Calculating Tropical Winds from Time Mean Sea Level Pressure Fields

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ABSTRACT

The time-mean tropical surface momentum balance is investigated with a simple model that calculates tropical surface winds from time mean sea level pressure fields. The model domain is the global tropical strip centered on the equator with lateral boundaries at ±30° latitude. Steady state surface winds are numerically calculated from the nonlinear horizontal momentum equations, with forcing from observed climatological monthly mean sea level pressures and prescribed lateral boundary winds. Dissipation is parameterized by linear damping and diffusion. Comparisons of model winds with observed climatological monthly mean winds show realistic simulations in most regions and in all months. The poorest simulations occur in the meridional component of the wind in near-equatorial areas of strongly convergent or weak winds. In these areas, and in the near-equatorial region generally, diffusion processes make a significant positive contribution to the realism of the model winds. Horizontal nonlinear advection also improves the simulation near the equator, though to a smaller degree. The generally skillful model winds refute the conventional idea that weak gradients make the tropical pressure field a poor tool for calculating tropical winds. To the contrary, tropical pressure fields contain substantial information about associated winds. Thus, a relatively complete momentum balance can be identified for the major features of the time-mean tropical wind field.

1. Introduction

Estimating the wind from a given pressure distribution is a very basic and old part of meteorology. The quasi-geostrophic nature of large scale dynamics in midlatitudes allowed observers around 1860 to interpret isobars on daily weather maps as quasi-streamlines. They also noticed that the flow near the ground makes a small angle with the isobars, directing the air toward lower pressure.

Starting first from principles and using scale analysis, one can in fact show that on large space scales the Coriolis and pressure gradient forces should be in near balance over much of the globe, while the time rate of change and nonlinear advection are of lesser importance. To understand flow in the planetary boundary layer, the consideration of friction in even its crudest form, linear drag, allows us to understand qualitatively the angle between wind and isobars in terms of the three-way balance between pressure gradient, Coriolis, and frictional accelerations.

Obviously, near the equator quasi-geostrophy breaks down, and as a result the relationship between wind and pressure is no longer simple. Various attempts have been made to determine just what the major elements of the tropical relationship are (e.g., Ramage 1971; Saha and Suryanarayana 1971; Mahrt 1972; Lettau 1974; Romanov and Romanova 1976; Rao et al. 1978; Stout and Young 1983; Panchev 1985). An important implication of some of these studies (e.g., Ramage 1971; Stout and Young 1983) is that small pressure gradients in the tropics and errors in measuring pressure are major obstacles to inferring the wind from the pressure field. Ramage (1971, p. 212), in a discussion of tropical monsoon winds writes:

> "One must conclude that most of the time in the monsoon area the pressure field can neither be quantitatively nor qualitatively related to the three dimensional field of motion . . ."

However, despite this conclusion, the motivations for pursuing and clarifying the relationship between tropical wind and pressure remain compelling. One of the more important of these motivations comes from the crucial role played by tropical surface winds in the dynamical theories for a variety of atmospheric and oceanic phenomena [e.g., the Hadley and Walker circulations, monsoons, El Niño/Southern Oscillation (Gill 1982; Hastenrath 1985)].

In this study, we examine the relationships between tropical winds and pressure with a simple numerical

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model that calculates tropical surface winds from a given sea level pressure distribution. Our method will be to integrate the nonlinear momentum equations toward equilibrium while keeping the pressure field fixed. Our focus is on long term monthly mean climatological wind and pressure fields for the global tropical strip ($0° \leq \lambda \leq 360°$, $30°S \leq \phi \leq 30°N$). The problem is a challenging one because of (i) the large deviation from geostrophy of the tropical momentum balance, (ii) the notoriously small pressure gradients, and (iii) the problematic nature of both the input pressure and verifying wind fields. Therefore, our study is also, in part, a mutual quality check of observed climatological pressure and wind fields, which are taken from Oort (1983), who analyzed the two fields completely independently. In philosophy, our approach has much in common with the studies of Rao et al. (1978) and Sardeshmukh and Hoskins (1985). In these studies, the question is asked: How much of the flow can be retrieved by specifying the forcing? We and Rao et al. retrieve the wind field from the specified pressure field, while Sardeshmukh and Hoskins retrieve the streamfunction from the specified divergence.

As the equator is approached, decreasing Coriolis effects allow only relatively weak pressure gradients to form. These weak pressure gradients suggest that with decreasing latitude, the tropical momentum balance may become less dependent on the local terms (e.g., the local pressure gradient, Coriolis force, and linear drag) and more dependent on remote dynamical conditions. Thus we anticipate that on nearing the equator, the nonlinear advection terms become increasingly more important in the simulation of observed winds. In an idealized numerical model of the near-equatorial boundary layer, Mahrt (1972) found indications that nonlinear advection may be significant for certain pressure gradient patterns. Hastenrath and Lamb (1978) and Greenhut and Bean (1981) found some empirical indications that nonlinearity may be important in the momentum balance over the tropical Atlantic and Pacific. However, Romanov and Romanova (1976) concluded from observations in the eastern Atlantic that the nonlinear terms are negligible in the surface balance, even near the equator. And Egger et al. (1981) found that over an area in the equatorial Pacific, the three-way balance alone gives reasonable results for departures from normal of monthly mean winds. Thus it is interesting to see how much the nonlinearity actually contributes to the realism of our model's winds, and to that end we use a formal skill measure. In particular, we apply this skill measure in comparisons of our model winds with those derived from simpler balances.

In this paper we limit the discussion to observed and calculated long term monthly mean winds in the tropical strip during December and January, and in the Indian Ocean region during the development of the Asian summer monsoon. A standard version of the model was run for these and all other months. The model is described in section 2. The data used to force the model and to verify its results are discussed, along with the verification procedure, in section 3. In section 4 the standard version results for December are presented for the tropical strip, while variations from the standard run are discussed in section 5. These variations include changes in the poorly known dissipation constants. In section 6 we discuss the model results for April through July over the Indian Ocean and southern Asia. We then consider in section 7 the possibility of uncertainties in the calculated winds arising from pressure field uncertainties. Finally, in the summary and discussion we discuss the successes of this study, as well as its limitations and directions for further study. We conclude by considering additional applications of this model (e.g., to shorter time scales, higher levels, initialization of wind fields, estimates of wind stress, or, when run in reverse, estimates of pressure from prescribed winds). From the outset, we expect the model to do poorly over elevated land, since there both sea level pressure and wind are artifacts. Thus we focus on pressure–wind relationships over the ocean in our analysis of model results and our consideration of additional applications.

2. The model

The model is based on the two approximate equations of horizontal time mean motion:

$$0 = \frac{\partial u}{\partial t} = -\frac{1}{\rho}\frac{\partial p}{\partial x} + fv - ku + D\nabla^2 u - \left(u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y}\right) \quad (1)$$

$$0 = \frac{\partial v}{\partial t} = -\frac{1}{\rho}\frac{\partial p}{\partial y} - fu - kv + D\nabla^2 v - \left(u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y}\right) \quad (2)$$

where the notation is conventional, except that all quantities are long term monthly means; ∇^2 is the horizontal Laplacian operator. The equations will be applied at sea level. We prescribe sea level pressure as a fixed function of space, and therefore (1) and (2) form a closed set in the horizontal wind components u and v, which can be integrated in time. Our aim is to find stationary solutions in u and v that balance the prescribed pressure field.

In (1) and (2) we assume that the time derivatives are small, which is justified since we calculate (u, v) forced by long term mean pressure fields. We neglect the $\tan\phi$ terms related to the earth's curvature, since these terms are very small in the tropical strip. We express dissipation processes as a linear damping ($-ku$, $-kv$) plus a horizontal diffusion ($D\nabla^2 u$, $D\nabla^2 v$) of momentum. The diffusion of course may drive the momentum in an area toward zero in much the same way as the linear damping, except that the effective damping coefficient of the diffusion increases with the square of the wave number. However, diffusion, as it dissipates momentum at one gridpoint, transfers momentum to nearby points. Thus the diffusion terms may simulate both momentum removal and mixing processes.

We neglect in (1) and (2) the vertical advection terms. For much of the tropics, this neglect seems acceptable since scale analysis (Charney 1963) indicates that vertical advection may be safely ignored outside of active precipitation areas. Further, we may expect that in precipitation areas, low level convergence would tend to produce vertical advections that remove momentum from the surface flow. Thus, the effects of vertical advection in precipitation areas may be partially accounted for by the model's dissipation terms.

Perhaps a more serious approximation in our equations for monthly mean winds is the absence of explicit transient effects, which results from averaging the nonlinear terms. The importance of this neglect is hard to assess, primarily because, to the extent that transients dissipate the time mean field (Lau 1979), their effects may already be included in the dissipation terms, and thus in our choice of the poorly known constants, k and D. We may therefore see the friction and diffusion terms as representations of several processes that, in explicit form, were left out when we reduced the governing equations to (1) and (2). The potential role of the transient terms in the real atmosphere's momentum balance is discussed further in section 8.

Equations (1) and (2) are discretized on a 2.5° by 5.0° latitude–longitude grid. Horizontal derivatives are approximated by centered differences. To find a stationary solution, (1) and (2) are integrated in time using a leapfrog scheme, with one forward time step at $t = 0$. The time step is one-half hour. We are not interested here in the evolution in time of u and v towards a stationary solution, but use time integration simply as an iteration technique to find that solution. In order to suppress the computational mode, a 1–2–1 time filter is applied at every 10th time step. The solution is reached when, at two successive time levels, $n\Delta t$ and $(n+1)\Delta t$,

$$|u_{n+1} - u_n| < 0.01 \text{ m s}^{-1}$$

$$|v_{n+1} - v_n| < 0.01 \text{ m s}^{-1}$$

at all gridpoints of the domain. As long as a reasonable initial state is chosen, it takes about 100 time steps to meet these convergence criteria.

The standard model runs that we discuss in the following sections have the following features in common. (i) The horizontal domain is periodic in the longitudinal direction ($0° \leqslant \lambda \leqslant 360°$) and extends from 30°S to 30°N in the latitudinal direction. (ii) ρ is a constant taken as 1.21 kg m^{-3}. (iii) The Coriolis parameter is fully variable. (iv) The lateral boundary conditions (at ϕ = 30°S and 30°N) are the observed u and v components. (v) k is 1.8×10^{-5} s^{-1} over the ocean and 12.6×10^{-5} s^{-1} over land (7 times k over the ocean). (vi) $D = 5 \times 10^6$ m^2 s^{-1} everywhere. Thus, for scales smaller (larger) than about 1000–3000 km, the diffusion is a stronger (weaker) damping mechanism than linear friction. The values for k and D were chosen after a subjective comparison of the simulated and observed wind fields for January and July.

In order to assess the value of a model based on (1) and (2), we will make comparisons with simpler and even cheaper models, such as the geostrophic balance (e.g., $\rho^{-1}\, \partial p/\partial x = fv$) and the three way balance (e.g., $\rho^{-1}\partial p/\partial x = fv - ku$). We will not explicitly compare our model to the nonlinear rotational balance equation (see Holton 1979, p. 179), but it is useful to realize that (1) and (2) differ from that balance equation only in that our solution includes the divergent part of the wind.

The question arises whether, given a pressure field, there will always be a solution for u and v. For the nonlinear balance equation, it is well known that solutions do not exist when the pressure field becomes "non-elliptic" (see for example Baumhefner 1968). Peagle and Peagle (1974) outline two answers to this problem. The first is to smooth the pressure field so as to make it elliptic, and the second is to allow some divergence. Obviously (1) and (2) allow divergence, and therefore they should, in general, have solutions for observed pressure fields. However, it is possible that errors in the input pressure field may cause (1) and (2) to violate our analogue of the ellipticity condition. For such errors, we expect the diffusion of momentum to ensure convergence to the solution. We have not encountered any case where more than one solution seems to exist.

In addition to the above time stepping procedure for determining the winds in balance with a given pressure field, we have also experimented with a number of other procedures. In addition to various time-stepping techniques, we tested a procedure based upon the following iteration equations.

$$fv_{n+1} - ku_{n+1} = \frac{1}{\rho}\frac{\partial p}{\partial x} - D\nabla^2 u_n + \left(u_n \frac{\partial u_n}{\partial x} + v_n \frac{\partial u_n}{\partial y}\right) \quad (3)$$

$$-fu_{n+1} - kv_{n+1} = \frac{1}{\rho}\frac{\partial p}{\partial y} - D\nabla^2 v_n + \left(u_n \frac{\partial v_n}{\partial x} + v_n \frac{\partial v_n}{\partial y}\right). \quad (4)$$

In this procedure we set the initial winds to zero ($u_0 = v_0 = 0$) and applied the above convergence criteria. We found that *if* the iteration converges, (3) and (4) require less computer time and converge toward the same solution as that found by time stepping. However,

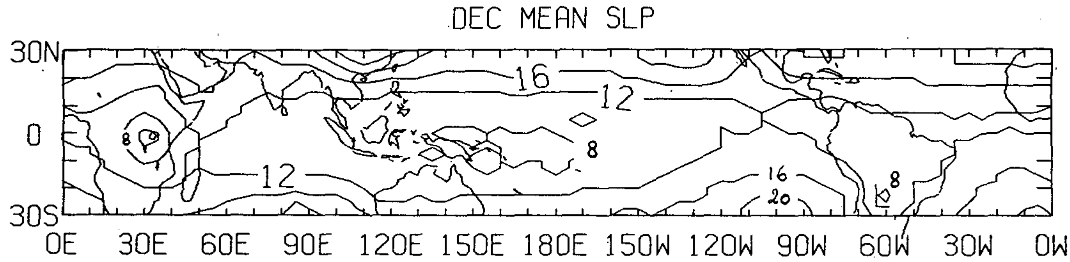

FIG. 1. December mean sea level pressure in millibars minus 1000 mb., from Oort (1983). Contour interval is 4 mb.

the iteration often does not converge, with divergence sometimes occurring after five to ten steps that deceptively resemble convergence. Thus, we chose time stepping as the more reliable method of finding stationary solutions.

3. Data and verification procedure

The input sea level pressure fields are the 10-year (1963–73) monthly means, a description of which can be found in Oort (1983). From the p fields, wind fields (u_c, v_c) are calculated using (1) and (2), and are then compared to the appropriate 10-year monthly mean observed surface winds (u_o, v_o). The observed fields were obtained on a 2.5° × 5.0° latitude–longitude grid coinciding with the model's grid. Oort's analyses of p and (u_o, v_o) are completely independent, and therefore comparing (u_c, v_c) to (u_o, v_o) is a fair mutual consistency check on the data and a check on the validity of (1) and (2). The accuracy of this climatological dataset, and comparisons of it to other sets, are extensively discussed by Oort (1983).

For one wind component (u, say), the agreement between observed and calculated winds is expressed as the component skill,

$$S_u = 1 - \sum_m (u_o - u_c)_m{}^2 / \sum_m (u_o)_m{}^2. \quad (5)$$

We also consider the joint skill for u and v, defined as

$$S = 1 - \sum_m [(u_o - u_c)^2 + (v_o - v_c)^2]_m / \sum_m (u_o{}^2 + v_o{}^2)_m. \quad (6)$$

For each skill measure, m is a gridpoint index over which we sum (e.g., $m = 1$ would give a two dimensional geographic distribution of the skill; $m = 72$ gives the skill as a function of latitude, that is, a zonally averaged skill). Our skill measures have the properties that a perfect simulation gives $S = 1$, and a zero wind simulation gives $S = 0$. A negative skill corresponds to a model wind that differs in sign from the observed wind, or that has a speed more than twice the observed. Thus in areas where the observed wind vector changes sign and/or is weak, negative skills are more likely. In sections 4 and 5, we will compare skills for the standard model with skills for several variations of the model and the skills of the geostrophic and three way balances, as well.

Our model is designed to simulate sea level winds that arise from sea-level pressure gradient forces. In areas of surface elevation we therefore expect errors, since observed winds at the surface cannot easily be converted to sea level winds, and since sea level pressure is artificial in these areas. Oort's sea level pressures over land do in fact show several major features that are not present in earlier analyses (e.g., Taljaard et al. 1969; Crutcher and Meserve 1970; Godbole and Shukla 1981). In particular, Oort shows a small scale, relatively deep low over East Africa (see Fig. 1, for example). These comparisons with other climatologies also show that Oort's pressures are generally higher, by about two to six millibars, over much of Australia and South America. Apparently the reduction of pressure to sea level is not a trivial matter. Thus, over land areas where significant elevations occur, our model is likely to have less skill. For this reason, we will de-emphasize in our

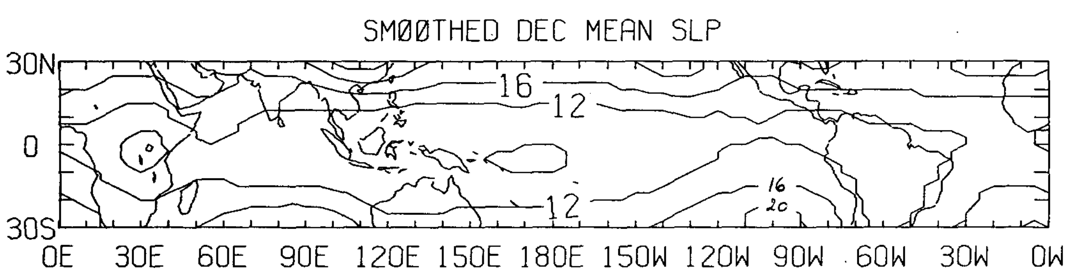

FIG. 2. As in Fig. 1 except after six passes through a 1–2–1 spatial smoothing filter.

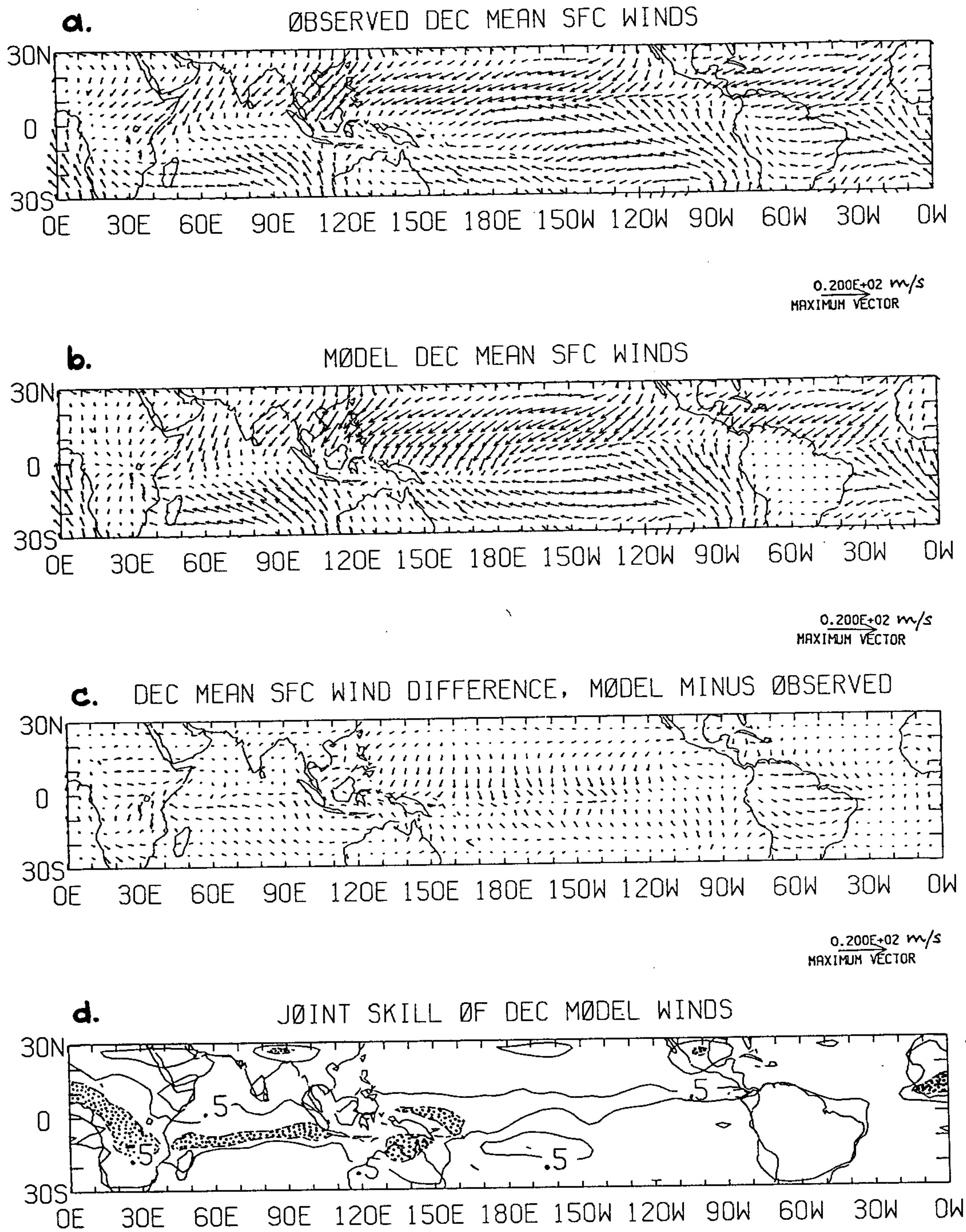

FIG. 3. Comparisons of December mean surface winds from observations (Oort 1983) and standard model: (a) observed winds after two passes through a 1–2–1 spatial smoothing filter; (b) model winds; (c) wind difference, model minus observed; (d) joint skill, S, of the model winds. Vector scale (in m s^{-1}) as shown. Vectors plotted at 5° latitude by 5° longitude resolution, less than model's 2.5° by 5° resolution, for clarity. Contours in (d) from -0.5 to $+0.5$; contour interval is 1 skill unit; negative contour dashed; zero contour omitted; areas where skill is less than -0.5 are stippled.

discussion of the model results the model's performance over land. The Oort pressure data also contain several significant (and questionable) small scale features in the Pacific along the equator. To smooth out these troublesome features, we pass the pressure data six times through a successive 1–2–1 space filter in both horizontal directions. This filter is also applied twice to the observed winds. On the whole, the Oort wind

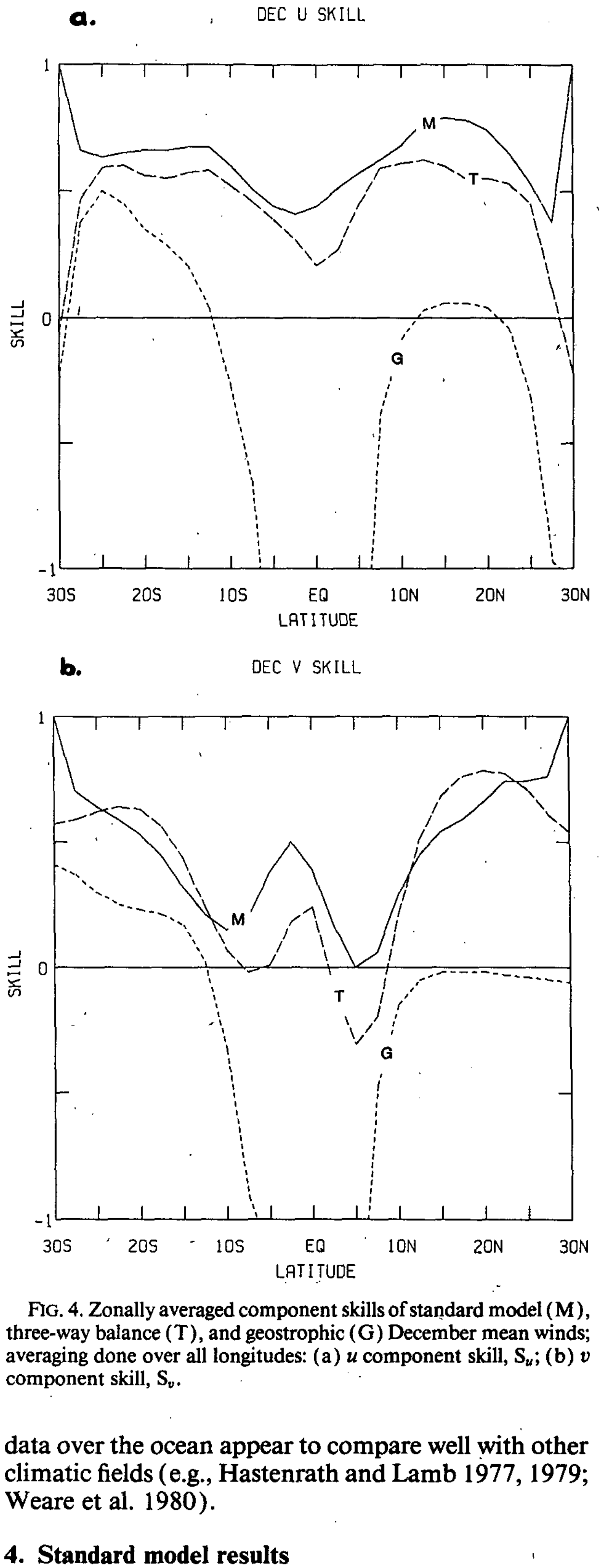

FIG. 4. Zonally averaged component skills of standard model (M), three-way balance (T), and geostrophic (G) December mean winds; averaging done over all longitudes: (a) u component skill, S_u; (b) v component skill, S_v.

data over the ocean appear to compare well with other climatic fields (e.g., Hastenrath and Lamb 1977, 1979; Weare et al. 1980).

4. Standard model results

The model has been run for all 12 months. The results for December provide a representative case study for the model's strengths and weaknesses. The smoothed pressure field used for the December case is shown in Fig. 2. In Fig. 3, the December observed and model wind fields are compared through map displays of wind vectors, wind vector differences (model minus observed), and simulation skill, S.

The wind vector maps (Figs. 3a and 3b) show a strikingly good agreement of the overall speeds and flow patterns. In particular, the near-equatorial convergence zones (CZ) are fairly well located by the model. However the model positions the western Pacific CZ too far south and seems to have too weak a South Pacific CZ. In the Indian Ocean sector the model CZ is located about one gridpoint too far north. This may indicate resolution problems arising from the relatively coarse grid (2.5° lat by 5.0° long). The excessive

TABLE 1. Case studies made in testing variations of the standard model. Different cases employ different combinations of terms from equations (1) and (2) to find mean surface winds from prescribed tropical December mean sea level pressure.

Case	Terms included
1	Coriolis, pressure gradient (geostrophic balance)
2	Coriolis, pressure gradient, nonlinear (unstable)
3	Coriolis, pressure gradient, diffusion
4	Coriolis, pressure gradient, diffusion, nonlinear
5	Coriolis, pressure gradient, friction (three-way balance)
6	Coriolis, pressure gradient, friction, diffusion
7	Coriolis, pressure gradient, friction, nonlinear

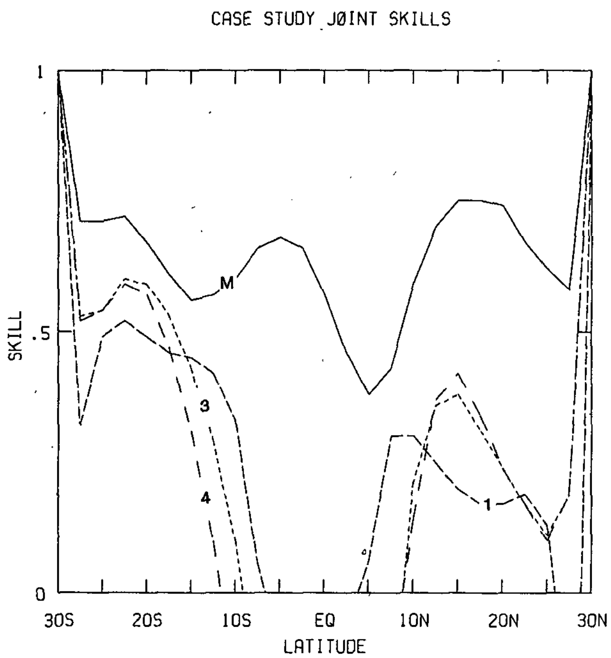

FIG. 5. Zonally averaged joint skills, S, of December mean winds for standard model (M), case 1 (1), case 3 (3), and case 4 (4). Cases explained in text and Table 1. Averaging done over central and eastern Pacific (180°–90°W).

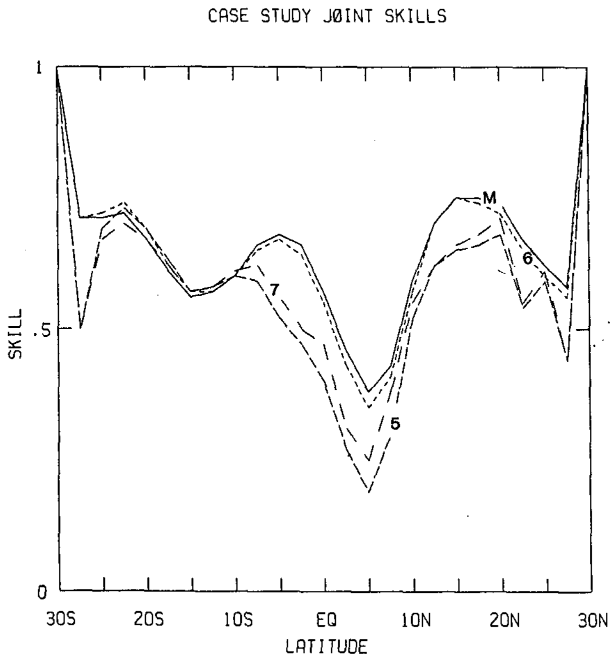

FIG. 6. As in Fig. 5 but for standard model (M), case 7 (7), case 6 (6), and case 5 (5).

convergence over Africa and the weak convergence and flow over South America suggest that the input sea level pressure in these areas may in fact be too low and too high, respectively, as discussed in section 3.

The model's successes and difficulties are perhaps more clearly seen in the wind vector difference map (Fig. 3c; same vector scale). Over most of the domain, the speed errors are less than 2 m s^{-1}. The largest errors are 3–4 m s^{-1}. These occur in the northeast trades of the central and western Pacific, where the model winds are too northerly. The problematic African and South American pressures are expressed again in the wind differences for these areas.

Figure 3d shows the joint skill, S. The major regions of low skill are: 1) the near-equatorial CZ in the eastern hemisphere; 2) areas of problematic pressure data; and 3) land areas. Obviously, by adjusting the spatial dependence of k and D, some of these low skills could be improved upon. We have not done this, however, because our simple model lacks the physical justification for such a step. Over most of the tropics though, the model's simulation of the basic flow pattern is very good and the skill is positive. It appears therefore that the equations we use are basically correct despite the neglect of possibly important terms.

These basic model strengths and weaknesses are clearly seen in the zonally averaged skills for December, shown in Fig. 4 for S_u and S_v. The skills for the geostrophic and three-way balance winds are included in this Figure for comparison. For the case shown, the model boundary winds are the observed values from Oort (1983). Thus the model skill is 1.0 at the boundaries.

As expected, geostrophy gives a poor representation of surface flow at low latitudes, with infinite errors near the equator. The addition of linear friction alone, to give the three-way balance, makes a big improvement, but skills are still low within about ten degrees of the equator, especially in the v component. Running the full model gives significant further increases in skill at almost all latitudes, most noticeably near the equator where considerable improvement comes from the inclusion of the nonlinear advection and diffusion processes. This improvement is especially striking for v in the CZ regions, between 10°N and 10°S.

The standard model winds for the other months show skills that are very similar to those for December. In all months, the model winds have their lowest skills in the v component and in near-equatorial areas where the observed wind is strongly convergent and/or weak. But in all months, the model's representation of the tropical, and especially the near-equatorial, surface momentum balance is an important improvement over simpler balances. Thus, the model's horizontal nonlinear advection and diffusion processes seem to be important terms in the tropical surface momentum balance. The precise impact of these terms will be analyzed in the next section.

5. Model variations

The standard model configuration was chosen to provide, on the whole, the most skillful simulations. Variations of this model, though giving less skillful results overall, reveal some interesting features of the

TABLE 2. Absolute values of the terms in the standard model solution for December, averaged over 180°–90°W. Units are 10^{-5} m s^{-2}.

Latitude	$u\frac{\partial u}{\partial x}$	$v\frac{\partial u}{\partial y}$	$D\nabla^2 u$	ku	fv	$\frac{1}{\rho}\frac{\partial p}{\partial x}$	$u\frac{\partial v}{\partial x}$	$v\frac{\partial v}{\partial y}$	$D\nabla^2 v$	kv	fu	$\frac{1}{\rho}\frac{\partial p}{\partial y}$
7.5	0.3	0.9	1.5	7.4	4.7	2.7	0.3	0.7	1.7	4.5	7.8	13.5
5.0	0.2	0.2	2.6	6.0	2.7	3.0	0.3	0.7	1.5	3.8	4.3	7.2
2.5	0.2	0.5	2.7	5.3	1.5	3.2	0.3	0.9	2.8	4.1	1.9	5.4
0	0.2	0.7	2.0	5.3	0.0	3.6	0.3	0.7	2.7	4.3	0.0	5.1
−2.5	0.2	0.8	1.2	5.9	1.6	4.0	0.3	0.4	1.7	4.5	2.1	6.3
−5.0	0.3	0.7	1.4	6.9	3.0	4.4	0.3	0.2	1.5	4.3	4.9	8.5
−7.5	0.3	0.6	1.6	7.9	4.0	4.9	0.3	0.2	1.2	3.8	8.4	12.2

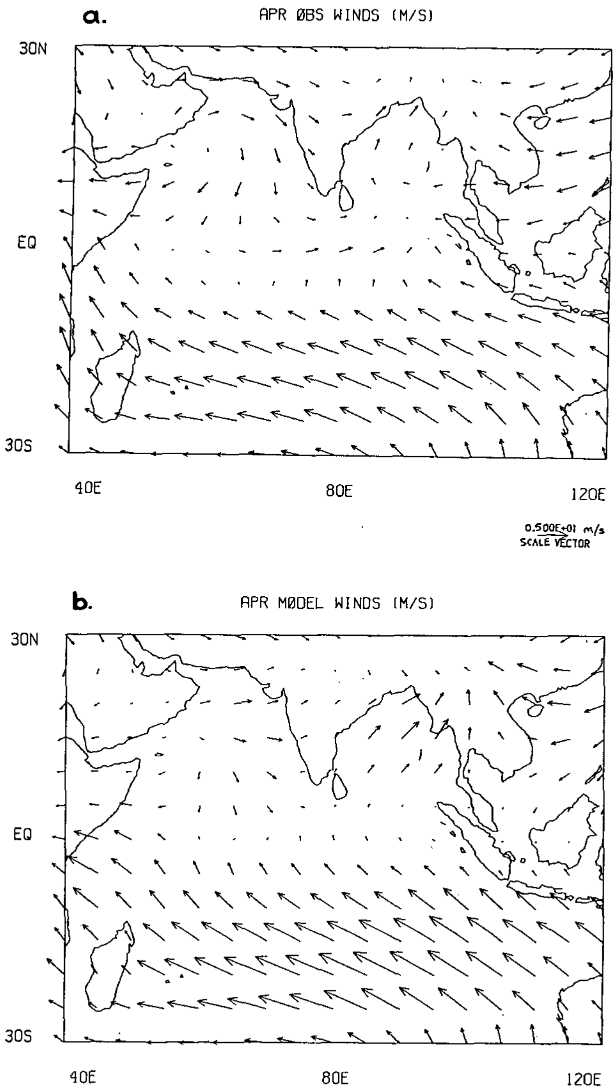

FIG. 7. As in Fig. 3a, b but for April mean surface winds in Indian Ocean region and with vector scale (in m s^{-1}) as shown here.

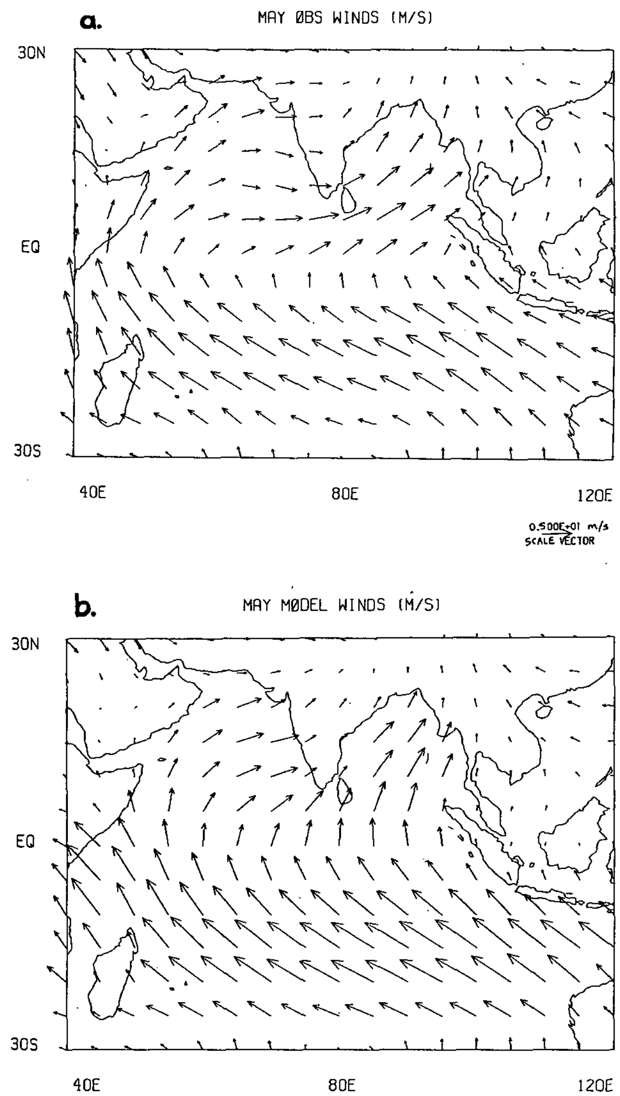

FIG. 8. As in Fig. 7 but for May.

model and, apparently, of the tropical surface momentum balance. The variations we tested include running the model *without* the effects of: (a) linear friction, (b) horizontal diffusion, (c) nonlinear advection terms, and (d) combinations of (a) through (c). In addition, various values for the dissipation coefficients, k and D, and the lateral boundary conditions were tested.

We need to make one point clear here. The nonlinear terms are (and should be) explicitly present in (1) and (2). Thus, experiment (c) gives an indication of how important the nonlinear terms are in the tropical atmosphere. On the other hand, the linear drag and horizontal diffusion terms are parameterizations (perhaps very crude ones) of processes not explicitly included in (1)–(2). Therefore, experiments (a) and (b) will yield less definitive insights into the balance of terms in the real atmosphere. However, we treat all the experiments in the same formal way by comparing their joint skills, S, to the skill of the standard version.

Seven cases are considered in which various terms are removed from (1) and (2). These are listed in Table 1. The zonally averaged skills for the seven cases are used as a measure of the relative importance of the terms included. The December joint skills for these cases, averaged over the central and eastern Pacific region (180°–90°W), are shown in Figs. 5 and 6, and are discussed below as representative examples. This Pacific region was chosen for these case studies because it is a large region that is relatively free of problems with topographic effects or conversion to sea level pressure. In addition, Oort's (1983) pressure fields for this area are fairly consistent with those of other workers (e.g., Shea 1986). The basic conclusions of these

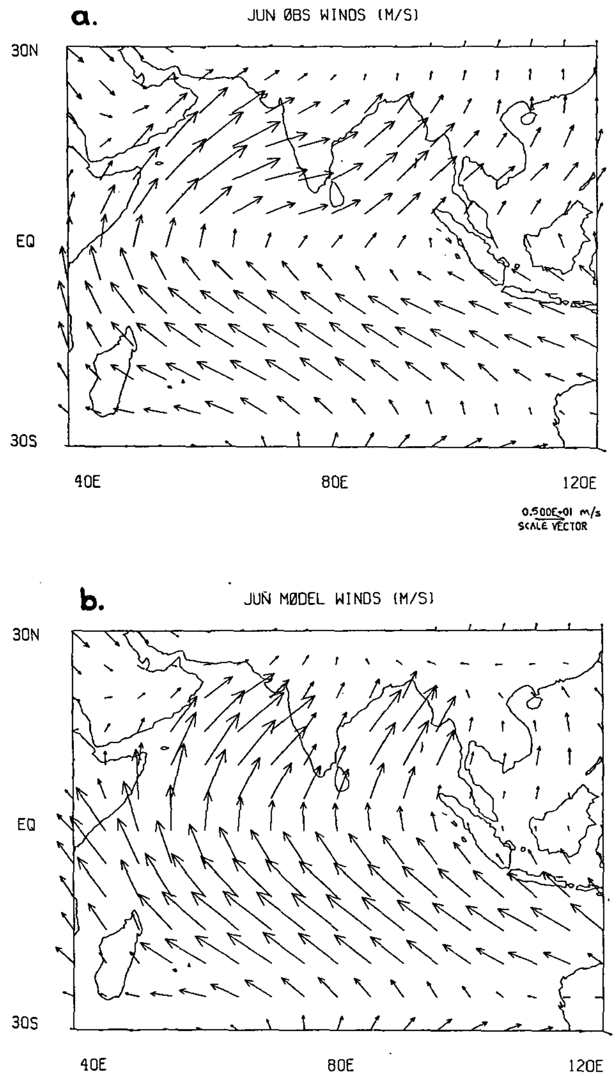

FIG. 9. As in Fig. 7 but for June.

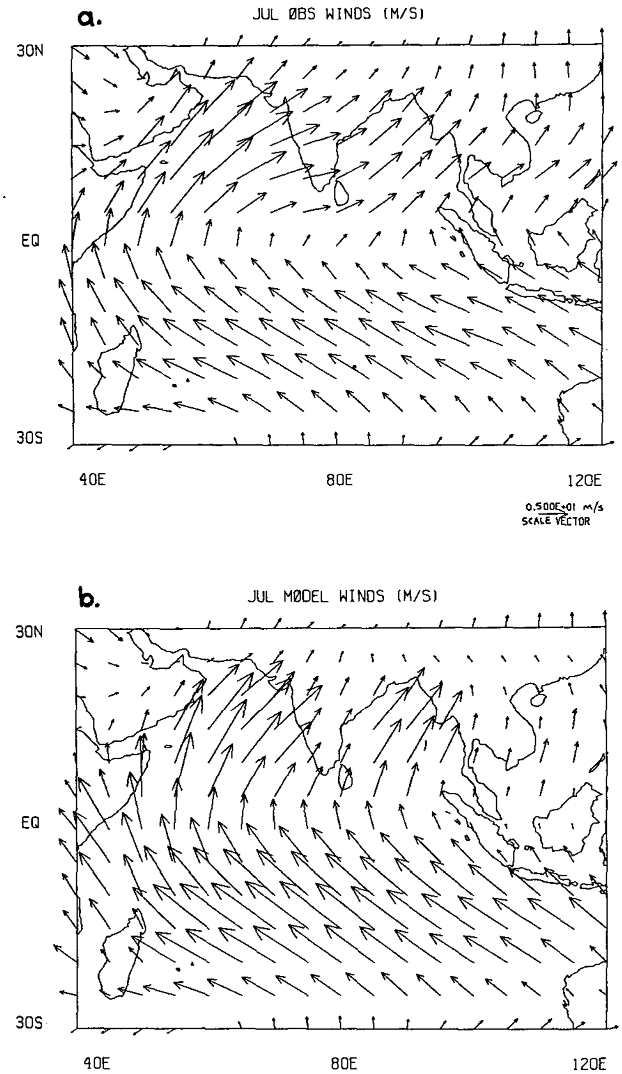

FIG. 10. As in Fig. 7 but for July.

model variation studies are borne out when the studies are made at different locations and times.

The exclusion from (1) and (2) of the nonlinear and dissipation terms gives the simplest balance (geostrophy, case 1, Fig. 5), but also, as seen before, the lowest skill. Taking the geostrophic balance as the lowest skill benchmark, we add terms to find those that are most important in explaining the skill of the standard run. The balance of geostrophy plus advection (case 2) does not converge to a solution. Thus, from a practical point of view, dissipation needs to be included. Geostrophy plus diffusion (case 3, Fig. 5) gives only a slight improvement over geostrophy and mostly near the equator (equatorial improvement not shown in Fig. 5). The same statement can be made about geostrophy plus advection and diffusion (case 4, Fig. 5). Geostrophy plus linear drag (the three-way balance, case 5, Fig. 6) is a large improvement over geostrophy in the entire region. We therefore take the three-way balance as the next skill benchmark and find that the addition of diffusion (case 6, Fig. 6) contributes significantly to the skill over most of the region. The skill improvements are most noticeable in the northeast trades and on the southern flank of the near-equatorial CZ. The addition of nonlinearity to the three way balance (case 7, Fig. 6) also contributes positively to the skill, though to a smaller degree than does diffusion. As with diffusion, the nonlinear effects are most clear in the CZ and the northeast trades. There is some decrease in skill in the southeast trades associated with the inclusion of nonlinearity. The upshot of these case studies is that linear drag appears to be the best first candidate to improve upon the geostrophic balance, with diffusion being an important additional term. The further addition of the nonlinear terms makes a smaller additional improvement in most areas.

The importance near the equator of the diffusion and nonlinear terms is also evident in Table 2, where the absolute values of the various terms in the standard model solution are shown, again averaged over the central and eastern Pacific. The three-way balance is the primary balance, but diffusion and nonlinear advection (especially $v\partial u/\partial y$ and $v\partial v/\partial y$) are clearly nonnegligible terms near the equator. These terms are similarly significant in the equatorial area of other sectors in the model domain, in both the standard model balance and the balance calculated from the observed winds.

To some extent, the diffusion and nonlinear terms seem to substitute for each other, although diffusion does add more to the skill than nonlinearity. The case 6 and 7 curves of Fig. 6 show that both terms improve the three-way balance skill, and both give their biggest contributions in the same regions. But the sum of the improvements when they are added singly to the three way balance (the sum of the case 6 and 7 skill increases) is greater than the improvement given when both are added together to the three-way balance (the standard model skill increase). A similar situation was encountered by Sardeshmukh and Hoskins (1985) who reported that a linear model with unrealistically high friction gave results similar to a nonlinear model with lower friction. Moreover, they found that the transients (as a prescribed forcing) can to some extent substitute for nonlinearity or linear drag. Thus, from their experience and ours, it seems fair to say that the precise effects of transients, nonlinearity, and dissipative terms are difficult to disentangle. Of these three processes, only nonlinearity is explicitly present in (1) and (2). The other two are jointly and crudely parameterized by the linear drag and horizontal diffusion.

This substitution is also indicated by experiments on the model's sensitivity to the dissipation constants. Model runs in which friction, diffusion, or nonlinearity has been removed show a greater sensitivity to the choice of k and D than does the standard model. The standard model shows only a moderate sensitivity to the k and D values when these values are one-half to two times the standard values. The choice of a spatially nonuniform k, in which the land friction is much larger than the ocean friction, as in the standard model, gives markedly better results than does the choice of a uniform k. Experiments with a narrow range of k values show that (1) and (2) give somewhat better u (v) skills if a smaller (larger) than standard k value is used. This indicates that the linear drag represents, in part, processes that operate differentially in the zonal and meridional directions.

Several variations of the boundary conditions were tested. These included setting the boundary winds equal to: (a) zero, (b) the observed winds, (c) the three-way balance winds, and (d) the geostrophic winds. In all these tests, very little change in the interior winds was found; significant changes occurred only within a few grid distances from the boundaries.

6. Regional results

It is also interesting to see the model's simulation of the evolution of a major tropical wind system. In Figs. 7–10, the onset of the Asian summer monsoon over the Indian Ocean is presented in vector plots of the observed and standard model winds for April through July.

In all four months shown, the overall pattern is fairly realistic. The southeasterlies over the southern Indian Ocean are well captured in direction and magnitude; the weakening of the northeasterlies in April and the onset of the southwesterlies in May also match well with the observed winds. Indeed, with the strengthening of the southerly flow after April, the model skills (not shown) improve everywhere. However, the model problems seen in the December case also appear in these regional results. In general, the model's meridional winds are too strong, and the skills are low in areas of weak and convergent winds. The model winds for April through July are excessively easterly over and near East Africa. This seems to be mainly a problem with doubtful input sea level pressures over land. In these months, Oort's (1983) East African sea level pressures are, from Kenya to the Gulf of Aden, two to eight millibars lower than those of other climatologies (e.g., Godbole and Shukla 1981; Shea 1986).

7. Effects of pressure field uncertainty

The above results suggest that tropical surface winds retrieved from observed pressures may be fairly skillful. Thus, tropical sea level pressure fields appear to be important sources of information about the associated winds. However, there is a question as to how sensitive the derived winds are to uncertainties in the pressure field. In particular, do small changes in pressure, representing our uncertainty about the pressure, produce large changes in the derived winds? If so, then one cannot meaningfully retrieve the wind from such pressure fields. As a test of our model, we have examined its response to uncertainties in the input pressure.

It is not trivial to determine what the uncertainty in the pressure really is. But, as a working definition, we simply declare the uncertainty to be the difference between two credible analyses. Figure 11 shows this difference for the January 1987 monthly mean sea level pressure analyses made at the National Meteorological Center (NMC) and at the Fleet Numerical Oceanography Center (FNOC). Over elevated land, the difference is large, since the two analyses use different methods for converting pressures to sea level. Over the oceans, the difference is less than 0.5 millibars in most

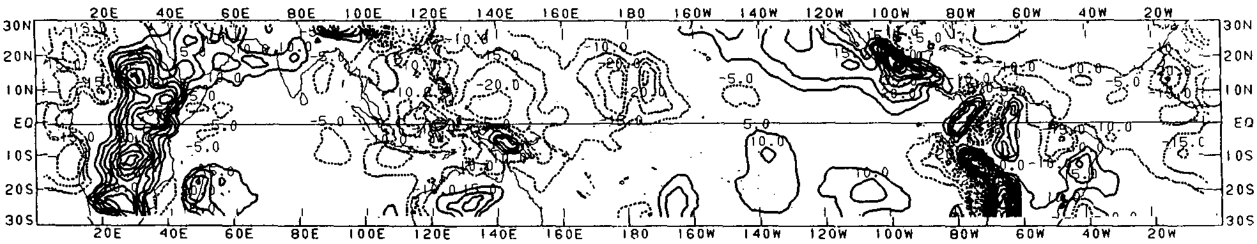

FIG. 11. January 1987 mean sea level pressure difference, NMC − FNOC, in 0.1 mb. Contour interval is 0.5 mb; zero contours omitted; negative contours dashed.

areas, except for some fairly large areas in the western Pacific and eastern Atlantic.

We test the model's response to these differences by running it first with the NMC pressures and then with the FNOC pressures. The differences between the resulting u and v components of the wind are shown in Fig. 12. Neglecting the large differences over and very near the land, we note that the near-equatorial u component suffers much less from pressure field uncertainties than does the v component. As discussed earlier, there are large problems with modeling the near equatorial meridional wind. Table 3 gives, as a function of latitude, the root mean square (rms) difference for the NMC and FNOC pressure, and for the u and v winds from the two model runs, for ocean gridpoints only. The uncertainty in the pressure (of about 0.6 millibars) leads to a typical uncertainty in u and v of about 1.2 and 1.5 m s^{-1}, respectively. For u, the uncertainty is small compared to the rms magnitude of u itself. For v however, the uncertainty is relatively large, as large as v itself between about 5°S and 5°N.

We do not know whether the pressure difference in Fig. 11 gives a realistic impression of pressure uncertainty. The difference does seem larger than one might have expected. This may be partly explained by the inclusion of an interactive ocean in the FNOC analysis procedure. The difference between the NMC and FNOC sea surface temperatures is also remarkably large in some months. Thus, we may be overestimating the pressure uncertainty. However, despite these complications, we conclude that the pressure uncertainty is small enough to make wind retrieval possible. The errors in the derived wind that do arise from this uncertainty seem to be, on the whole, relatively small, al-

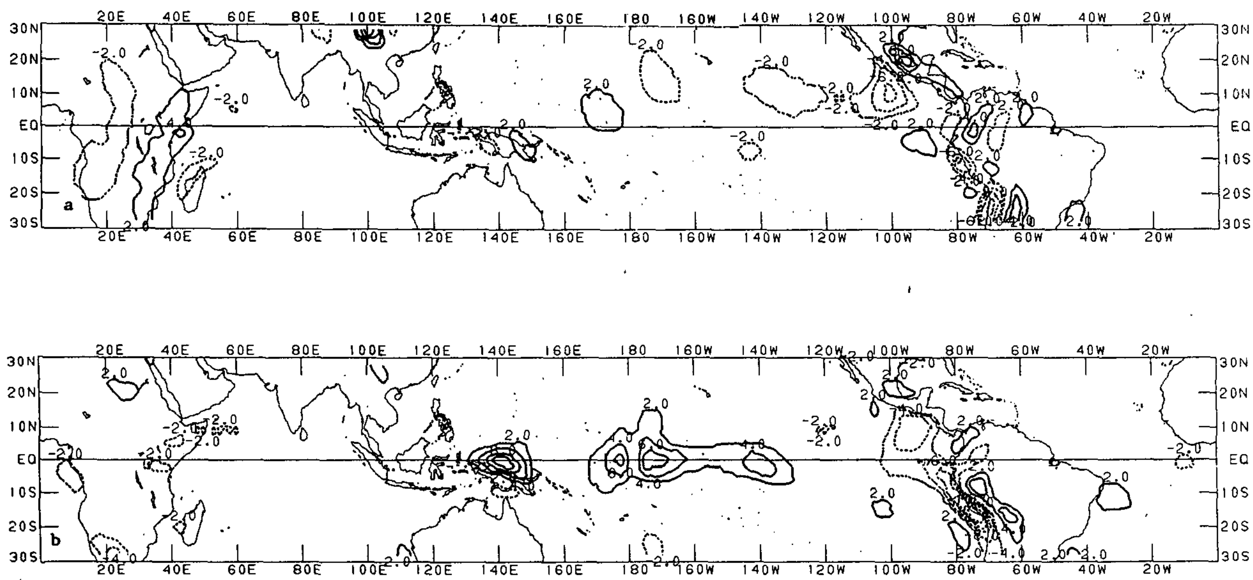

FIG. 12. Differences between January 1987 mean surface winds calculated by standard model in separate runs using January mean sea level pressures from NMC and from FNOC, NMC-based calculations minus FNOC-based calculations: (a) zonal wind difference; (b) meridional wind difference. Contour interval is 2 m s^{-1}; zero contour omitted; negative contours dashed.

TABLE 3. Zonally averaged rms differences, for ocean gridpoints only, between: (i) NMC and FNOC sea level pressure fields, and (ii) the winds calculated by the standard model in separate runs based upon these two pressure fields. Pressure differences in millibars. Wind differences, by zonal and meridional components, in m s^{-1}.

	rms differences		
Latitude	p	u	v
27.5	0.5	0.6	0.5
25.0	0.5	0.7	0.4
22.5	0.5	0.9	0.5
20.0	0.6	1.1	0.6
17.5	0.7	1.2	0.7
15.0	0.9	1.5	1.0
12.5	0.9	1.7	1.3
10.0	0.8	1.9	1.5
7.5	0.8	1.7	1.6
5.0	0.8	1.4	1.8
2.5	0.7	1.3	2.3
0	0.5	1.0	2.7
−2.5	0.4	1.0	2.3
−5.0	0.5	1.0	1.7
−7.5	0.5	1.1	1.3
−10.0	0.6	1.2	1.3
−12.5	0.6	1.1	1.5
−15.0	0.6	1.0	1.2
−17.5	0.5	1.1	1.5
−20.0	0.5	1.0	1.4
−22.5	0.6	0.9	1.0
−25.0	0.5	1.0	1.0
−27.5	0.5	1.0	1.0

though with the potential to be large in the near equatorial v component.

8. Summary and discussion

We have examined the momentum balance for tropical surface winds with a simple numerical model. The model, based on the nonlinear horizontal momentum equations and forced by observed long term monthly mean sea level pressures, calculates surface winds that are verified by comparison with observed surface winds. The model's results are also compared with those generated by simpler momentum balances.

These comparisons suggest several interesting conclusions about the tropical wind field. First, with a relatively simple model forced by observed pressures, realistic simulations of time mean tropical winds, even near-equatorial surface winds, can be obtained. Uncertainties in the observed pressures, although potentially problematic in the calculation of the meridional wind near the equator, are not, overall, a barrier to wind retrieval. Thus the model results indicate that time mean tropical pressure fields are an important source of accurate information about associated tropical winds. This refutes the conventional notion that weak tropical pressure gradients and measurement errors make the pressure field a very poor or even useless tool for calculating tropical winds. Finally, the model's improvements over the three-way balance winds show that diffusion effects and nonlinear advection are important to the time mean tropical momentum balance, especially near the equator.

The physical processes represented by the model's dissipation terms still need clarification, however. In addition, it would be useful to explore further the possible roles of the terms omitted, in explicit form, from (1) and (2). The transient eddy forcing terms are especially interesting in this regard (see section 2). In an earlier observational study of the equatorial momentum balance, Romanov and Romanova (1976) found indications that these terms may be important near the equator. However, the results of Stout and Young (1983) suggest that, equatorward of the subtropics, horizontal momentum transports by synoptic scale transient eddies may be negligible. We have attempted a simple estimate of the transients' relative importance by calculating the horizontal transient forcing from the January 1987 NMC analysis. Using NMC's twice daily winds at 1000 millibars for this period, we have calculated the transient eddy terms

$$\mathrm{FEX} = -\left(\frac{\partial}{\partial x}\overline{u'^2} + \frac{\partial}{\partial y}\overline{u'v'}\right) \tag{7}$$

$$\mathrm{FEY} = -\left(\frac{\partial}{\partial x}\overline{u'v'} + \frac{\partial}{\partial y}\overline{v'^2}\right) \tag{8}$$

where the overbars indicate time means and the primes indicate departures from the time mean. The spatial distribution of FEX and FEY are shown in Fig. 13. The transient terms are largest in the higher latitudes of the domain and smallest near the equator. Near the equator the transient terms are about one-fifth the size of the leading terms and roughly equal in importance to the nonlinear terms of the mean flow. Thus, in the equatorial region, the transient terms would seem to have a small, but non-negligible, role in the overall momentum balance. It is also interesting to notice in Fig. 13b that, poleward of about 15°S and 15°N, the transients tend to act as a brake on the equatorward flow of the trades. This may be a clue for explaining the model's excessively strong equatorward flow (section 4 and Fig. 3c) and the differential response of the u and v components to the choice of k (section 5). We are currently involved in further studies designed to clarify the role of transient eddy and dissipation processes in the tropical surface momentum balance.

The model's ability to give realistic wind simulations suggests several possible applications. The model is likely to be applicable to shorter term mean pressure and wind fields, and, especially, to higher levels where poorly represented interactions with the surface are less important. Also, because the model results show the pressure field to be a reliable source of information about tropical winds, pressure data and the model may be useful as a means of improving the analysis of sur-

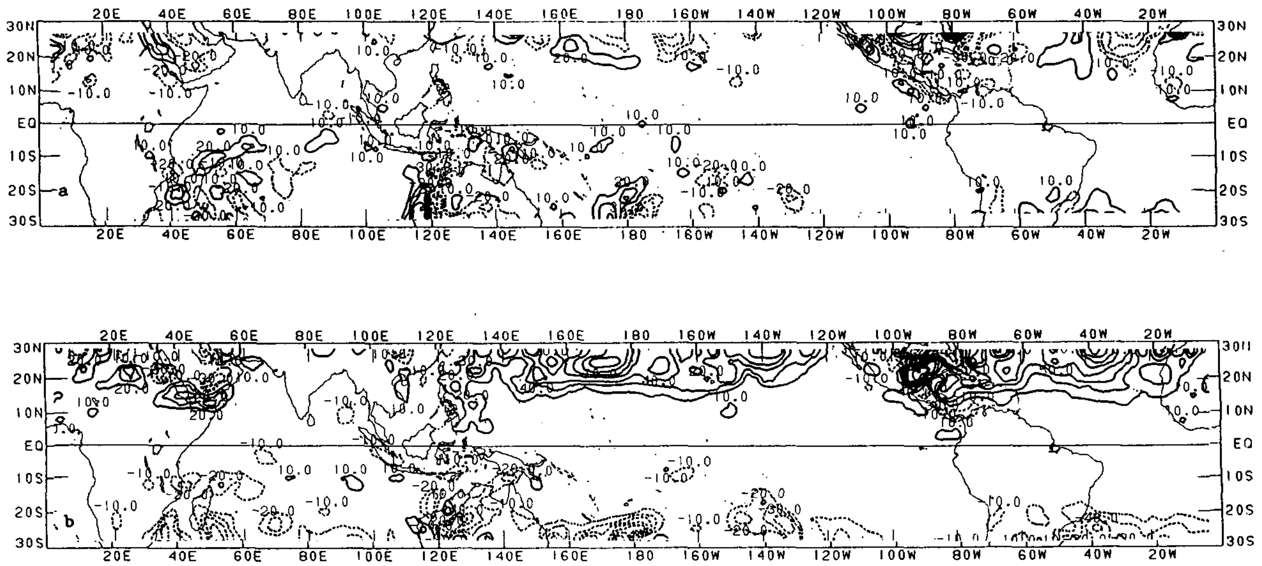

FIG. 13. January 1987 transient eddy forcing, in 10^{-6} m s^{-2}, from NMC 1000 mb wind analysis. Contour interval is 10×10^{-6} m s^{-2}, zero contour omitted: (a) zonal component, FEX; (b) meridional component, FEY. Abundance of positive (negative) contours in NH(SH) in (b) indicates the retarding effect of eddies on the meridional flow.

face wind fields. Thus the model might be useful in estimates of wind stress on the ocean (cf. Hantel 1970). It may also be applied to tests of the consistency between wind and pressure fields; for example, in the initialization of wind driven ocean models. A model of this sort is easily run backwards, using the wind field as the forcing in order to calculate pressures (cf. Saha and Suryanarayana 1971). In the reversed form, the model would provide a convenient means of deriving tropical pressure fields from, say, scatterometer-derived winds. Examples of these applications will be presented in future publications.

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