NOTES AND CORRESPONDENCE

Equatorial Sea Surface Temperature Sensitivity to Net Surface Heat Flux: Some Ocean Circulation Model Results*

D. E. HARRISON

NOAA/PMEL, Seattle, Washington

19 June 1990 and 21 November 1990

ABSTRACT

Several primitive-equation ocean general circulation model experiments have been carried out in order to explore the sensitivity of equatorial sea surface temperature (SST) results to uncertainty in the net surface heat flux (Q) imposed at the surface. Both climatological seasonal cycle experiments and hindcasts of the 1982/83 ENSO event are considered. It is found that regions of light winds, which typically reach values of SST in excess of 31°C using this ocean model and past Q parameterizations, attain more realistic SST values of 29°–30°C when Q is reduced by as little as 10 W m^{-2}. Sensitivity in this regime is about 0.1–0.2 °C (W m^{-2})$^{-1}$ for low-frequency SST changes. In regions of easterly winds with their associated upwelling, horizontal advection, and stronger mixing, changes of Q in excess of 50 W m^{-2} produce SST changes typically of 0.7°C, for a sensitivity of about 0.02°C (W m^{-2})$^{-1}$. These results apply equally well to the ENSO hindcasts and the seasonal cycle studies. The reasons for the large variation in sensitivity and the very large sensitivity under light winds are described. To the extent that these results are representative of oceanic conditions, very accurate Q information will be required for studies of the low-frequency variability of SST in light wind regions like the western Pacific; much less accurate fluxes appear needed for studies of comparable variability in upwelling regions.

1. Introduction

Ocean–atmosphere coupling processes lie at the heart of the El Niño–Southern Oscillation (ENSO) phenomenon which has received so much attention over the past decade. During ENSO periods there is large-scale disruption of the normal patterns of the trade winds, precipitation, sea surface temperature, and ocean currents. The mechanisms of coupling are the focus of much research interest at present. All mechanisms under investigation focus on the roles of sea surface temperature (SST) changes and surface wind changes. Insofar as the atmosphere is concerned, the only oceanic variable of concern is SST. Clearly, our ocean models must be able to produce accurate mean SST values as well as accurate SST changes, given accurate atmospheric forcing information, if we are to understand the behavior of the coupled system.

Sea surface temperature changes occur in the ocean as a result of advective, diffusive, and convective processes. Because the physics of the ocean-surface mixed layer (OML) (and, under some conditions, the skin layer at the top of the OML) is complex, it is generally represented rather schematically in present tropical Pacific ocean models. There are many reasons for this, one of which being that our knowledge of the net surface heat flux between the atmosphere and ocean is so limited, and OML physics so strongly constrained by it. Some models, like the GFDL primitive-equation model used by many tropical oceanographers (thanks to the help of George Philander and Ron Pacanowski), use a Richardson-number-dependent vertical mixing to provide a crude parameterization of OML processes; some others fix the depth of their "mixed layer" and then include many OML processes. Reduced gravity models make the assumption that SST changes are negatively correlated with changes in the depth of the thermocline, and so assume SST changes inversely with upper-layer depth changes.

Our purpose here is not to critique the different approaches to SST modeling. However, some comments are perhaps in order. The reduced gravity models (e.g., Busalacchi and O'Brien 1981) are often able to obtain qualitatively correct behavior on long time scales or for special circumstances (like the passage of downwelling Kelvin waves), though many instances have now been documented in which SST increases while the upper layer thins (and vice versa). The work of the Lamont group, which fixes mixed-layer depth while preserving many other OML processes, has shown that their approach can yield interesting SST results under a variety of circumstances (e.g., Seeger 1989; Seeger et al. 1988). However, mixed-layer depth is known to vary significantly in both the seasonal cycle and during ENSO periods across much of the equatorial Pacific,

* Contribution No. 1206 from NOAA/Pacific Marine Environmental Laboratory.

Corresponding author address: Dr. D. E. Harrison, NOAA/PMEL, 7600 Sand Point Way NE, Seattle, WA 98115.

so the physical correctness of these results remains subject to verification.

The GFDL primitive-equation model with the Pacanowski and Philander (1981) Richardson number mixing parameterization has been found to have some considerable success in hindcasting the large-scale, upper-ocean behavior of the tropical Pacific, at least for the seasonal cycle (Philander et al. 1987), and the 1982/83 ENSO event (Philander and Seigel 1985; Harrison et al. 1989). Qualitatively satisfactory patterns of SST variability have been obtained, but quantitative SST changes have either not been reported or, when reported, are less satisfactory than changes in other upper-ocean variables (Harrison et al. 1989, hereafter HKG).

The 1982/83 ENSO hindcasts of HKG showed that different analyses of the surface wind field over the tropical Pacific produced qualitatively similar large-scale patterns of SST change, but quantitatively very different SST evolution. Because the different results tended to bracket the observed SST values, HKG argued that the primary need for improved ocean model SST validation was improved surface wind stress information. Efforts are underway within the Tropical Ocean–Global Atmosphere (TOGA) program to greatly enhance the number and quality of surface wind and temperature data in the equatorial Pacific, so that one may hope that ENSO events occurring after 1992 will be better observed than their predecessors, and that model studies will be better constrained by having more accurate wind stress forcing information.

However, the role of surface heat fluxes in the evolution of SST in the tropical Pacific is also a large concern at present. Particularly in the central and western Pacific, SST changes are small, both within ENSO and non-ENSO periods. Whenever SST changes are small, the net surface heat flux must be well known if accurate SST variability is to be obtained. The work of Blumenthal and Cane (1989) addresses the sensitivity of SST results to surface heat flux uncertainties in the context of the fixed mixed-layer depth ocean model used by Seager et al. (1988) and Seager (1989), but no comparable results are available for the primitive equation general circulation model.

The purpose of this note is to describe the results of several experiments that have been conducted to investigate the degree of SST sensitivity of the GFDL ocean general circulation model system to net surface heat flux uncertainty, using the form of Richardson number mixing that has been widely used. Results from simulations of the seasonal cycle and of the 1982/83 ENSO period will be presented.

2. The ocean model and surface heat flux parameterizations

The ocean circulation model used in these experiments is described by Philander and Seigel (1985), and used in the 1982/83 ENSO studies reported by HKG (1988), Harrison et al. (1990), and Harrison (1989). It is described fully in Philander and Seigel (1985). Briefly, the model has a nonuniform grid in depth, with 27 levels spanning 5000 m, but with 10-m separation in the first 100 m in order to better resolve upper-ocean processes. The horizontal resolution is 1° in longitude and 0.33° in latitude within 10° of the equator (the region of interest here). The standard Arakawa *B*-grid second-order differencing is followed. Old-style Bryan and Cox convective adjustment on density is permitted. Surface wind stress is a specified function of location and time—monthly mean surface wind stresses are linearly interpolated in time.

In nature the net surface heat flux is the sum of the shortwave, longwave, and sensible and latent heat flux components. The net flux is frequently much smaller than the largest components. Each of these components varies with changing air–sea parameter changes. In the tropics the monthly mean shortwave and latent components can each vary by more than 100 W m^{-2} depending upon location, and whether ENSO or non-ENSO conditions prevail. The sensible and longwave components vary by smaller amounts typically, although the variability of the longwave component in the central equatorial Pacific may approach 50 W m^{-2} under extreme conditions. At present we estimate each component by use of bulk formulae. Very few direct flux measurements exist in the tropics to permit us to assess the relevance of bulk formulas derived from midlatitude conditions. Also, there is limited data available on cloudiness and cloud type as well as near-surface humidity, which are needed to estimate the latent and shortwave components. Thus ocean modelers have typically heavily parameterized the net surface heat flux for their model calculations.

The surface heat flux parameterization used here is described by Philander and Seigel (1985). Both SST and surface wind speed affect the surface heat flux, and either an arbitrary air–sea temperature difference or an atmospheric "reference" temperature must be specified. This parameterization is crude; although latent and sensible heat flux variability are incorporated, solar and net longwave variability are not. Because the objective here is to explore the sensitivity of SST results in this model to changes in net heat flux, the physical shortcomings of this parameterization are immaterial; it gets SST near to observations so we can look at sensitivity about "normal" conditions.

Despite the fact that a minimum wind speed is imposed to mimic the unresolved high-frequency variability, all studies known to the author tend to have unrealistically warm SST values under sustained light monthly mean wind conditions. In particular the 1982/83 hindcasts reported by HKG typically produced maximum Pacific SSTs in excess of 30°C at the peak of the ENSO event; in the cases with the weakest wind stress some values in excess of 31°C were found. Here we explore the effect of introducing the assumption that the net radiative component of the surface heat flux declines as SST exceeds 27°C, due to cloudiness effects. In the interest of simplicity we take the Philander and Seigel parameterization value and diminish it with a linear ramp in SST, so that (in W m^{-2}):

$$Q = Q(\mathrm{PS}) - 13.3(\mathrm{SST} - 27) \quad \text{for} \quad \mathrm{SST} > 27°\mathrm{C}. \tag{1a}$$

It was also clear in HKG that minimum temperatures in the cold tongue, under conditions of climatological easterlies, tended to be colder than observed. So, in this instance without any physical motivation, we arbitrarily ramp up the heat flux from its Philander and Seigel value whenever SST is colder than 26°C:

$$Q = Q(\mathrm{PS}) + 20(26 - \mathrm{SST}) \quad \text{for} \quad \mathrm{SST} < 26°\mathrm{C}. \tag{1b}$$

When 26° > SST > 25°C, we keep the Philander and Seigel heat flux value. For convenience we shall refer to the parameterization described by Eq. (1) as RAMP2. All standard cases will be with the unmodified Philander and Seigel (1985) net flux, which shall be denoted by PS.

Again, please note that these flux changes are arbitrary (albeit motivated to improve the results of previous experiments), but this arbitrariness in no way lessens their utility for this study. The goal is to explore SST sensitivity to Q variations in this model. Any change in Q would suffice for this purpose. These changes result in Q differences that are usefully large, and take the model results toward observations.

3. 1982/83 ENSO hindcast equatorial SST

Several of the 1982/83 ENSO hindcasts reported by HKG were repeated with the RAMP2 heat flux parameterization. Here are reported results from runs using the National Meteorological Center (NMC) 1000-mb winds (converted to stress by Philander and Seigel) and using the Florida State University (FSU) pseudostresses evaluated by O'Brien and coworkers (converted to stress using air density of 1.25×10^{-3} g cm^{-3} and a constant drag coefficient of 1.2×10^{-3}). The NMC case had among the weakest equatorial stresses for this period, while the FSU case had strong equatorial stresses. In each case the resulting RAMP2 surface currents barely differed from the corresponding reference case (rms differences less than 10 cm s^{-1}), so attention shall be focused on SST. The reader interested in how these results compare with observations may consult HKG.

Figure 1 shows equatorial longitude–time plots of 30-day running averages of the NMC SST results using the original heat flux parameterization (1a) and the difference in SST between the original experiment and the RAMP2 experiment (1b). Figure 2 shows equatorial longitude–time plots of 30-day running averages of the net surface heat flux in the original experiment (2a) and the difference in net surface heat flux between the original run and the RAMP2 run. The very warm western Pacific water (SST > 31°C) in the original case is clear in Fig. 1a, as is its expansion all the way across the basin late in 1982. Figure 1b shows that the maximum temperatures obtained with the RAMP2 parameterization are between 29° and 30°C. Figure 2a shows that the high original experiment SSTs are obtained with net surface heat fluxes of typically 20–30 W m^{-2}. This is possible because the winds are light. There is net heat flux in the ocean and there is no ocean dynamical process acting to counter these factors that produce a very thin surface layer. Basically the only process available to diminish the SST is the model vertical mixing out of the shallow surface layer. Figure 2b shows that a reduction in the net heat flux of only about 10 W m^{-2} suffices to bring about the 2°–3°C reduction in SST. While there is still net heating of the surface, the model vertical mixing evidently is able to maintain a more reasonable SST value at this level of heating.

In the equatorial cold tongue, which is present through midsummer 1982 at most longitudes in the NMC experiments, there is little change in SST between the two experiments. Only in the coldest part of the cold tongue do the differences exceed 1°C, and differences smaller than 0.5°C predominate. The maximum surface heat flux differences are about 50 W m^{-2} and are in the coldest part of the cold tongue.

Figures 3 and 4 present comparable results from the FSU experiments. Because the FSU stresses are larger than the NMC stresses, the maximum SST values in the original FSU experiment are smaller, typically 30°–31°C. The RAMP2 parameterization again reduces the maximum temperatures to between 29° and 30°C, and the reduction in net surface heating is again about 10 W m^{-2}. The equatorial cold tongue results are similar to those in the two NMC cases—SST differences are typically 0.5°C, maximum differences are a bit greater than 1°C, and the maximum surface heat flux change is 60–70 W m^{-2}.

4. Seasonal cycle experiments

The seasonal cycle experiment of Philander et al. (1987) is the starting point for this sensitivity study. Figure 5 presents 30-day running average equatorial SST from the rerun of this experiment (Fig. 5a), together with the difference in SST between this experiment and the identical experiment run with RAMP2 surface fluxes (Fig. 5b). In the warm western Pacific, waters warmer than 29°C in the basic experiment are cooled by 0.5° to 1.0°C with RAMP2 fluxes. In the spring warming in the cold tongue, RAMP2 fluxes warm typically by only 0.2° to 0.5°C; when the cold tongue is fully formed in September–October, the maximum warming is about 1.0°C.

Figure 6 shows the surface fluxes from the basic ex-

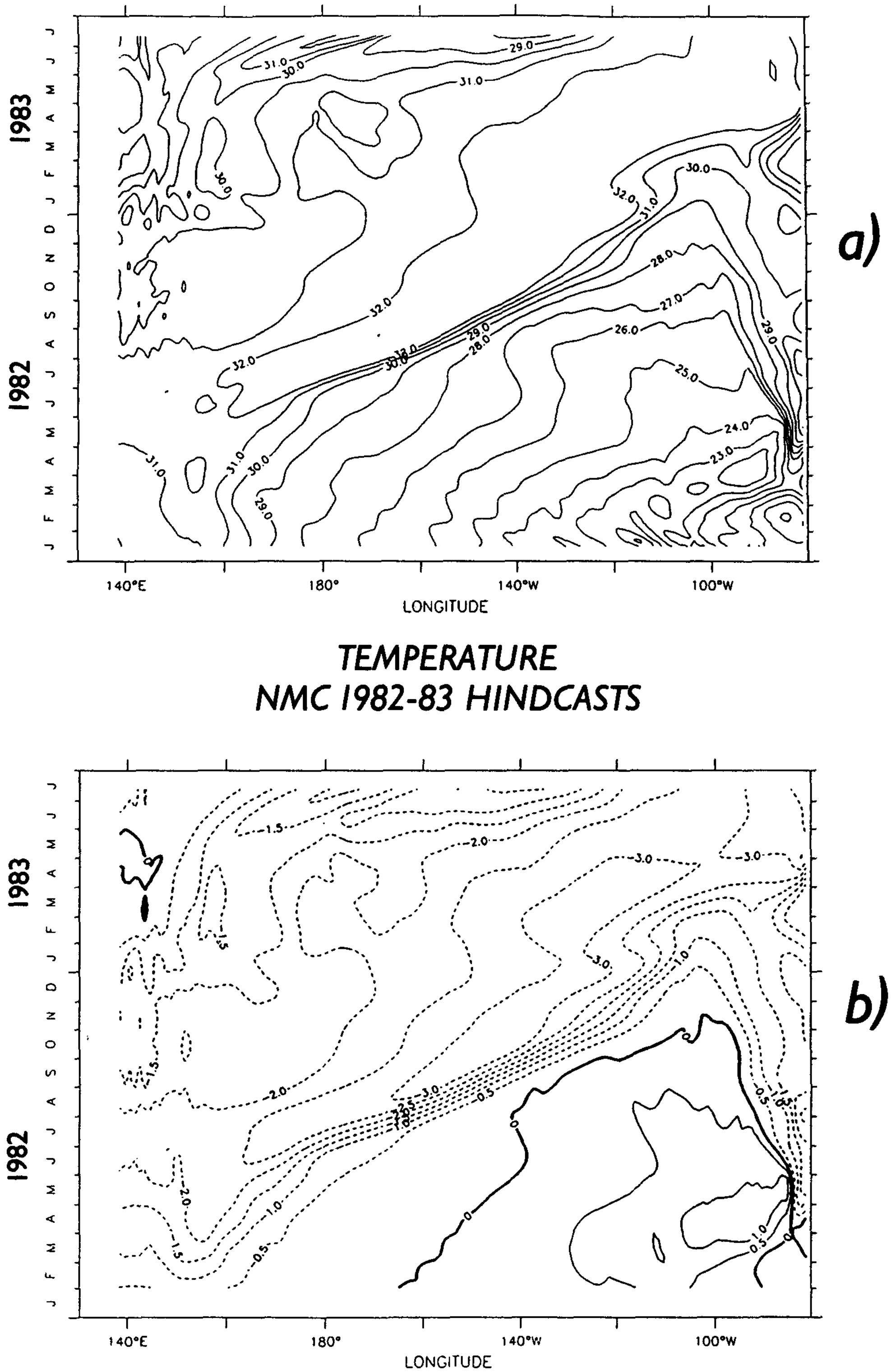

FIG. 1. Thirty-day running average sea surface temperature results from 1982/83 El Niño hindcast experiments. (a) SST from an experiment identical to that of Philander and Seigel (1985). (b) The difference in SST between the Philander and Seigel experiment and an experiment identical except for use of the RAMP2 surface heat flux modification (see text).

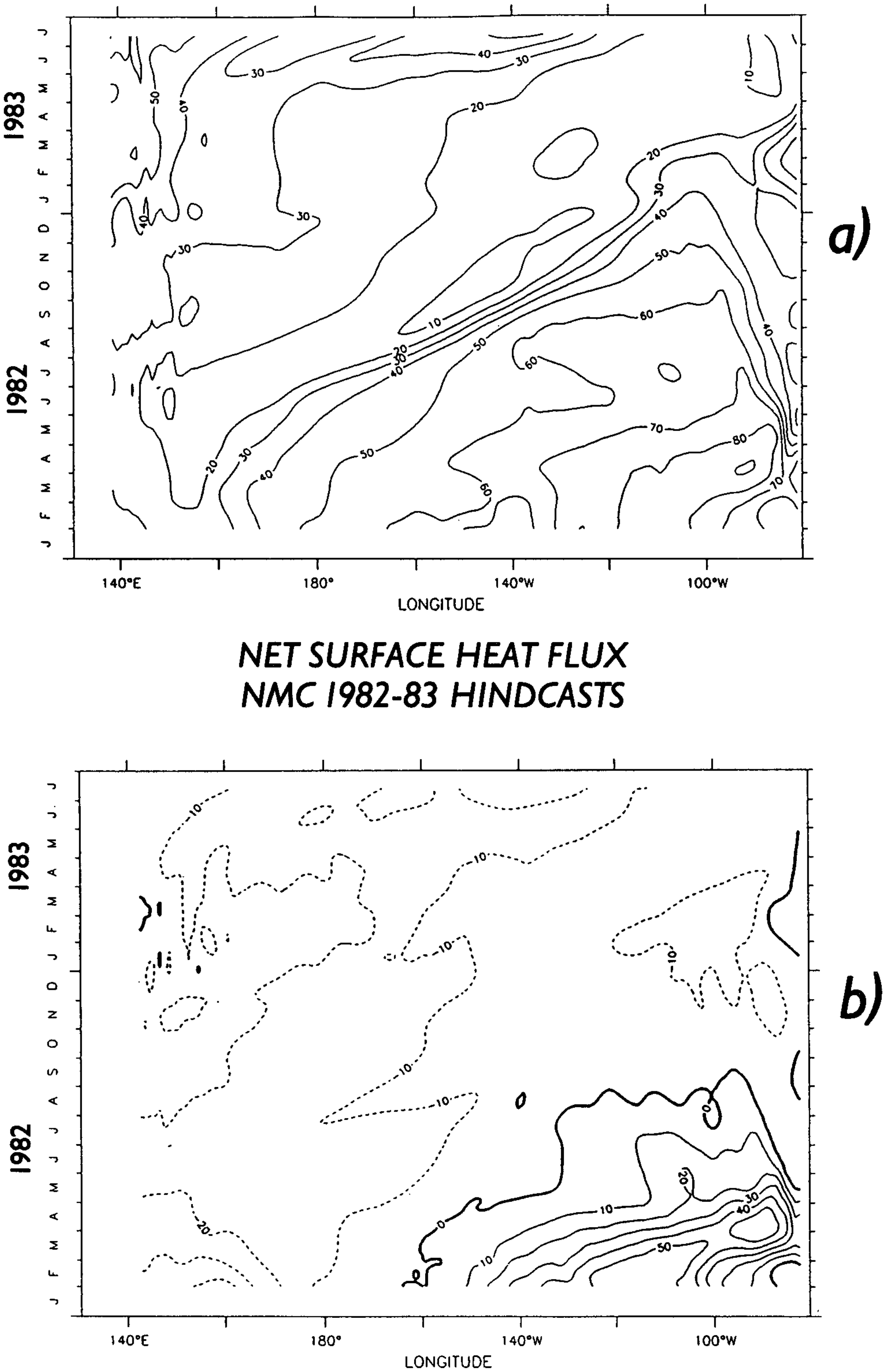

FIG. 2. Thirty-day running average net surface heat flux results (a) from the Philander and Seigel experiment. (b) The difference in net surface heat flux between the original experiment and the one using the RAMP2 flux.

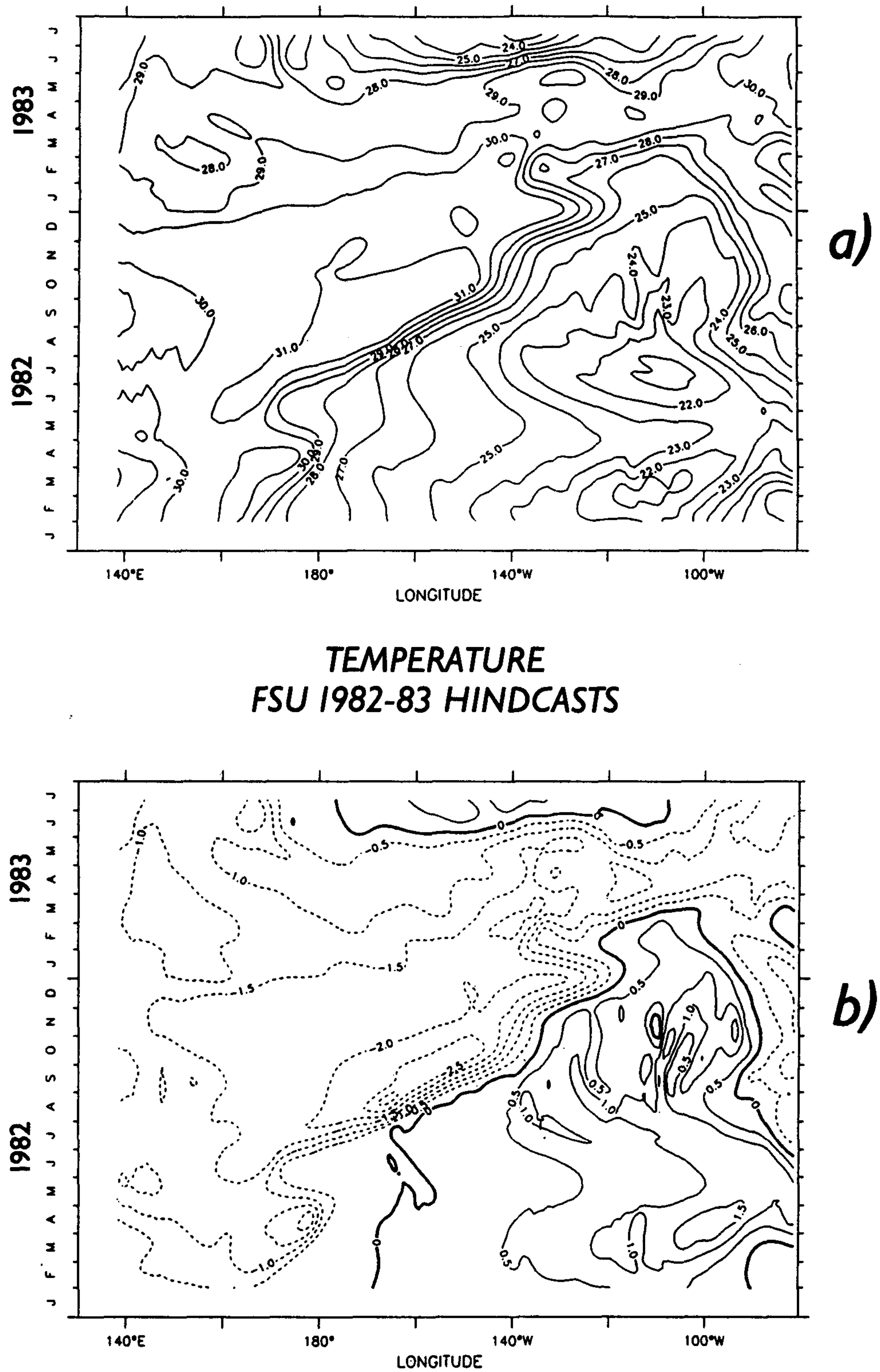

FIG. 3. SST results, as in Fig. 1, except for 1982/83 El Niño hindcast experiments using the FSU surface wind stress fields with the Philander and Seigel net surface heat flux parameterization and with the RAMP2 modification (see text).

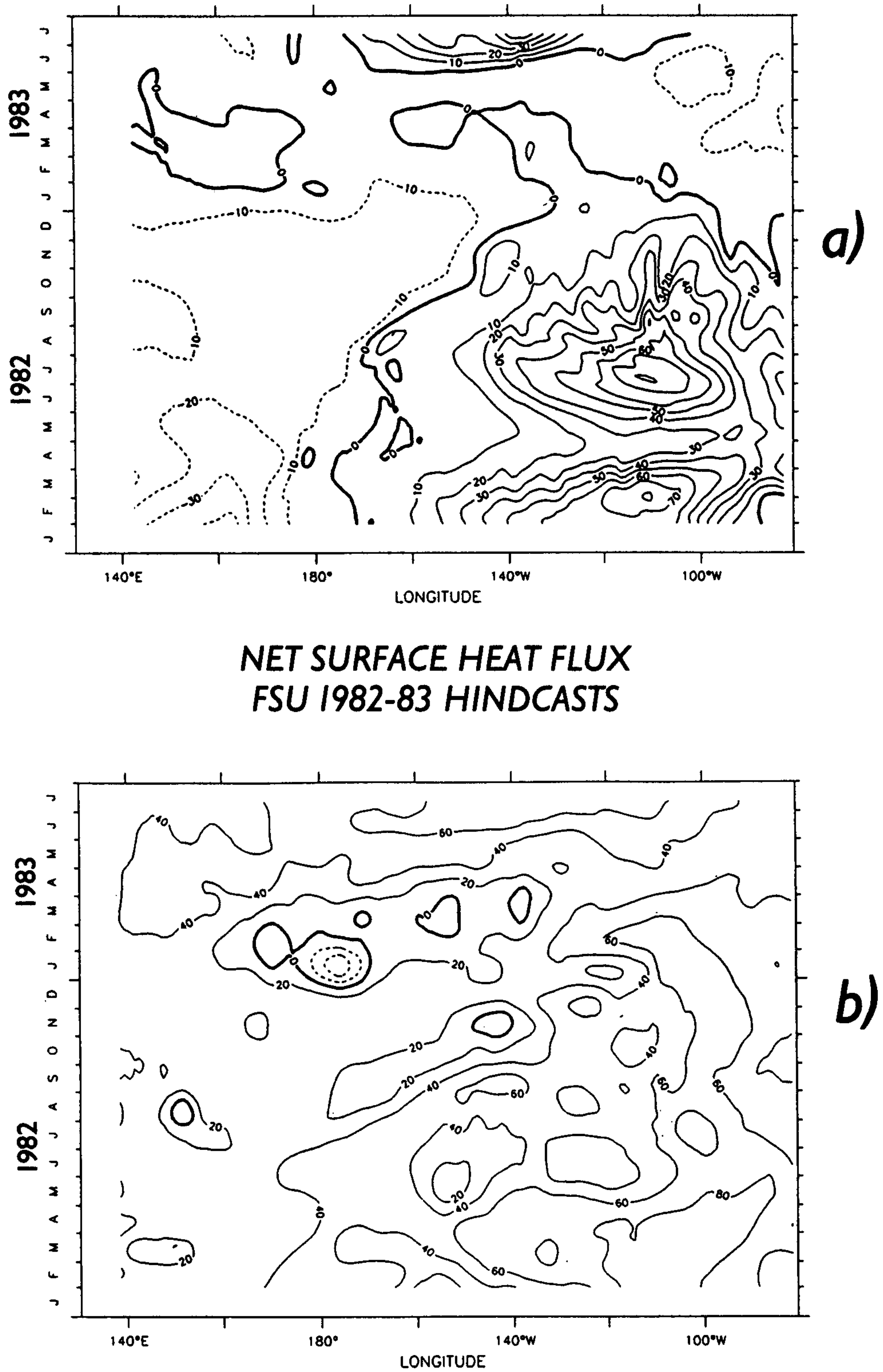

FIG. 4. Surface heat flux results, as in Fig. 2, except for the 1982/83 El Niño hindcast experiments using the FSU surface wind stress fields of Fig. 3 (see text).

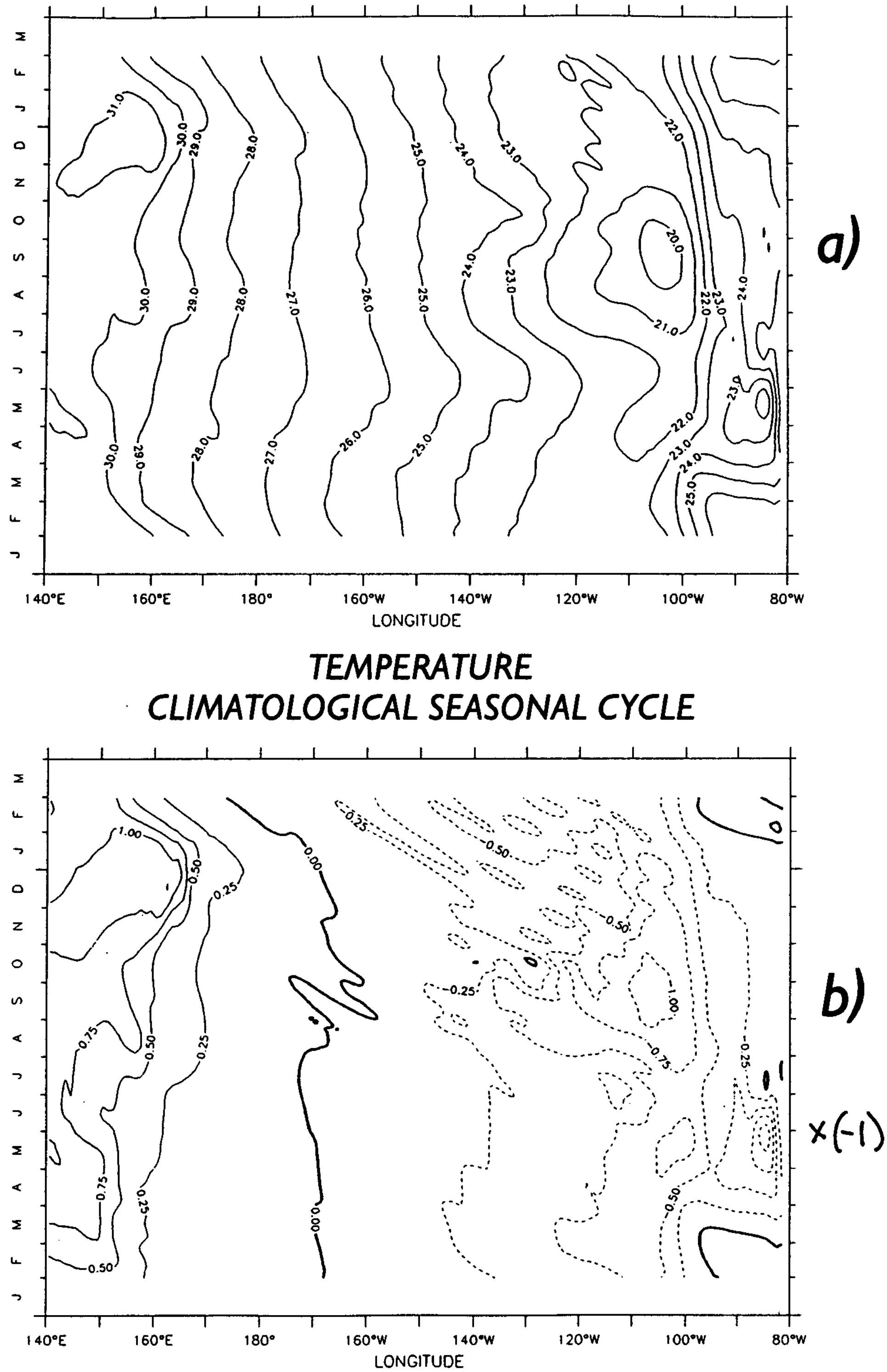

FIG. 5. SST results, as in Fig. 1, except seasonal cycle experiments using climatological monthly mean surface wind stresses and surface heat flux, as in Philander et al. (1987), and modified with the RAMP2 change (see text).

periment, and the flux differences between it and the RAMP2 experiment. Again we find that the warm water flux differences are quite small, typically 10 W m^{-2} or less. In the cold tongue in the warm season, flux differences are typically 30–50 W m^{-2}; in the cold season they are perhaps 50–70 W m^{-2}.

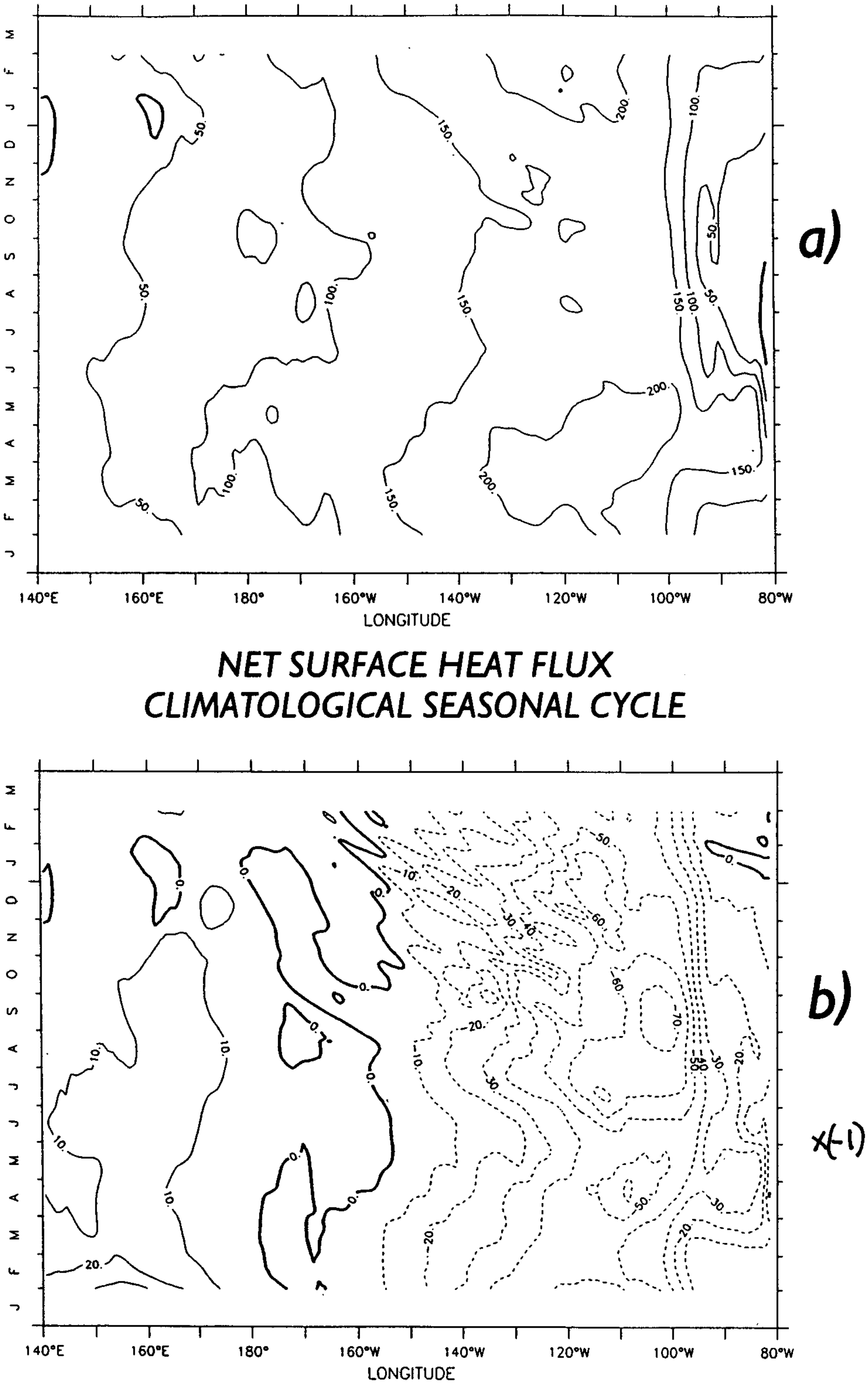

FIG. 6. Surface heat flux results, as in Fig. 2, except for the seasonal cycle experiments of Fig. 5 (see text).

Note that the net surface heat flux into the western Pacific is larger in these seasonal cycle experiments; this is because the Hellerman and Rosenstein wind stresses are larger in the western Pacific than the FSU stresses, so that the surface layer heat budget typically involves both more mixing and more upwelling than when the stresses are weaker. The net result is that western Pacific SST values of 29°–30°C are obtained with net surface fluxes of about 30–50 W m^{-2}. This example illustrates the need to consider both wind stress and Q in mean SST studies. However, in sensitivity studies like these, the mean fluxes are of no special interest.

Sensitivity in these seasonal cycle experiments is similar to that found in the 1982/83 hindcast experiments.

5. Summary and discussion

Several experiments, concerning model SST sensitivity on month-to-seasonal time scales to surface heat flux variations with the primitive equation ocean circulation model of Philander and Seigel (1985) and a modified surface heat flux parameterization, show that reducing the net surface heat flux by 10 to 20 W m^{-2}, under light wind conditions, reduces the model-predicted equatorial SST values from unrealistically high maximum values (in excess of 31°C) to more realistic values between 29° and 30°C. This is true for hindcasts of the 1982/83 ENSO event using two different surface wind stress analyses, and for simulations of the seasonal cycle using the Hellerman and Rosenstein climatological surface wind stresses. Thus the model exhibits very strong SST sensitivity to small changes in net surface heat flux—on the order of 0.2 °C (W m^{-2})$^{-1}$—in regions of warm water and weak winds.

This very large sensitivity results from the fact that the effective model upper-ocean mixed layer is quite shallow ($\leq$20 m) under light wind conditions and with SST values typical of the western Pacific. Because the layer is shallow and because there is little upwelling, horizontal advection, or horizontal diffusion, the heat put in at the surface basically has to be balanced by a vertical diffusive heat flux downward. With the present parameterization of vertical mixing, very little heat can be placed in the ocean if realistic light wind SST values are to be obtained.

The situation is quite different whenever there is easterly wind stress, with any significant upwelling and wind-driven mixing, as is generally the case in the equatorial Pacific east of the Dateline and west of about 90°W. In this oceanic regime there is strong upwelling into the surface layer and strong Ekman meridional divergence because of the prevailing easterly winds. Further, because the upwelling is confined rather narrowly to the equator, there are strong meridional temperature gradients. Thus the surface layer heat equation balance involves horizontal and vertical advections as well as horizontal diffusion (though it is not a dominant term in the balance). The surface heat flux no longer controls SST as powerfully as it does in the light wind regimes, and we find a much reduced SST sensitivity to surface heat flux—roughly 0.02 °C (W m^{-2})$^{-1}$ in the 1982/83 hindcast experiments and as little as 0.01 °C (W m^{-2})$^{-1}$ in the climatological experiments.

The climatological experiment sensitivity is so weak in the cold tongue because it has very strong equatorial easterly stress, and very strong upwelling and Ekman divergence. Thus the net surface heating is of even less importance in these equatorial SST budgets. Other climatological seasonal cycle experiments have been carried out using the Harrison (1989) climatological wind stresses that are weaker by roughly 20% than those of Hellerman and Rosenstein. These experiments simulate the observed seasonal cycle of SST substantially better than the ones reported here, and have marginally larger SST/surface heat flux sensitivity because of the weaker easterly wind stress. These latter seasonal cycle results will be presented and compared with observations and their dynamics will be discussed in another manuscript.

Observations from the western Pacific suggest that the very strong model sensitivity there, as well as the need to have small net fluxes in the annual mean in order to get reasonable SST values, may be physically reasonable. In particular, Godfrey and Lindstrom (1989) suggest that less than 10 W m^{-2} appears to be as much heat as can be mixed downward on the basis of their estimates of vertical mixing. Further, they find that the actual layer of nearly homogeneous density typically is very shallow.

Seager et al. (1989) describe a simplified model for the mean and seasonal cycle of tropical Pacific SST that, with a fixed mixed-layer depth of 50 m, produces quite reasonable agreement with the rather spatially smoothed SST climatology of Rasmussen and Carpenter (1982). Their Fig. 12 indicates model SST sensitivity results of about 0.07 °C (W m^{-2})$^{-1}$ in the western equatorial Pacific and about 0.04 °C (W m^{-2})$^{-1}$ in the eastern/central equatorial Pacific. Their western Pacific sensitivity is smaller than reported here; this difference almost surely arises from the fact that the circulation model effective mixed-layer depth is shallower under light wind conditions (perhaps 20 m) than their fixed 50-m depth layer. In the equatorial cold tongue the circulation model sensitivity is much less than the Seager et al. As a result we speculate that this is due largely to substantially greater upwelling on and near the equator in the circulation model. The circulation model sensitivity results span a significantly broader range than their results for a simpler model. At this time it is difficult to assess the correctness of either the GCM or Seeger results.

The maintenance and evolution of the western Pacific warm pool is of great interest to ENSO researchers at present. If the model results reported here are indeed representative of the SST sensitivity of the ocean in this region on monthly to seasonal time scales, then extremely accurate surface heat flux data will be required in order to explore the thermal processes of the pool.

On the other hand, if the model results for the equatorial cold tongue are representative, variations in the SST of the equatorial waters are not controlled primarily by surface heat flux variations, and it may be possible to gain improved understanding of SST change processes there, with significantly less accurate net heat flux variability information than is needed in the western Pacific.

Acknowledgments. All model runs were done on the NIST CYBER 205 computer. This work is possible only because of the support from George Philander and Ron Pacanowski, who have made their model available and helped in its adaptation to the NIST facilities. The TMAP group at PMEL (Steve Hawkin, Jerry Davison & Kevin O'Brien) provided all local computing assistance. Financial support for TMAP is provided by EPOCS, PMEL and NASA (project UPN 161-25-35-04).

REFERENCES

Blumenthal, M. B., and M. A. Cane, 1989: Accounting for parameter uncertainties in model verification: An illustration with tropical sea surface temperature. *J. Phys. Oceanogr.*, **19,** 815–830.

Blusalacchi, A. J., and J. J. O'Brien, 1981: Interannual variability of the equatorial Pacific in the 1960's. *J. Geophys. Res.*, **86,** 10 901–10 907.

Godfrey, J. S., and E. J. Lindstrom, 1989: The heat budget of the equatorial western Pacific surface mixed layer. *J. Geophys. Res.*, **94,** 8007–8017.

Harrison, D. E., 1989a: Local and remote forcing of ENSO ocean waveguide response. *J. Phys. Oceanogr.*, **19,** 691–695.

——, 1989b: On climatological monthly mean wind stress and wind stress curl fields over the world ocean. *J. Climate,* **2,** 57–70.

——, W. S. Kessler and B. S. Giese, 1989: Hindcasts of the 1982–83 El Niño: thermal and dynamic height variability along the ship of opportunity XBT tracks. *J. Phys. Oceanogr.*, **19,** 397–418.

——, B. S. Giese and E. S. Sarachik, 1990: Mechanisms of SST change in the equatorial waveguide during the 1982–83 ENSO. *J. Climate,* **3,** 173–188.

Pacanowski, R., and S. G. H. Philander, 1981: Parameterization of vertical mixing in numerical models of tropical oceans. *J. Phys. Oceanogr.*, **11,** 1443–1451.

Philander, S. G. H., and A. Seigel, 1985: Simulation of El Niño of 1982–83. *Coupled Ocean-Atmosphere Models,* J. C. J. Nihoul, Ed., Elsevier, 517–541.

——, W. Hurlin and A. D. Seigel, 1987: A model of the seasonal cycle in the tropical Pacific ocean. *J. Phys. Oceanogr.*, **17,** 1986–2002.

Rasmussen, E. M., and T. Carpenter, 1982: Variations in tropical sea surface temperature and surface wind fields associated with the Southern Oscillation. *Mon. Wea. Rev.*, **110,** 354–384.

Seager, R., 1989: Modeling tropical Pacific sea surface temperature: 1970–87. *J. Phys. Oceanogr.*, **19,** 419–434.

——, S. E. Zebiak and M. A. Cane, 1988: A model of the tropical Pacific sea surface temperature climatology. *J. Geophys. Res.*, **93,** 1265–1280.