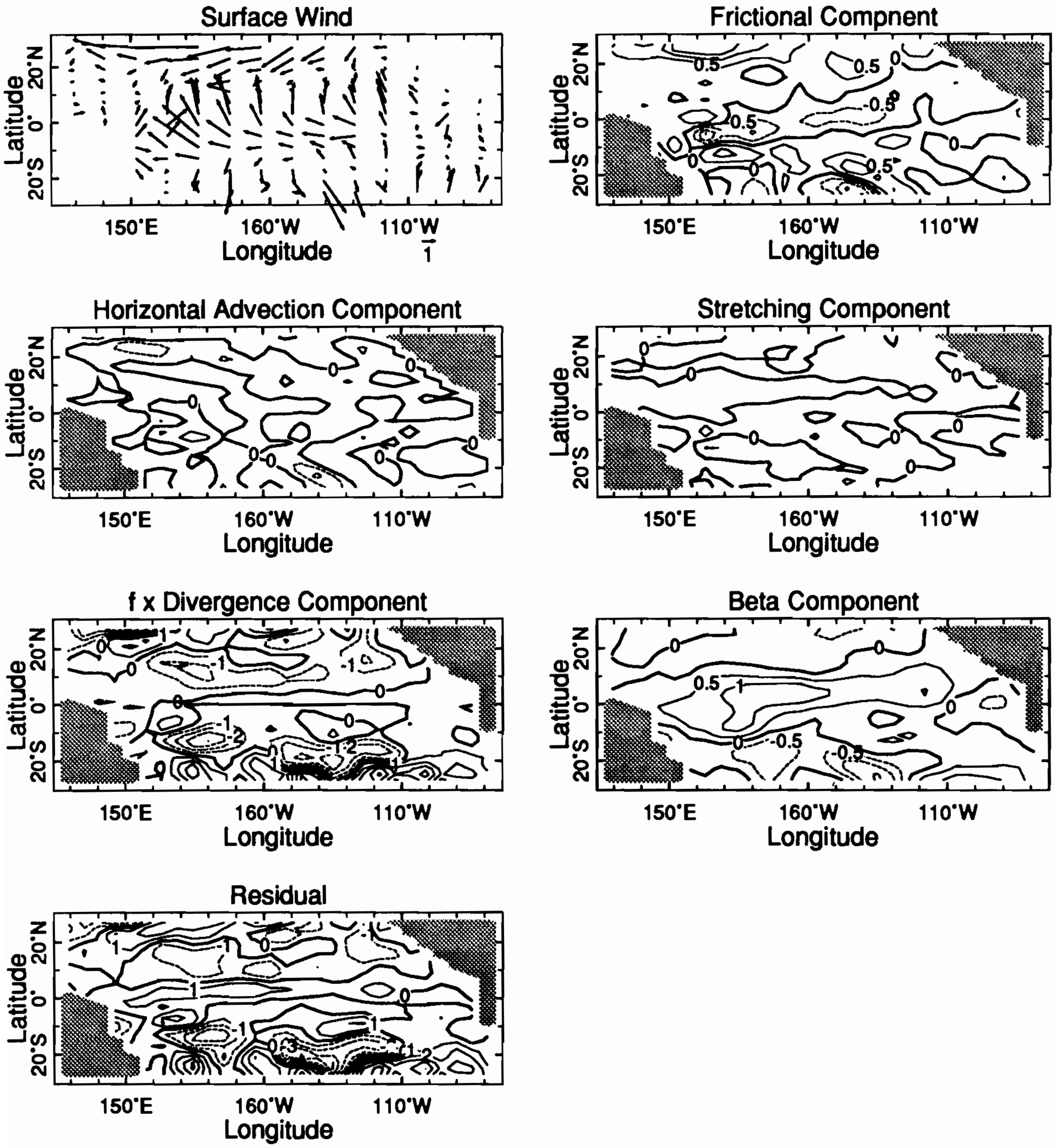

Fig. 2. As in Fig. 1, except for March 1985.

data quality, as will be shown below. Secondly, we present an algorithm for estimating surface pressure fields from the winds, and make some comparisons with observed surface pressure. From the pressure field, it is posible to calculate a modified, and dynamically consistent wind field. This procedure selectively filters out certain "noisy" (and unrealistic) features in the divergent component of the winds, while reproducing most other characteristics closely. Thirdly, the combined wind and pressure fields are used to test particular models, and to infer the structure of forcing functions that would give the "correct" surface winds. The study focuses primarily on (monthly) anomalies; ENSO variability is thus the dominant signal being analyzed.

2. Vorticity Budget

We start with the vorticity equation in the following form:

$$u\zeta_x + v\zeta_y + \zeta\delta + f\,\delta + \beta v + F = R, \tag{1}$$

where ζ and δ are the vorticity and divergence, respectively, f is Coriolis parameter, β its latitudinal gradient, F is the curl of the frictional force, and R a residual. In spherical coordinates, $x{=}a\cos(\theta)\phi$ and $y{=}a\,\theta$, where a is the earth's radius and θ and ϕ measure latitude and longitude, respectively. The residual R contains the effects of time dependence and transients, vertical advection, the nonlinear curvature terms, and errors. For the monthly mean fields being analyzed, most of these can safely be regarded as small. The possible exceptions are the error terms, and transient forcing. Note that errors can arise from the parameterization of F, as well as the data itself.

Two forms were investigated for the frictional stresses: a linear and a quadratic drag. The former is commonly used, though the latter is formally better justified. Near the surface, such parameterizations should be appropriate, whereas, if dealing with depth-averaged winds over a deeper layer, the effects of vertical structure within and above the boundary layer would become important. In fact, the results proved to be insensitive to which form was used, so we will describe here the results using the linear form. For this case, $F{=}\epsilon\zeta$. Unless stated otherwise, the dissipation time (ϵ^{-1}) is taken to be 1 day.

The monthly FSU data are produced on a 2° latitude by 2° longitude grid between 29°S and 29°N, 124°E and 70°W. These fields were first smoothed with a 1-2-1 filter in time and space, interpolated to a coarser longitudinal grid (5.625° intervals, as used by the models to be described below), and then the terms in (1) evaluated, for each month of the data set. (As a check, some of the calculations were repeated with unsmoothed fields; none of the results changed significantly.) Representative cases are shown in Figs. 1-2.

July 1982 (Fig. 1) represents the beginning phase of the major warm event of 1982-1983. As is typical of antecedent periods, strong westerly anomalies appear in the western Pacific, from the equator northward. There is little signal in the eastern Pacific, and a rather incoherent pattern of anomalies in the southern subtropics. The frictional stress curl (which in the linear case is proportional to the vorticity anomaly) features a large dipole pattern in the west Pacific in the vicinity of the westerlies, and smaller, less coherent pattern elsewhere. In comparison, the zonal and meridional advection terms (displayed together in Fig. 1c) are small, and rather "noisy". The same is true for the $\zeta\delta$, or stretching term. On the other hand, the $f\,\delta$ term is very large, especially in the subtropical latitudes, where already sizeable divergence anomalies are amplified most. In the equatorial region the field is more coherent, and roughly in quadrature with the vorticity field, as might be expected. The βv term is slightly smaller in magnitude than the frictional term, but larger than the nonlinear terms; moreover, its spatial structure is unique in being coherent over large meridional regions spanning the equator. Unfortunately, the residual term is as large or larger than any other, but inspection reveals that its structure is almost identical to the $f\,\delta$ term. A reasonable hypothesis is that the divergence field contains spurious features that cannot be balanced by the remaining terms.

March 1985 represents approximately the opposite phase of ENSO; there were anomalously cold sea surface temperatures and strong equatorial easterlies (Fig. 2). Southerly anomalies across the normal ITCZ position indicate its northward displacement. Despite the very different conditions, the same conclusions follow with respect to the relative contributions of vorticity budget terms. The nonlinear terms are small, and the $f\,\delta$ term is largely unbalanced.

In order to make a more general assessment of the vorticity budget, the root-mean-square (RMS) of the various terms were computed, based on all spatial points and all months between 1970 and 1988. Once again, the nonlinear terms are seen to be the smallest contributions, and the $f\,\delta$ and residual terms are more than twice as large as the next largest terms (Table 1). As a test, the residual was recomputed with the nonlinear terms removed; this resulted in a *reduction* of the residual. Our interpretation is that these terms are incorrectly represented due to the data quality combined with the amplifying effect of second order differentiation. In any case, it is clear that they play a minor role in the budget.

The size of the residuals in these calculations is alarming. The fact that the structure of the residual field is very similar to that of $f\,\delta$ suggests that the source of the imbalance is data errors, and not some other term excluded from the budget. This is reinforced by additional calculations with the $f\,\delta$ term artificially reduced. In all cases, this resulted in a proportional reduction in the residual, as would be expected if this term were largely unbalanced by any other. We note that similar calculations were carried out with other terms suppressed: the residuals either increased or were unchanged. Also, the frictional parameter ε was varied over a considerable range. Probably due to the overwhelming imbalances from other sources, the sensitivity of the residual to this parameter was rather low, but the "best" values were in the range of $1 day^{-1}$. Replacing the linear frictional form by a quadratic one produced no detectable difference.

3. Pressure Field Estimation

The results so far indicate that the nonlinear terms of the surface momentum balance are of secondary importance. On the other hand, the remaining linear terms do not balance, even approximately, based on the wind data. Is the problem just data errors, or something more? One way to attack this question is to assimilate the data into a model based on the linear equations, and determine whether the differences between the two are within expected limits of uncertainty. We did this by using the winds to estimate the surface pressure field, and then reconstructing the winds from the pressure in accordance with the governing equations. This approach, by definition, yields a dynamically consistent combined wind-pressure set.

The linear momentum equations may be written as follows:

$$\varepsilon u - f v = -p_x/\rho \tag{2a}$$

$$\varepsilon v + f u = -p_y/\rho, \tag{2b}$$

where p is the surface pressure and ρ the density. These equations imply a particular relation between the wind components which does not hold in the data, due to errors and neglected terms. Thus the pressure field is not uniquely determined. Our approach was to compute many estimates of the pressure field by integrating (2a-b) along different paths, and average them. The individual estimates were calculated as follows: for a particular latitude of the grid (the reference latitude), (2a) was integrated along the entire longitude range. Then, for each longitude, (2b) was integrated from the reference latitude northward and southward to the limits of the grid. By choosing all possible reference latitudes, 30 separate versions of the pressure field were produced. The final pressure field was then calculated as a weighted average of these, with the weighting proportional to $1/\sin\theta$. This was important because of the general increase of pressure

Data source	$\varepsilon\zeta$	$u\zeta_x+v\zeta_y$	$\zeta\delta$	$f\delta$	βv	R
Original winds	.50	.30	.15	1.28	.49	1.39
Adjusted winds	.47	.13	.06	.49	.39	.05

Table 1. Root-mean-square contributions of vorticity budget terms based on original and adjusted winds.

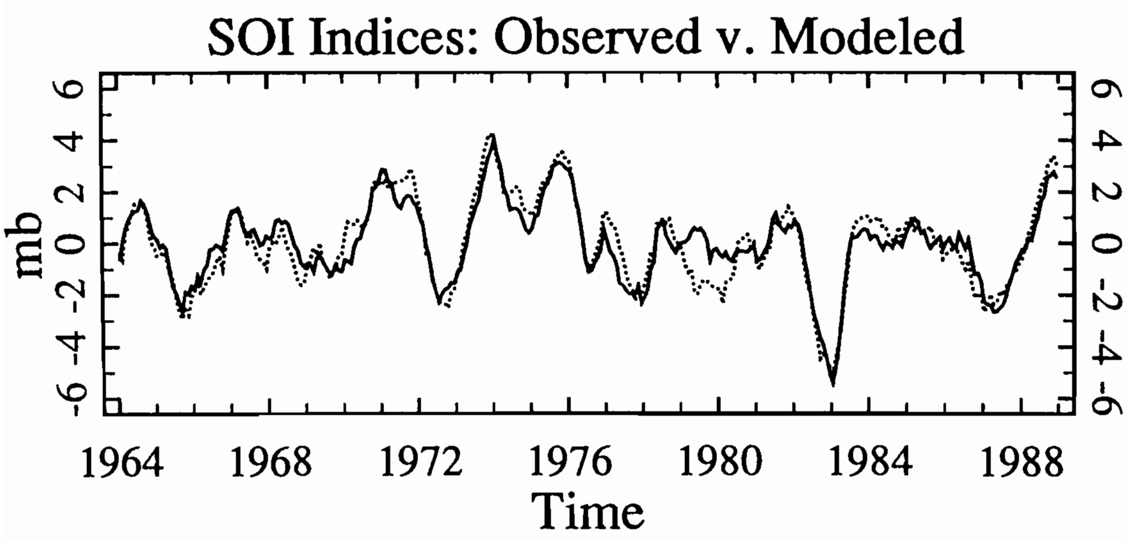

Fig. 3. Southern Oscillation Indices for the period 1964-1988, from pressure observations (solid line) and from pressure estimates based on surface winds (dotted line).

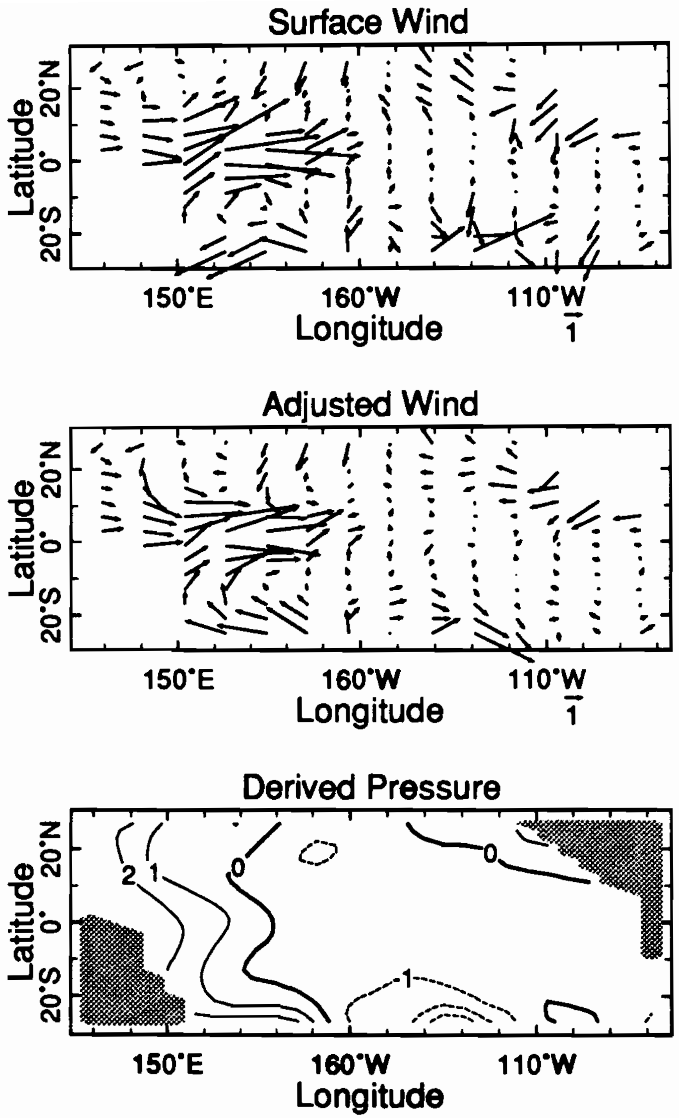

Fig. 4. Surface wind anomalies, estimated pressure anomalies, and adjusted wind anomalies for July 1982, derived using the procedure described in the text.

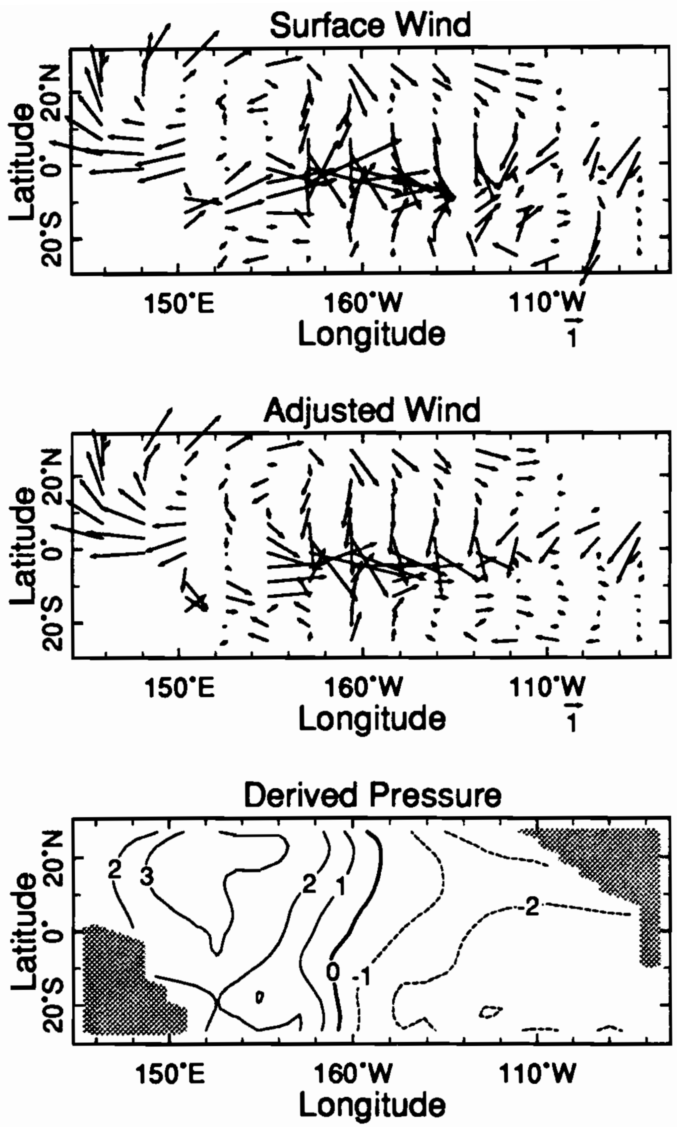

Fig. 5. As in Fig. 4, except for December 1982.

perturbations relative to wind perturbations at higher latitudes (in accordance with geostrophy). The normalization effectively gives equal weight to winds at all latitudes, and is consistent with a given uncertainty in the wind data, independent of position. Also, as the area-mean pressure is not determined by the wind field, it was arbitrarily set to zero.

Other schemes could easily be imagined for estimating the pressure field. For example, one could derive an equation for the Laplacian of pressure, and invert it. The advantage of the present scheme is that it relies on averaging and integrating: fundamentally stable operations that are not scale-sensitive. One expects such a procedure to minimize individual errors quite effectively in producing a "best guess" field.

An obvious test for the derived pressure fields is direct comparison with pressure data. We computed a Southern Oscillation Index based on derived pressure for the period 1964-1988 (Fig. 3). For the present purposes the index was defined simply as the monthly anomaly at Tahiti minus that at Darwin, with no normalization (thus it is an anomalous pressure difference). Comparison with the same index computed from observations shows a very close relationship. The only arguably significant discrepancy occurs in the 1979-1980 period, when the derived index shows a drop that is absent in the observed one.

Another means of testing the pressure fields is through reconstructing the associated wind fields. That is, using (2) and the derived pressure, one can calculate a modified wind field. This modified field will satisfy the linear vorticity balance, and will therefore differ from the original winds in certain ways. In order to judge the whole analysis reasonable, it should happen that the two wind fields show generally good agreement to within the uncertainty of the data, and that the differences are not strongly biased with respect to location (except possibly in very data poor regions).

The reconstructed winds were calculated for all months of the data set. They are presented along with the original winds and derived pressure fields for selected months in Figs. 4-7. For July 1982 (Fig. 4) the pressure field shows a strong zonal gradient in the equatorial west Pacific, extending poleward into the subtropics of both hemispheres. Also featured is a localized low in the vicinity of the cyclonic circulation in the southeastern region. The reconstructed winds are generally similar to the data, with all major features reproduced. The strength of the equatorial westerly anomalies are reduced slightly, as is the divergent component of v to the north. A notably large change is at the center of the cyclonic circulation near 25°S and 130°W: whereas the original wind field shows a somewhat chaotic and strongly divergent flow superimposed on a more coherent cyclonic circulation, the reconstructed field retains only the latter. This is typical of the sorts of changes found in the reconstructed fields; in many (but not all) instances, the changes are obvious improvements.

A mature warm event pattern is found in December 1982 (Fig. 5). The derived pressure field shows the quintessential Southern Oscillation pattern: a very large-scale dipole with centers in the western Pacific and southeastern Pacific. The two wind fields are similar except for the region near 10°S and 160°E, where the reconstructed anomalies are much weaker, and the meridional component is reversed. Also, the meridional component is generally weaker in the southeastern region; our guess is that the reconstructed version underestimates it in this case.

January 1984 presents a very different scenario: strong easterlies in the equatorial west Pacific, continued westerlies in the east, and very strong subtropical anomalies in both hemispheres (Fig. 6). The pressure field depicts a band of low pressure stretching across most of the subtropical southern region, extending all the way to the northern subtropics west of the dateline, with high pressure prevailing in the rest of the domain, and especially strong in the northeastern region. In this case, as the previous ones, the reconstructed winds match the original winds rather closely, with a slightly "cleaner" appearance due to the suppression of strongly divergent and incoherent features.

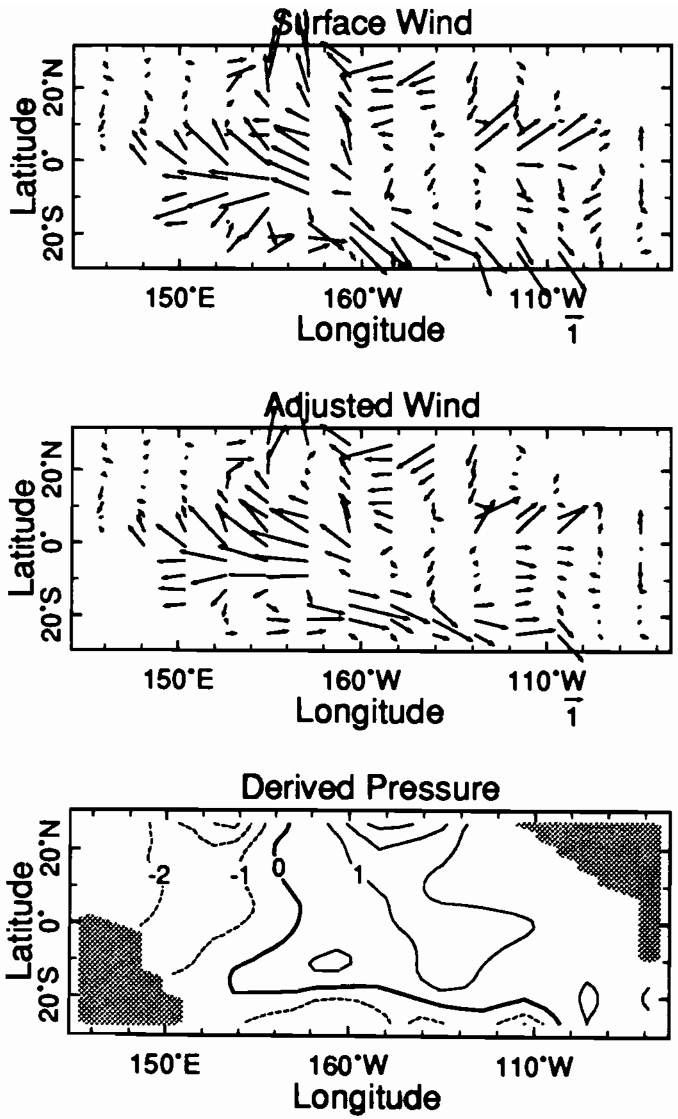

Fig. 6. As in Fig. 4, except for January 1984.

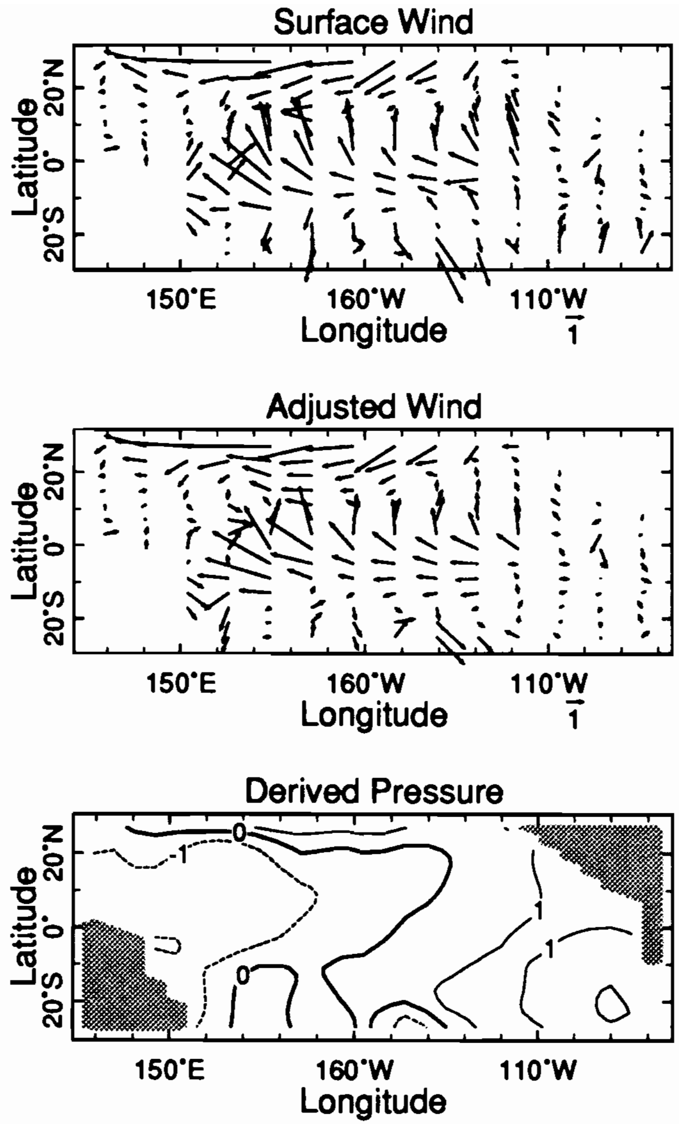

Fig. 7. As in Fig. 4, except for March 1985.

Yet another pattern obtains in March 1985 (Fig. 7). This is a "cold event" scenario; that is, anomalously cold SST in the equatorial east Pacific. Along with this are found strong easterly anomalies in the central equatorial region, and especially strong northward flow to the north, resulting in very strong near-equatorial divergence. Interestingly, the anomalies remain large but become easterly again further to the north, so that a strong convergence results at about 10°N. (A more typical pattern is found in the south: a general reduction or reversal of the zonal component between the equatorial zone and the subtropics.) The pressure field resembles the Southern Oscillation pattern (opposite phase from Dec 1982), except for the band of rapidly increasing pressure in the northernmost latitudes. This is the counterpart to the subtropical easterlies here; the more characteristic pattern tends to have relatively little meridional gradient in the subtropics. The easterlies, and divergence/convergence zones are all reproduced well in the reconstructed wind field.

The examples shown give the impression that the procedure for calculating pressure, and then modifying the winds, results in generally minor changes of the original wind field. To quantify this more precisely, we computed the RMS difference in u and v between the two wind fields, based on all spatial points and all months. The results for both were roughly 0.5 m/s. Thus the adjustments are well below the uncertainty of the data itself, which is roughly 2m/s (Reynolds et al.,1989). Moreover, by construction, the adjusted winds are consistent with the pressure field, and they satisfy the vorticity balance that was not even approximately satisfied in the original wind field. We consider that this alone is sufficient reason to warrant using the adjusted data in subsequent modeling applications. For comparison, the RMS values of the vorticity budget terms based on the reconstructed winds are given in Table 1. The residual term in this case is based only on the linear terms (it is not exactly zero because of differences in the numerical methods used for integration and differentiation). It can be seen that, as before, the nonlinear components are relatively small, but now the linear components are all roughly equal in magnitude. Among these, the divergence term is by far the most changed from the original data.

4. Model Applications

Having a consistent surface wind-pressure data set based on the linear momentum equations affords a good oportunity to evaluate models. Two classes of simple models have been applied to simulating surface winds: one based on a single mode baroclinic structure forced by internal heating (Gill,1980; Zebiak,1982,1986; hereafter Z), and another based on a boundary layer structure forced by buoyancy perturbations associated with surface fluxes (Lindzen and Nigam,1988; hereafter LN). For quite different reasons, both models assume the form of the steady linear shallow water equations, but the forcing is different. Neelin (1989) has shown that the boundary layer model can be written in such a way that the forcing appears in the "thermal" equation, as it does with the baroclinic model, but the theory assigns its structure directly to the SST field. For the baroclinic model, the forcing is assumed to be associated with organized convection. Though this has been parameterized in terms of SST frequently, it clearly has an important dependence on the circulation itself, and particularly the boundary layer convergence (see Z).

How well does either of these models square with observations? This question can now be addressed, since for either model, the combined surface wind and pressure data are sufficient to determine a forcing field uniquely. This "inverted" forcing field is the one that would, in either case, produce the correct circulation (as approximated by the adjusted wind fields). The extent to which this forcing satisfies the theoretical constraints of the models can be taken as an indication of their validity.

The thermal (or mass) equation for either model can be written in the following (nondimensional) form:

$$\varepsilon p + \gamma(u_x+v_y) = -Q\,, \tag{3}$$

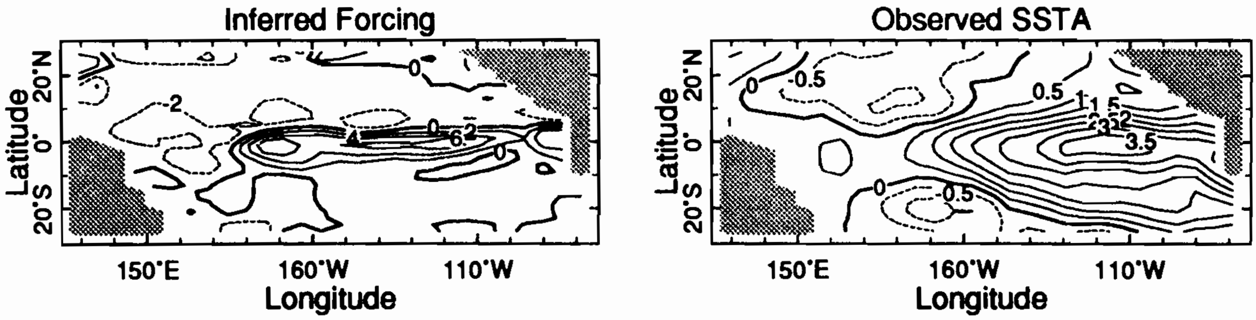

Fig. 8. Inferred forcing (from Eq. 3), and observed sea surface temperature anomalies for December 1982.

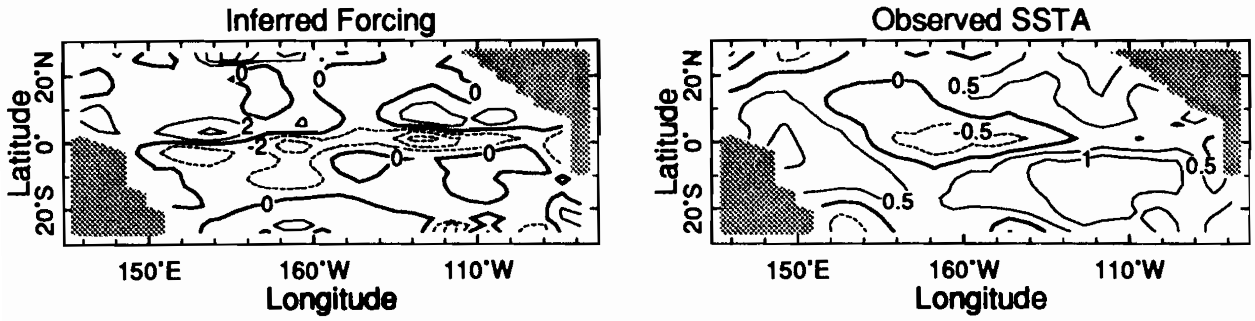

Fig. 9. As in Fig. 8, except for January 1984.

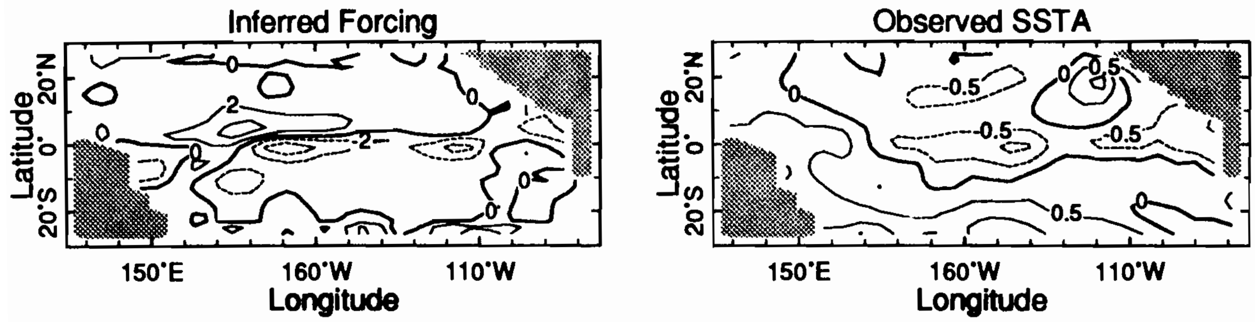

Fig. 10. As in Fig 8, except for March 1985.

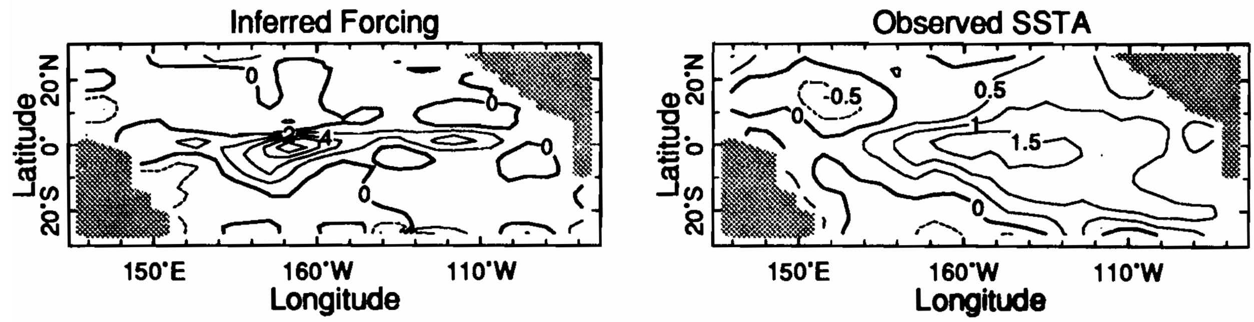

Fig. 11. As in Fig 8, except for December 1986.

with the value of γ and the form of Q depending on the choice of model. For the baroclinic model, Z assumed an internal (Kelvin) wave speed (c_a) of 60 m/s, and nondimensionalized such that γ had a value of unity. The parameterized heating, meant to represent mainly convective latent heat release, depended partly on the local SST anomaly, and partly on low level convergence. However, for the chosen parameters, the SST dependence was strong enough to control the character of the response.

The reader is referred to LN and Neelin (1989) for a detailed description of the boundary layer model. From the latter paper, combined with the nondimensionalization of Z, one can determine that $\gamma = \varepsilon H_0 g / \varepsilon_T c_a^2$, where g is the acceleration of gravity, H_0 is the boundary layer height, and ε_T^{-1} is a "cumulus uptake time". LN chose ε_T^{-1} to be 30 minutes, and H_0 to be 3 km, which gives $\gamma = .68\varepsilon$. The forcing in this model depends only on the SST field; for the LN parameter values, the result is $Q = .85\varepsilon T'$, where T' is the SST anomaly in ° K. Finally, the nondimensional value of ε corresponding to a 1 day dissipation time is 0.25.

It is apparent that the ratio of Q and γ for the baroclinic model is close to the ratio of T' and γ for the boundary layer model. Given that the divergence term tends to dominate the left-hand side of (3), this means that the structure and magnitude of Q and T' from the respective models will be very similar. In fact, we found them to be nearly identical in all cases, and only the baroclinic model heating will be presented here. In dimensional terms, a unit value of Q corresponds approximately to 1.5 mm/day of rainfall. For the boundary layer model, it corresponds to a 1 °C SST anomaly.

The derived forcing field, together with the observed SST anomaly field (CAC analysis), for December 1982 are shown in Fig. 8. Two important points are immediately evident: first, the structure of the two fields is very different -- the forcing has a much smaller meridional scale; and second, the magnitude of the forcing is too large to be explained by the boundary layer model alone. Notice that the heating anomaly changes sign at about 5°N, whereas large positive SST anomalies extend to nearly 15°N in the eastern Pacific. A cautionary note is required here: because the area-mean pressure anomaly is undetermined, and set to zero, the same is true of the derived heating field. Any fluctuation of the net heating over the entire domain that may occur in nature cannot be captured here. In any case, the issues of spatial structure are unaffected by this uncertainty.

For January 1984 (Fig. 9), there is again a conspicuous mismatch in spatial structure between the SST anomaly and derived forcing fields. The forcing field shows intense negative anomalies along the equator in the central and east Pacific, and large positive anomalies to the north. Disregarding the overall mean, there is in this case an identifiable correspondence between the two fields in the equatorial east Pacific, but not in the central and western sectors. Furthermore, it is implausible that such negative anomalies in the eastern Pacific could arise solely from convective sources, as there is too little convective activity in this region to begin with. Thus, in agreement with Gutzler and Wood (1989), there is the suggestion of boundary layer mechanisms being important in the eastern region, and perhaps less so elsewhere. Similar conclusions could be drawn from the March 1985 fields (Fig. 10). In this case the scale difference between SST and heating in the central Pacific is unmistakable.

December 1986 was an instance of a moderate warm event (Fig. 11). Equatorial westerly anomalies were large in the western and central regions, with equatorward flow to the north and south, but little response to the east. This typical pattern was discussed by Z as problematical; the model response for mature phase ENSO conditions tends to resemble the familiar pattern of Gill (1980), with broad easterlies to the east of the heat source. Many models, including some GCMs, tend to do this; yet the observed pattern shows the westerlies and cyclonic circulations in the heating region without the easterlies to the east. Neelin and Held (1987) achieved a more realistic simulation by specifying the vertical motion from a GCM calculation, suggesting that the problem in other models may be in the parameterized forcing. The "inverted" forcing field shows, once again, a substantial scale reduction relative to the SST anomaly field, especially in the eastern sector. Positive heating anomalies are situated in the equatorial central Pacific, and along a very narrow band farther to the east, with predominantly negative anomalies in western Pacific. The general pattern is not unlike what is inferred from OLR anomalies during ENSO events (with the possible exception of the eastern equatorial region), but tends to show even smaller scales. This difference could easily arise from the extensive smoothing of OLR data.

The origin of the "easterly problem" discussed in Z is now evident: the parameterized heating is too strongly tied to the SST field. Whereas the prescribed heating tends to have the broad structure of the SST anomaly field, the required forcing is much more confined, with a tendency for nearby compensation (that is, anomalies of the opposite sign). The impact on the easterlies is dramatic, since they arise from the Kelvin wave component, as discussed by Gill (1980). This component is forced at a given longitude by the projection of the heating onto the first Hermite function, which has the form of a Gaussian (centered on the equator) with y-scale of order 10° latitude. A function with the structure of the SST anomaly field of Fig. 11, for example, projects strongly onto the Kelvin component, whereas one like the inferred heating field projects very weakly. On the other hand, the projection onto the gravest Rossby modes, those responsible for the equatorial westerlies, is if anything stronger with the more confined forcing.

5. Conclusions

Starting from the FSU wind analyses for the tropical Pacific, we have performed several calculations. First, a vorticity budget for the surface layer was computed. It reveals the relative unimportance of the nonlinear components, but is not even approximately balanced. Based on the former result, and an hypothesis that the latter is largely due to data errors, an algorithm was developed to estimate the surface pressure field from the wind observations and the linear momentum equations. The estimated surface pressure agrees well with direct observations in terms of the SOI variability over the past 25 years, and also matches closely recent operational analyses that were examined.

Using the same linear equations, and the derived pressure field, a revised wind field was determined. This wind field, by construction, is consistent with the pressure field, and satisfies the linear vorticity budget. We feel confident that the modified winds are superior to the original data since the adjustments amount to only about .5 m/s in a root-mean-square sense, well below the uncertainty of the data, and yet a dynamical balance is achieved. This result is very dependent on a carefully chosen method of pressure field construction. Rather than using a method involving derivatives of the data, we adopted an integral technique, and one that further allowed for preferential weighting of information in the very sensitive equatorial zone.

The combined surface wind and pressure fields were then used to infer forcing fields within the context of two simple atmospheric models, one based on a tropospheric scale baroclinic circulation forced by internal heating, and the other based on a boundary layer circulation forced by buoyancy perturbations associated with surface fluxes. Both have been presented as relevant to the variability of tropical surface winds. We found that the structure of the inferred forcing for the two models was nearly identical, within reasonable parameter ranges. This might be taken as evidence that the two models are indistinguishable, except for the fact that the derived forcing differs importantly from the SST anomaly fields. The buoyancy forced model is based on the theory that the flow is driven by pressure gradients set up hydrostatically by temperature perturbations tied to SST. Thus, for the model to be validated, it should happen that the derived forcing has the SST anomaly structure. There is some evidence of this relationship holding in the eastern tropical Pacific, but not elsewhere.

On the other hand, the baroclinic model as presented in Zebiak (1986) is also shown to be wrong, as its parameterized heating depends heavily on the SST field. The most systematic failure of this model in simulating ENSO is the fictitious easterlies it produces in the eastern Pacific. This problem vanishes when the forcing assumes a smaller-scale structure as in the inferred heating fields.

It might be argued that the boundary layer calculations are incorrect, since the model really applies to layer-averaged winds, rather than the surface winds used to infer forcing. It is certainly true that the meridional component of winds at the surface, with associated strong convergence, is largely absent at 850 mb. Thus the more appropriate layer-averaged convergence will typically be smaller than that implied by the surface wind. This might account for the discrepancy in amplitude between inferred and observed SST fields. Nonetheless, even allowing boundary layer turning and deceleration, one cannot account for the needed scale reduction -- additional mechanisms seem to be required. Fundamentally, the problem for this model is that the surface pressure and SST anomaly fields do not closely match.

What mechanism could be responsible for forcing at such small scales? The most plausible candidate is cumulus convection. Our results are not inconsistent with OLR analyses, considering the resolution and degree of smoothing typically applied to such fields. If convective heating is indeed the mechanism, then it follows that the large scale dynamics minimally represented by the Gill (1980) model must be included in order to simulate the real variability accurately. On the other hand, the results for the eastern Pacific are suggestive of the importance of boundary layer dynamics. In such convectively suppressed regions, this is the most likely mechanism controlling surface winds.

These results should be verified with other wind products, but it is hard to imagine that the major conclusions could change, as they depend on rather striking and consistent features. Apparently, what is needed (even for simple models) is a reasonable parameterization of organized convection, based on the large scale circulation and SST. Unfortunately, a satisfactory theory for this is lacking; issues such as what controls the scale of convective complexes, and the interaction and competition of convectively active regions, and the modulating influence of SST, are poorly understood. Nonetheless, it appears that the benefits of further progress in this area could be great. An additional implication of this work is that a rather fine resolution is required to simulate the real atmosphere faithfully. The scale of heating anomalies that should be resolved appears to be 2-3 degrees of latitude. This may be the explanation for why several low resolution models, including GCMs, show some of the same spurious easterlies as described above in ENSO simulations. If this is true, then only an increase in resolution can remedy what has frequently proven to be a devastating effect in coupled models.

ACKNOWLEDGEMENTS

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