

Modeling Tropical Pacific Sea Surface Temperature: 1970–87*

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ABSTRACT

An attempt is made to model the sea surface temperature (SST) of the tropical Pacific Ocean between January 1970 and August 1987. The SST is computed using the model and heat flux parameterization described in the climatological study of Seager et al. The model is forced with the observed winds as given by the Florida State University analysis.

The results indicate that while short period variability in the model and observations is largely uncorrelated, variability with periods greater than about one year is well represented in the model. Each El Niño that occurred in this time period is captured by the model. The main discrepancies in the evolution of the 1982/83 and 1986/87 model El Niño events are the inability of the model equatorial SST anomaly to cool in early 1983 and the disappearance of the 1986/87 anomaly during 1987.

The results suggest a number of conclusions. In the east Pacific equatorial Kelvin waves, excited by variations in the trade wind strength in the central and west Pacific, increase the SST via depression of the thermocline. In this region the surface heat flux acts as a negative feedback on the SST anomaly. However, in the central Pacific the surface heat flux anomalies are reinforcing heating through suppression of latent heat loss as a result of weakened trades. Zonal advection of warm water from the west, associated with Rossby waves excited by trade wind relaxation, contributes to warming in both the central and west Pacific. Anomalous cooling by entrainment is a negative feedback on the SST anomaly in the central Pacific.

1. Introduction

Considerable interannual variability has been observed in the region of the tropical Pacific Ocean. In the atmosphere this is inextricably linked to the Southern Oscillation (e.g., Rasmusson and Wallace 1983) and in the ocean to the occurrence of El Niño (e.g., Cane 1983). Though these phenomena are two aspects of a coupled system, in this paper we will consider only the interannual variability of the ocean in response to prescribed atmospheric forcing. The period considered covers January 1970 to September 1987.

The first explanations for the warming of sea surface temperature (SST) in the eastern Pacific observed during El Niño invoked a cessation of the local upwelling responsible for the strength of the equatorial cold tongue. From observations, Wyrtki (1975) was able to demonstrate that there was no systematic weakening of winds in the east Pacific and coastal regions that preceded the warming. He suggested that a relaxation of the trade winds in the central Pacific was responsible. The communication to the east was then provided by an equatorial Kelvin wave that effected a net transfer of warm water eastward. The inflow of warm water

from the west depresses the thermocline (raises sea level) in the east and displaces the cool subsurface water. Water upwelled towards the surface is then warmer and the SST increases.

McCreary (1976) and Hurlburt et al. (1976) demonstrated such a mechanism for thermocline displacement in models forced by idealized winds. They also discussed the simultaneous excitement of westward propagating Rossby waves by the trade wind relaxation, and later, by reflection of Kelvin waves at the eastern boundary.

Busalacchi and O'Brien (1981) considerably extended this work by forcing a linear, reduced-gravity, equatorial β -plane model with the observed winds for the period 1961–70. The variability of upper layer depth in their model was related to the excitation of Kelvin and Rossby waves. They attributed the development of the 1965 and 1969 El Niño events to the excitation of downwelling Kelvin waves (i.e., waves which deepen the thermocline as they propagate east) by relaxation of the trades west of the dateline combined with an absence of the normal seasonal strengthening of the southeast trades over the central Pacific. Busalacchi et al. (1983), in a study of the variability during the 1970s, demonstrated that only zonal stress anomalies contribute to the variability of the upper layer depth at the eastern boundary, which is corroborating evidence of the importance of Kelvin waves. Inoue et al. (1987) extended this work through the period 1979–82. McCreary (1976) and Cane (1984) in-

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icated that only stress anomalies within the equatorial waveguide region are important in determining the ocean variability near the equator, that is, in generating El Niño.

The above studies addressed the variability of thermocline depth or sea level and only discussed SST variability through qualitative reasoning. For example, it is argued that depression of the thermocline decreases the vertical gradient of temperature such that water entrained into the surface layers is warmer, resulting in an increase of SST. In this reasoning SST anomalies (SSTAs) are directly related to anomalies in upper layer depth. However, upwelling is limited to near equatorial and some coastal regions so this relationship would not be expected to hold over most of the tropical oceans. Anomalous horizontal advection and surface heating would be expected to contribute, even in the upwelling regions. For example, Gill (1983) and Schopf and Harrison (1983) suggest that anomalous zonal advective warming on the equator plays an important role in El Niño. If surface heating anomalies are important then the atmosphere contributes not only through forcing ocean motions but also through enhancing or suppressing evaporative and sensible cooling and by causing perturbations in the net radiation budget at the surface.

Few attempts have been made to model the SST variability or its climatology. Zebiak and Cane (1987) described, amongst other variables, the SSTA in an El Niño–Southern Oscillation (ENSO) event simulated by a coupled atmosphere–ocean model. The ocean component of that model, on which the model used in this paper is based, contained linear dynamics, fully nonlinear thermodynamics, and a surface heat flux anomaly proportional to the SSTA. Seager et al. (1988) have simulated the tropical Pacific SST climatology using an ocean model dynamically identical to that of Zebiak and Cane (1987) but with the thermodynamics reformulated to compute total rather than anomaly SSTs. Philander and Siegel (1985) forced an ocean GCM with the observed winds for the 1982/83 El Niño and simulated many of the features of the observed SST for that time. However the model was several degrees too warm in the east and central Pacific. They attributed the equatorial SST variability to equatorial Kelvin waves, excited by westerly wind anomalies in the west Pacific, which depress the thermocline in the east, and to anomalous zonal currents that bring warm water from the west. Latif (1987) forced a primitive equation model of the equatorial Pacific with realistic winds for the period 1947–78. The surface heat flux was proportional to the SST in degrees Celsius less 26°C. The correspondence between observed and model SSTAs was good in the west and central Pacific, where local forcing dominated, but less impressive in the east Pacific where both local and remote forcing were important.

The work presented here is an attempt to model the variability since 1970 of the SST in the tropical Pacific. We use the model of Seager et al. (1988), developed to simulate climatological SST, and force it with the observed winds as given by the Florida State University (FSU) analysis (Goldenberg and O'Brien 1981). The goal of this work is twofold. First, we wish to discover if a simple linear dynamical model, but with nonlinear thermodynamics, capable of simulating climatological SST with reasonable accuracy, can also simulate interannual variability. Second, if that is the case—and we will argue that it is, albeit with exceptions—the mechanisms altering SST in the model will be located and analyzed, with the hope of gaining insight into processes operating in nature. This will be attempted by focusing on the El Niño events of 1982/83 and 1986/87.

The following section describes the model, section 3 provides an overview of the model and observed variability between January 1970 and September 1987, section 4 discusses the mechanisms behind the SST variability in the model during the 1982/83 and 1986/87 El Niños. Discussion and conclusions follow in section 5.

2. Model

The model is discussed in detail by Seager et al. (1988). It describes the linear dynamics of a homogeneous upper layer overlying a motionless deep layer, on an equatorial β -plane, subject to a low frequency approximation. A constant 50 m deep, frictional surface layer is added to capture the surface intensification of the wind driven circulation. This layer is also linear but only because we use Rayleigh friction as a crude approximation to nonlinear effects. The pressure gradient in the surface layer is influenced only by variations in the thermocline depth and neglects the effects of any temperature variation in the layer. These latter effects are usually, but not always, negligible. The ocean basin is rectangular covering 29.75°S to 29.75°N, 124°W to 85°E.

The thermodynamics are, however, fully nonlinear. The surface temperature is determined by horizontal advection, entrainment into the surface layer (which equals the vertical velocity at the base of a constant depth surface layer), diffusion, and the surface heat flux. The temperature of entrained water is parameterized in terms of the thermocline depth using a relationship between the temperature at 50 m depth and thermocline depth derived from the data of Levitus (1982). The surface heat flux includes a formulation for the solar flux as a function of time, noon solar altitude, latitude and cloud cover, as given in Weare et al. (1980). For the full length of the simulation we use climatological annual mean cloud cover as given by

Weare et al. The latent heating term includes the effect of wind speed but avoids specification of air temperature and humidity by assuming a constant humidity factor which multiplies the saturation humidity evaluated at the sea surface temperature. Sensible and long wave heating are combined and made proportional to the quantity $(T_s - T^*)$ where T_s is the SST and T^* is, somewhat arbitrarily, taken to be 273.15 K.

Details of the model, the method of solution, and parameter sensitivity along with the results of a simulation of the tropical Pacific SST climatology are presented in Seager et al. (1988).

The model is forced with the total FSU monthly winds beginning in January 1964 and run through to mid-September 1987. For the purpose of this work the winds have been smoothed in time and space as described in Cane et al. (1986). At the beginning of the run the model ocean is motionless and a uniform 25°C. We allow six years for the model to spin up and begin analysis of the results in January 1970. All the anomalies in model fields discussed in the paper (temperature, terms in the heat budget, and currents) are relative to a model climatology computed over the period January 1970 to September 1987. In contrast, the observed temperature anomaly is taken from the analysis of cast and surface marine temperatures described by Reynolds (1988) and is relative to a long term climatology. The basin mean temperature of the tropical Pacific in the former averaging period is a few tenths of a degree warmer than in the longer period.

3. Overview of model SST for the period January 1970 to September 1987

Figures 1a–d show time series of model and observed temperature anomalies (SSTAs) averaged over the four ocean regions NINO1, EQ1, EQ2, and EQ3. NINO1 covers the South American coastal region 10° to 5°S, 90° to 80°W. EQ1, EQ2 and EQ3 are all equatorial regions spanning 5°S to 5°N and, respectively, 130° to 90°W, 170° to 130°W, and 150°E to 170°W. These regions were chosen because of the different dynamical regimes they represent.

It is clear that the model captures the El Niño events of 1972/73, 1976, 1982/83, and 1986/87 with a distinct peak in SSTA in NINO1, EQ1, and EQ2. The model also captures the long term variability of a cold period between 1972 and 1976, a period of near climatological SST between 1976 and 1982, and a tendency in EQ1 and EQ2 towards a cold period between 1983 and 1986. These longer term trends are clearly apparent in EQ3 where they overwhelm the signatures associated with El Niño events. Variability with periods shorter than the El Niño timescale is largely uncorrelated between the model and observations. The greater high frequency variability in the observations is explicable in terms of

the smoothing that has been applied to the winds used to drive the model plus the low frequency approximation in the model dynamics.

Computation of the coherence and phase between the time series indicated significant coherence at the 95% level was achieved in all regions for variability with periods longer than about 13 months. With the partial exception of EQ1, the phase in this frequency range is small indicating the ability of the model to capture the timing of this low frequency variability. The coherence was low and the phase highly variable for the higher frequency variability in all four ocean regions.

Looking more closely, in the coastal region NINO1 (Fig. 1a), the El Niño of 1972/73 is represented by a brief peak in SSTA of the correct amplitude in late 1972. The observed SSTA is much broader, remaining above 1°C from mid-1972 to spring 1973. Such continued presence of warm water is prevented in the model by the appearance of southeasterly wind anomalies in August 1972, which persist to March 1973. These easterlies are also the reason why in EQ1 the model fails to perpetuate a warming into early 1973. In EQ2 the wind anomalies remain westerly until spring 1973, and hence the model SSTA remains positive to mid-1973.

In NINO1 and EQ1 the 1976 El Niño is simulated reasonably well except for a spurious peak in SSTA in late 1975 and early 1976. This peak is caused by local westerly anomalies (northwesterly at the coast) that allow warming in spite of the presence of easterly anomalies in the central and western equatorial Pacific. The emergence of westerly anomalies in the central and west Pacific in mid-1976 causes the overly strong, broad peak of SSTA in 1976 in EQ2. It is not clear whether errors in the wind field or model failings are responsible. The second peak in late 1977 is caused by a similar burst of strong westerly anomalies. In between 1977 and 1982 the model and observed anomalies bear little resemblance to each other in any of the four regions.

Figures 2a and 2b show model and observed SSTAs for the season December, January and February (DJF) 1982/83 and Fig. 3 presents these for the period June, July and August (JJA) 1983. From Fig. 3 it is clear that the model provides a reasonable simulation of this stage of the El Niño with two major differences. First, the maximum SSTA in the model is a little cool and displaced 20° west of the observed maximum such that it is more disconnected from the coastal warming than it should be. Second and less importantly, the model fails to simulate the cooling of greater than 1°C at the dateline between 10° and 20°N. The results of Seager et al. (1988) indicate this to be a region where dynamics contribute little to the SST so we expect this failure to be a consequence of errors in the surface heat flux. Nonetheless the essential feature of a large magnitude warming centered on the equator in the central and

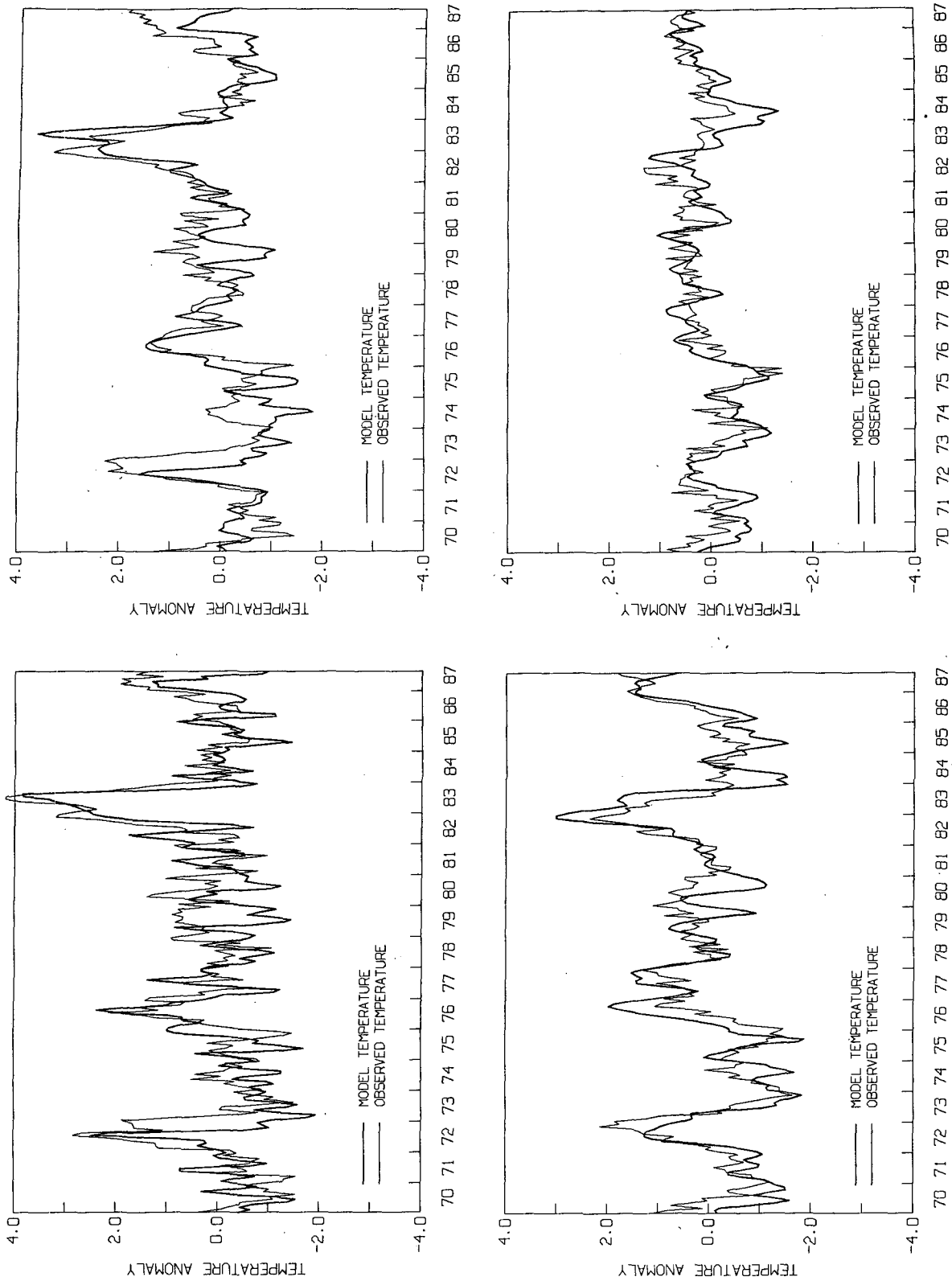


FIG. 1. Simulated and observed temperatures for the ocean regions (a) NINO1, (b) EQ1, (c) EQ2 and (d) EQ3.

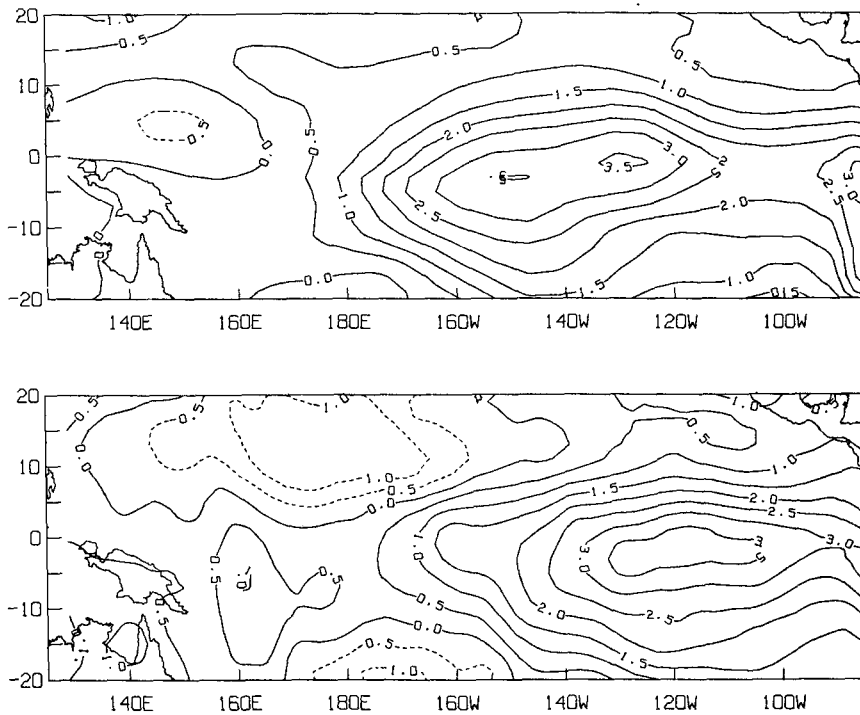


FIG. 2. Seasonally averaged temperature anomaly for December, January and February 1982/83, (a) model and (b) observed.

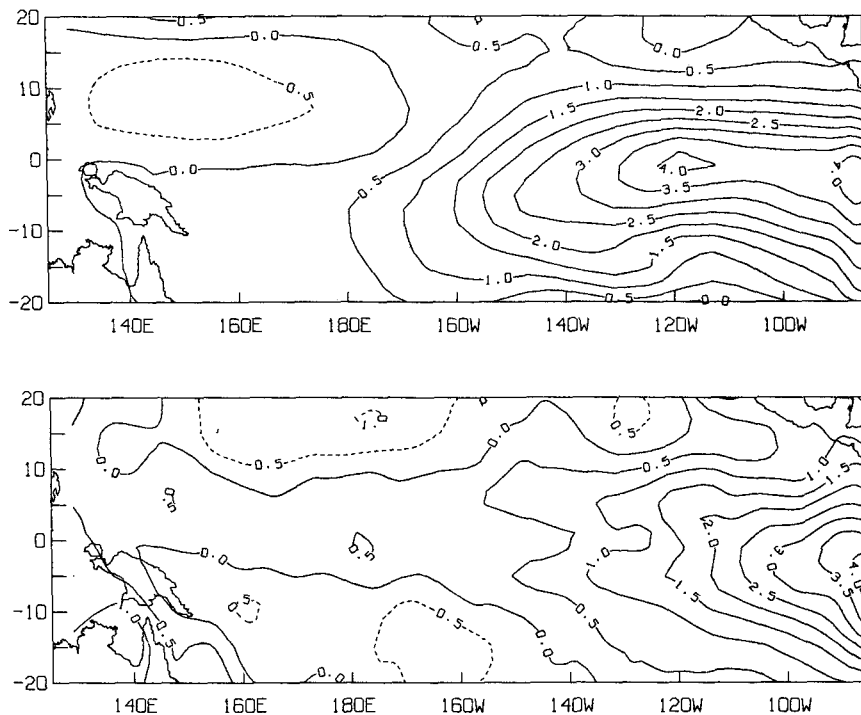


FIG. 3. As in Fig. 5 but for June, July and August 1983.

east Pacific but spanning some 30° of latitude is captured in the simulation.

Disagreement is more severe in JJA 1983 (Fig. 3). At this time the observed SSTA has retreated to the coast where there is a maximum of about 4°C. While the model SSTA has also retreated and is approximately correct at the coast, a 4°C anomaly remains at 120°W—some 2.5°C too warm. Overall the model SSTA has a magnitude that is too large, and too meridionally and zonally extensive.

In late 1986 the model SSTA warms in NINO1, EQ1, and EQ2 but falls again in early 1987 whereas observations indicate that the SSTA continues warm in NINO1 and warms further in EQ1 and EQ2. Figures 4 and 5 present comparisons of model and observed SSTAs for the season DJF 1986/87 and JJA 1987. The agreement is quite good in DJF 1986/87 with both having a 1.5°C anomaly in the central Pacific and a smaller maximum at the coast in the east. However, the model anomaly has a bias towards south of the equator that is not observed. Further, a large area off the Mexican coast has a negative anomaly whereas observations indicate this area to be warm. This is another area which Seager et al. (1988) identified as one where dynamics play a minor role in determining the SST so we expect the problem here to lie with the surface heat flux. This is probably also the reason why the model fails to reproduce the observed cooling between 10° and 20°N west of 160°W. Neglecting temporal cloud cover variability is a likely contender.

The disagreement is severe in JJA 1987. Observations indicate that the anomaly has strengthened to 2°C and is now centered at 130°W. It looks very suggestive of the composite mature phase anomaly of Rasmusson and Carpenter (1982), even if it is occurring some six months too late (or early). However the model retains only a positive anomaly of 1°C centered at 170°W and has gone so far as to produce a negative anomaly in the east. This model cooling is clearly related to the appearance of easterly wind anomalies in the east Pacific in January 1987 and the arrival of reflected upwelling Kelvin waves from the west. Off equatorial anomalies are considerably more in agreement with observations. The evolution of these two later El Niño events is considered in the next section.

3. Mechanisms of SST variability

The total zonal wind stress on the equator during the 1982/83 El Niño is shown in Fig. 6 as a function of time and longitude. Figures 7, 8, and 9 show time-longitude plots of the model thermocline depth, surface zonal current, and temperature anomaly on the equator for this same period. Figures 10 and 11 show model and observed SSTAs, and the terms in the heat budget capable of contributing to temperature anomalies, as a function of time for the ocean regions EQ1 and EQ2.

The corresponding quantities for the 1986/87 period are presented in Figs. 12 to 17.

The mechanism of SST variability were similar for the two El Niño events. In both cases the emergence of westerly wind anomalies in the west and central Pacific (Figs. 6 and 12) excited Kelvin waves that depressed the thermocline in the east and central Pacific (see Figs. 7 and 13). The heat budgets shown in Figs. 10 and 16 indicate the associated changes in vertical advection to be the most important process forcing SSTAs in the east Pacific.

In both events the westerly wind anomalies also excited Rossby waves that contained anomalous eastward surface currents on the equator (Figs. 8 and 14). The anomalous zonal advection of warm water eastward contributed to the warming. The strength of this contribution decreased eastward in line with the decline of the magnitude of the current anomalies. The role of the surface heat flux was also spatially variable. In the east Pacific the increased SST and the persistent strength of local easterly winds allows the surface heat flux to act as a damping on the SSTA (Figs. 10 and 16). This is not the case in the central and western Pacific (Figs. 11 and 17) where weakening of the trades suppresses evaporation and causes a warming. This has been previously noted for the west Pacific by Meyers et al. (1986).

In the central Pacific during the 1982/83 event, vertical advection damps the SST anomaly (Fig. 11b) even though the thermocline here is anomalously deep. The rise in SST which increases the temperature jump across the base of the surface layer and the continuation of upwelling are sufficient to overwhelm the effect of a deepened thermocline. Figures 17a and 17b indicate that during the 1986/87 event the SSTA was of insufficient magnitude for this to happen for any but a brief period in late 1986. Until that time anomalous vertical advection was a significant warming term. For the duration of both events anomalies in meridional advection and diffusion are overwhelmed by the other terms and do not contribute to the character of the SSTA evolution (hence, for the sake of clarity, they are omitted from the heat budget figures).

The 1982/83 event ends suddenly (Fig. 9) when the trades reestablish themselves after July 1983. This is several months later than the observations indicate (Figs. 10a and 11a). For the first half of 1983 the model is responding to trade winds that remained weak and the appearance of westerly anomalies in the east (Fig. 6). Considering the dynamics contained within the model and the fact that the primary source of variability in the surface heat flux is the wind speed, it is not surprising that the model failed to cool through this period.

In contrast to the 1982/83 event the main discrepancy during the 1986/87 event was that the model cooled throughout 1987 (Fig. 15) whereas observations indicated continued warming. The data of Behringer

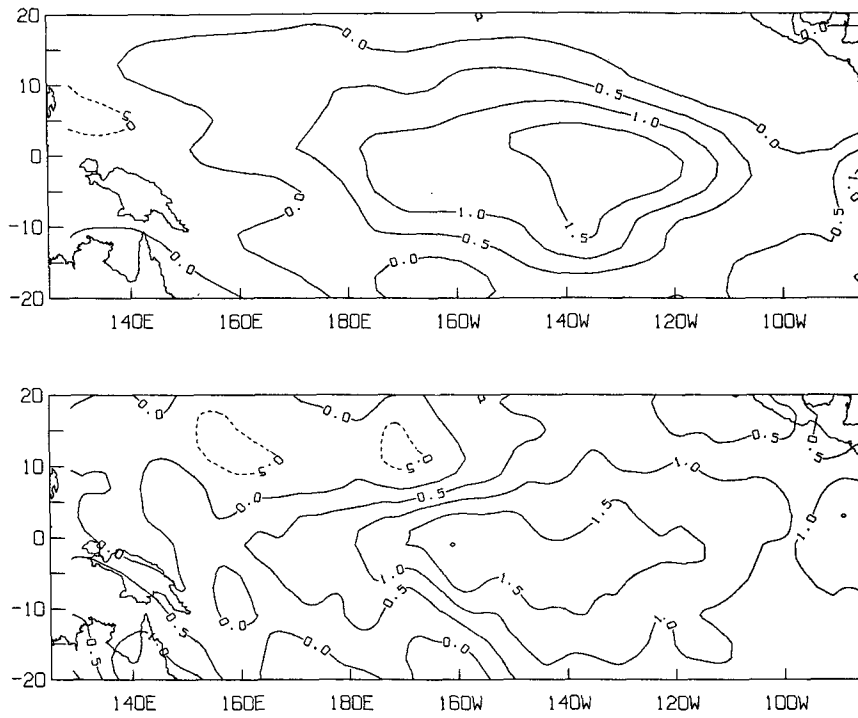


FIG. 4. Seasonally averaged temperature anomaly for December, January and February, 1986/87 for (a) model and (b) observed.

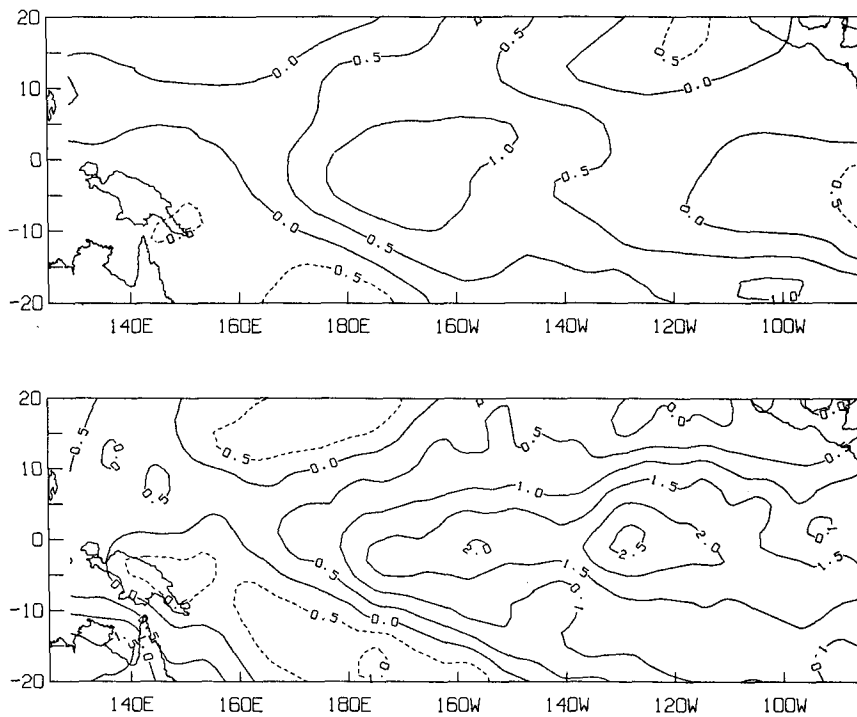


FIG. 5. As in Fig. 7 but for June, July and August 1987.

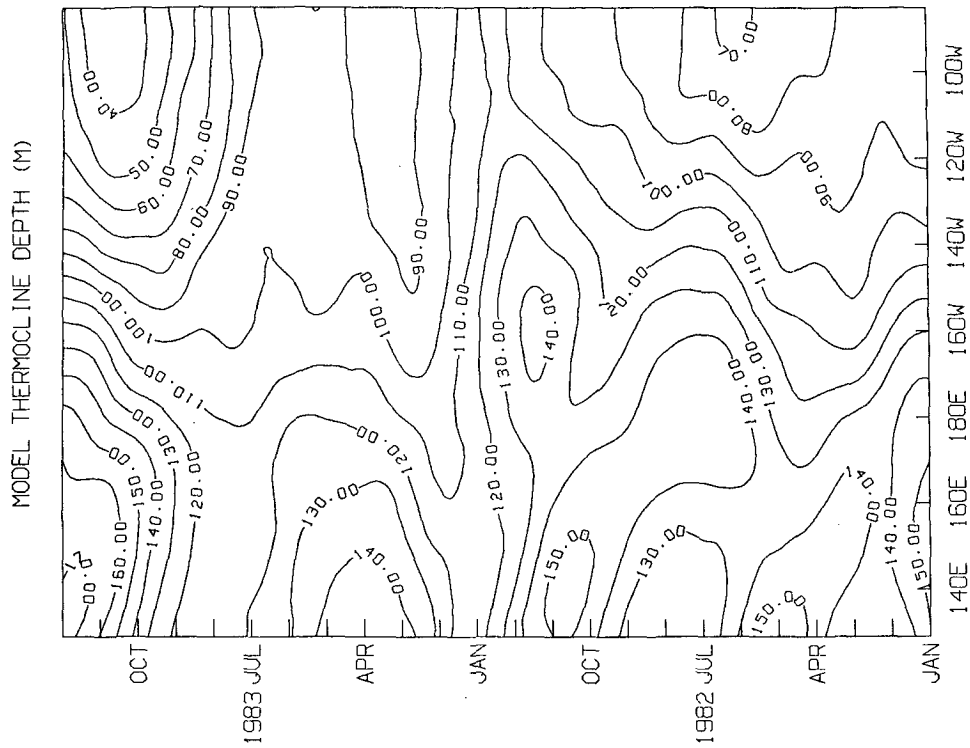


FIG. 7. As in Fig. 6 but for model thermocline depth.

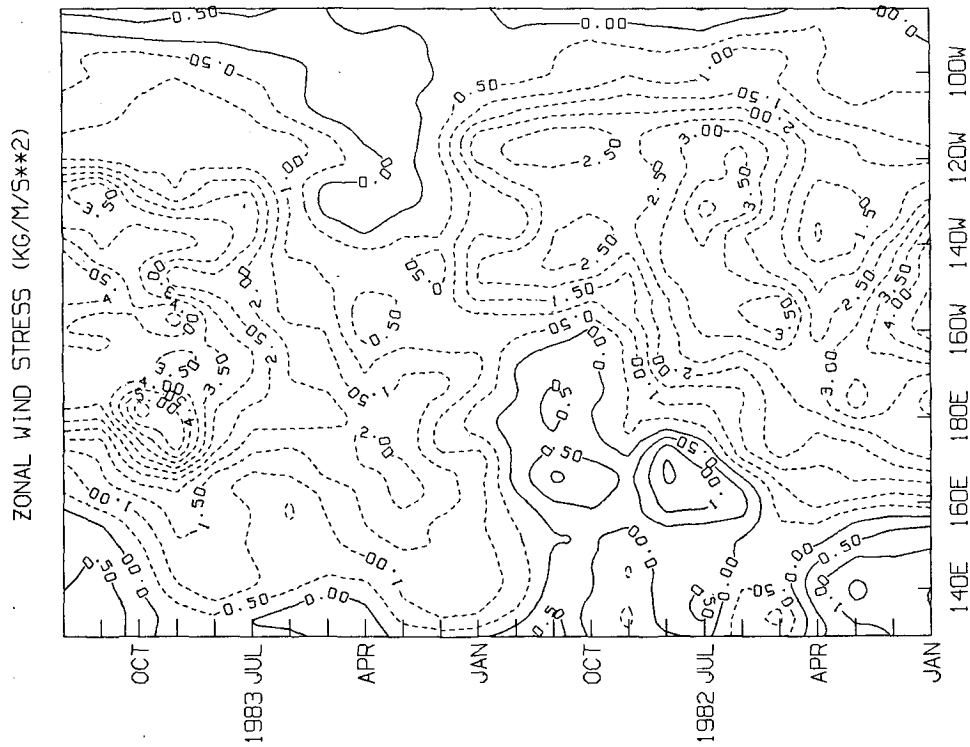


FIG. 6. Zonal wind stress on the equator as a function of longitude and time during the 1982/83 El Niño.

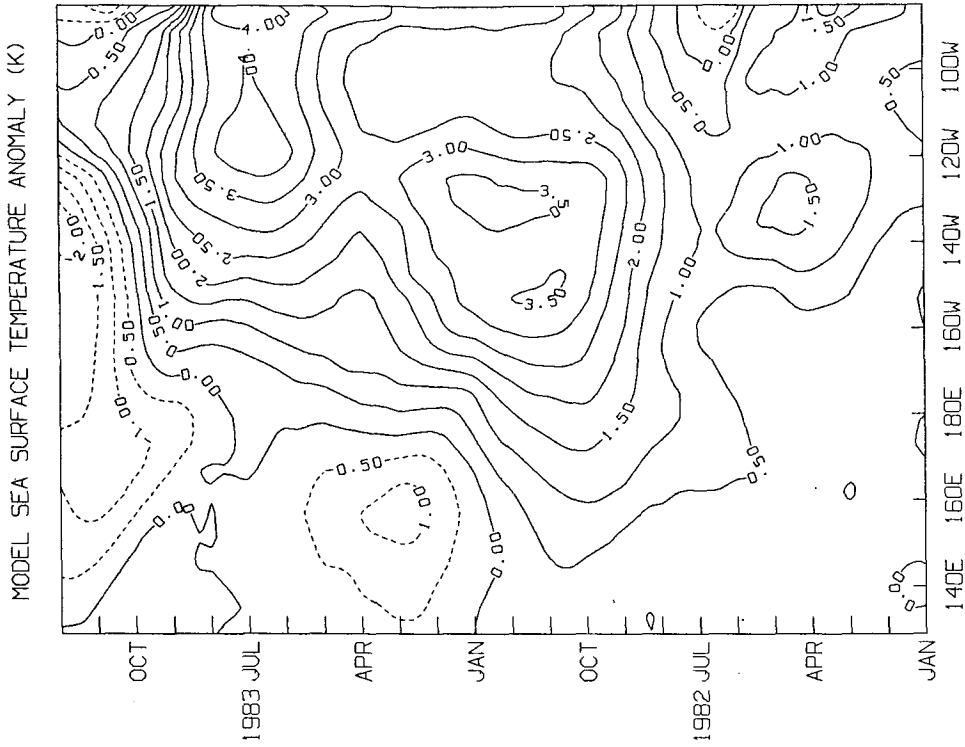


FIG. 9. As in Fig. 6 but for model surface temperature anomaly.

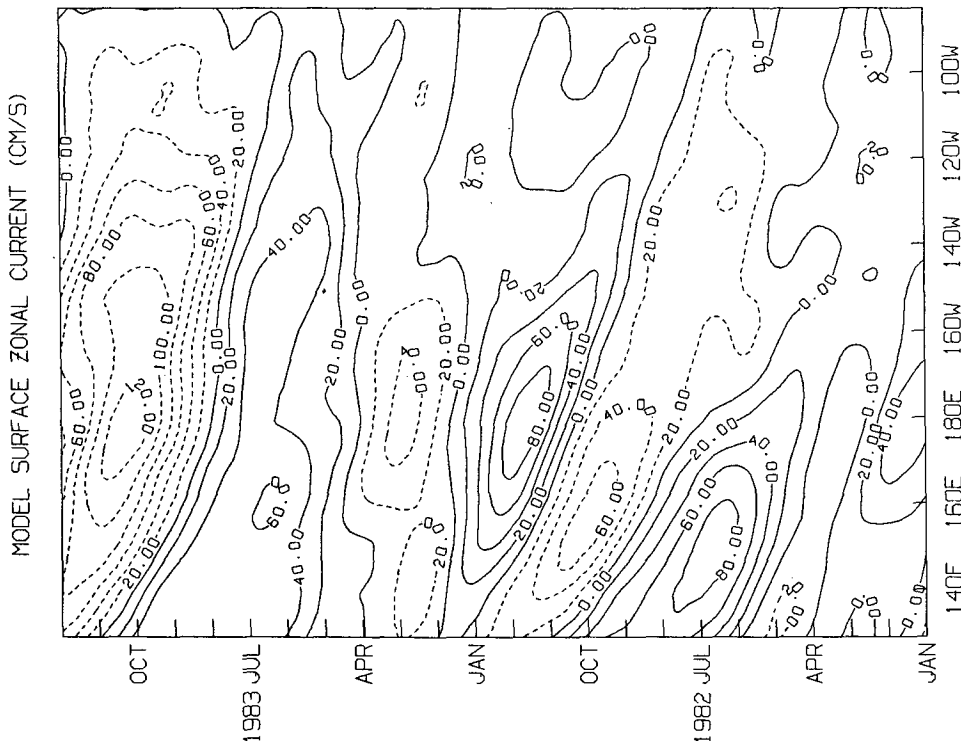


FIG. 8. As in Fig. 6 but for model surface layer zonal current.

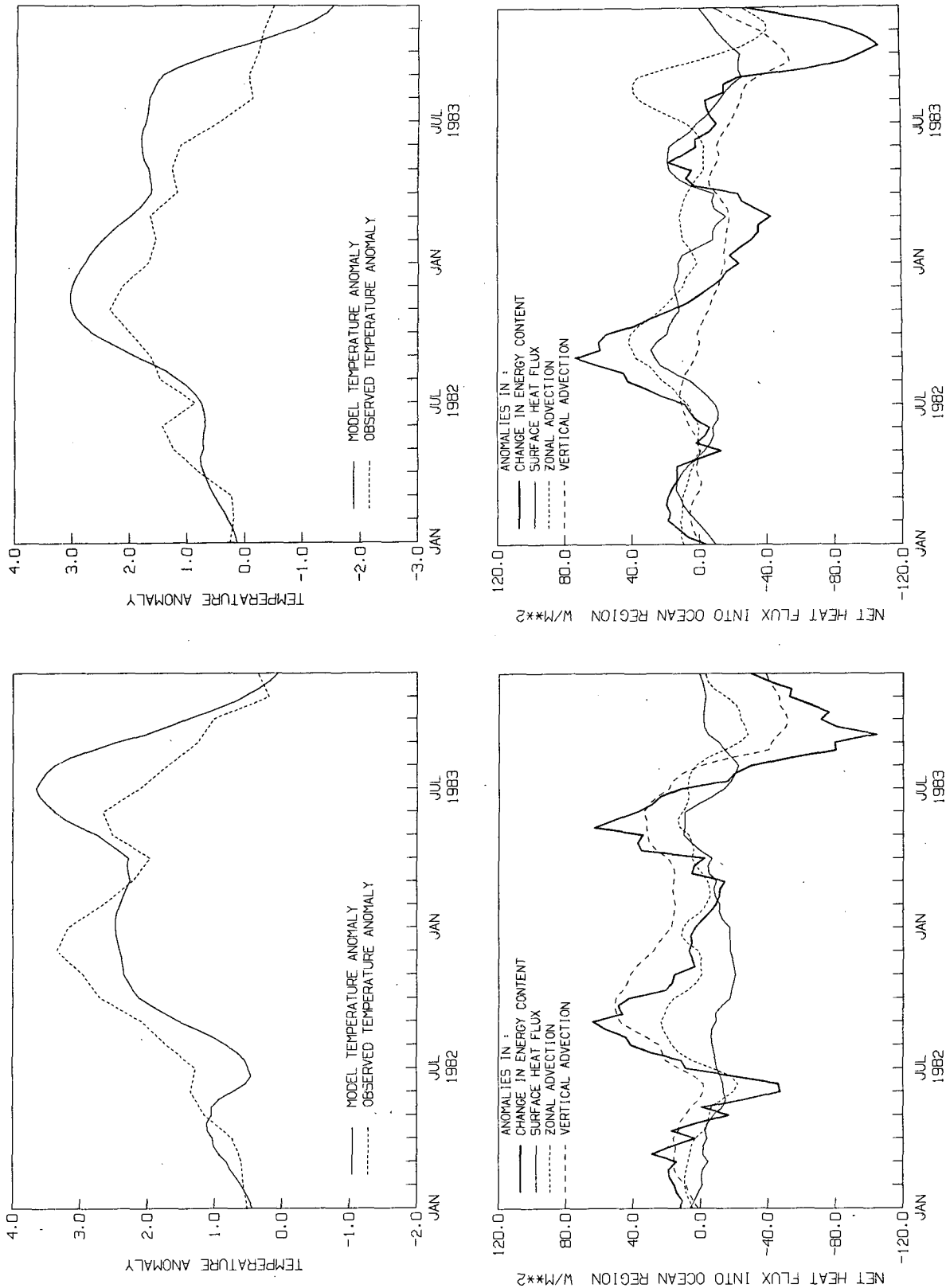


FIG. 11. As in Fig. 10 but for EQ2.

FIG. 10. (a) Model and observed temperature and (b) components of the heat balance contributing to temperature anomalies in EQ1 during the 1982/83 El Niño.

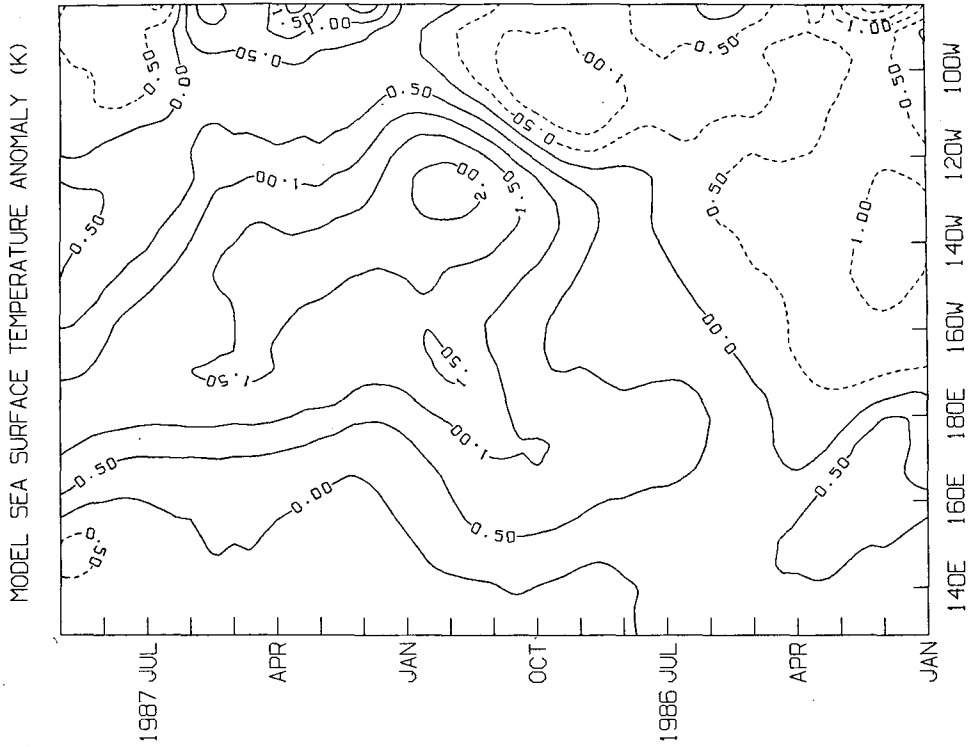


FIG. 15. As in Fig. 12 but for model surface temperature anomaly.

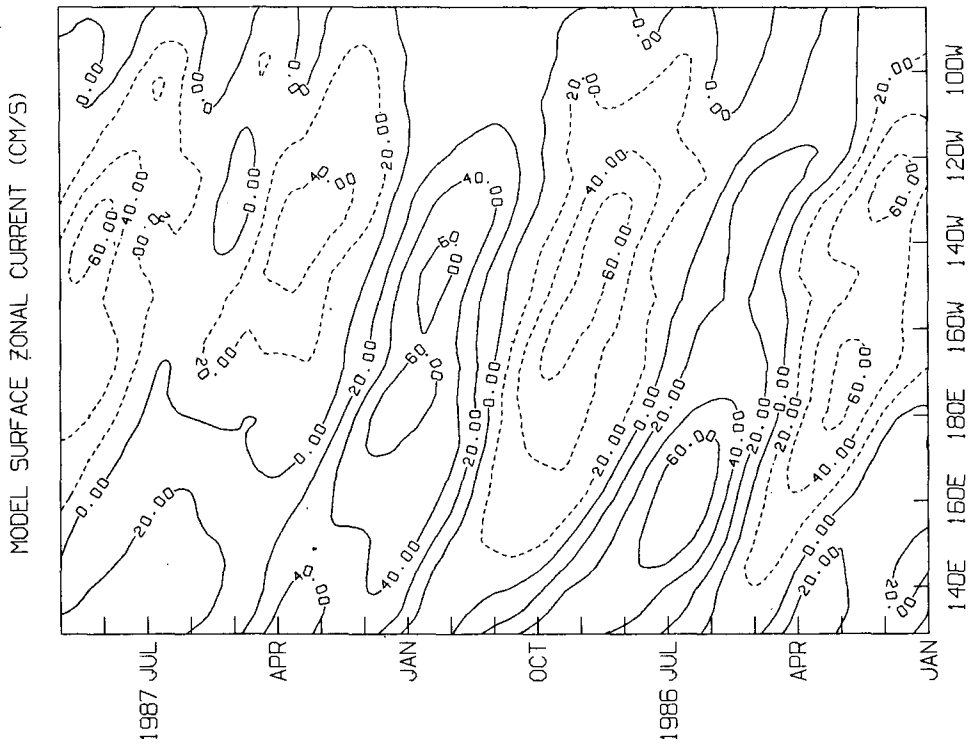


FIG. 14. As in Fig. 12 but for model surface layer zonal current.

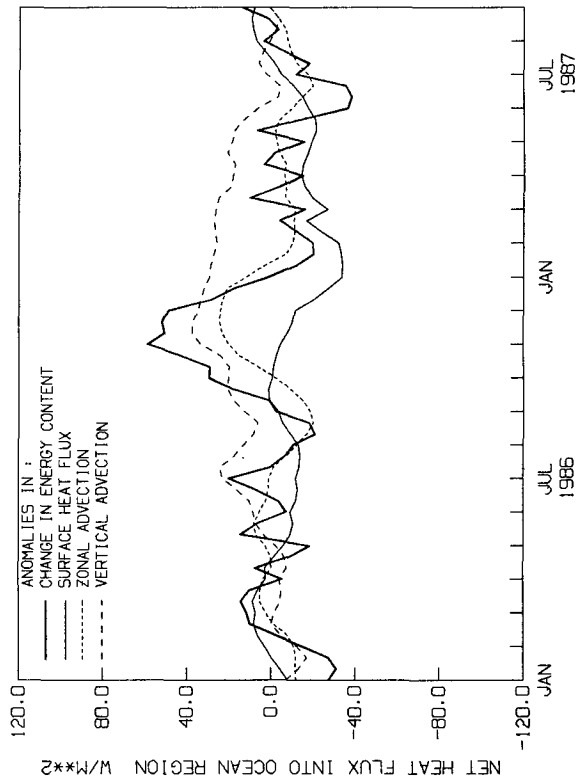
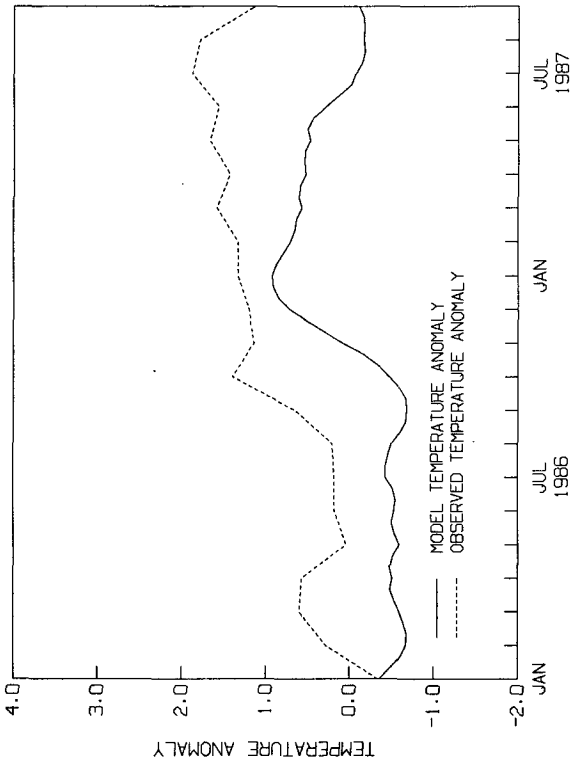
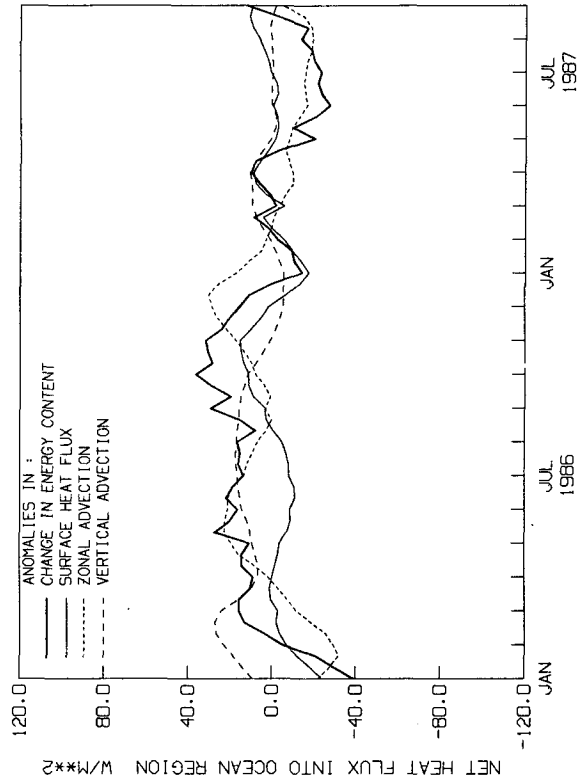
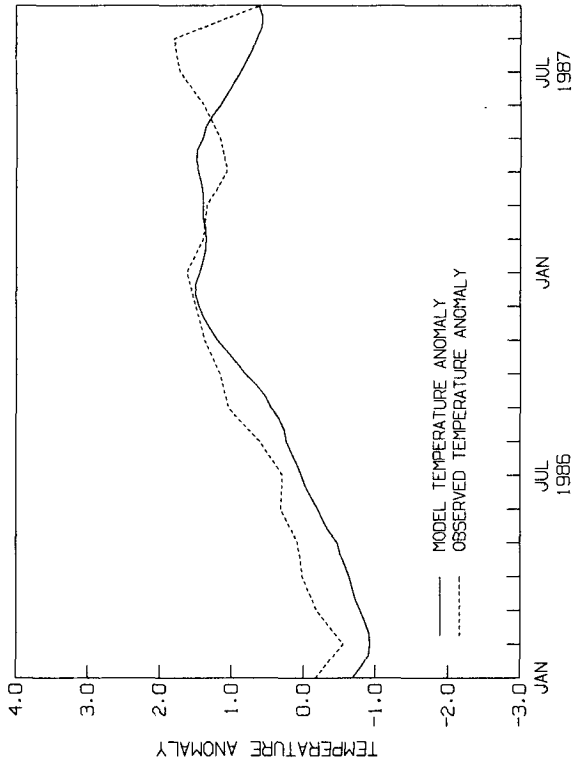


FIG. 17. As in Fig. 16 but for EQ2.

FIG. 16. (a) Model and observed temperatures and (b) components of the heat balance contributing to temperature anomalies in EQ1 during the 1986/87 El Niño.

(personal communication 1987) indicates that, although it was shallowing, the thermocline remained anomalously deep through 1987. However the model thermocline had returned to climatological depth by June. Zonal current data (McPhadden, personal communication 1987) show that during NH spring 1987 the surface current at 110°W was eastward as opposed to the weak westward current in the model (Fig. 14). It is worth noting that easterly winds occur at 110°W in the FSU analysis throughout 1987 so, if this is correct, the eastward current must be remotely forced, possibly in the reflection of a Kelvin wave at the eastern boundary. These differences of zonal currents and thermocline depth are of the sign to produce a model SSTA colder than observed. The longevity of the discrepancies suggests failings of the dynamical model are the source of error. Possibilities include neglect of nonlinear dynamics and overestimation of the efficiency of wave reflection at the western boundary. The latter would result in generation during 1987 of exaggerated amplitude upwelling Kelvin waves that raise the thermocline in the east.

4. Discussion and conclusions

The results presented in this paper provide a picture of a model that has some ability to simulate the interannual variability of the tropical Pacific. This ability is skewed towards the lower frequency variability; the major departures from climatology (El Niño events) are captured but between events, model and observations correspond poorly. Further, the evolution of each El Niño event reveals discrepancies between model and observations. While these reservations raise doubts about the model the extent of agreement suggests that the linear dynamics and nonlinear thermodynamics within this model capture much of the fundamental physics operating in the real system. Much of the smaller amplitude variability in the observations can be fairly reliably attributed to similarly small amplitude variability in the wind field. Paucity of data is such that this variability in the winds would not be expected to be accurately represented in the wind product used to force the model. For this reason alone we would expect little correspondence between the model and observed short timescale, small amplitude variability. Perhaps all that can be expected, and required is that the model will pick up the major departures from climatology. If the wind data were more reliable our standards would rise.

It should also be remembered that, because climatological annual mean cloud cover is used, no perturbations appear in the solar heat flux term. The data of Arkin et al. (1983) show large anomalies of outgoing longwave radiation, corresponding to increased cloud cover, in the central Pacific during the 1982/83 El Niño. Increased cloud cover would reduce the receipt of solar

radiation and may help explain why the model failed to cool in 1983 in line with observations. However, the problem in 1987 was that the model cooled in the east and central Pacific after the SSTA maximum at the end of 1986 but the real ocean did not. The cloud cover changes would be of the same sign as in 1982/83, suggesting there are problems other than those associated with the cloud cover and surface heat flux.

It has been demonstrated that the processes that occur during a model El Niño can be related to equatorial Pacific wind anomalies across the basin. Warmings in the east are primarily related to anomalies in upwelling advection that are a response to an anomalously deep thermocline. The thermocline is deepened by the passage of equatorial Kelvin waves excited by relaxation of the trade winds in the central and west Pacific. The resulting eastward mass transfer increases the temperature of water available for entrainment and hence capable of influencing the SST. Warming in the east, due to thermocline depression in the wake of Kelvin waves excited by westerly anomalies to the west, was common to all the El Niño events that occurred during the simulation.

Anomalous zonal advection also contributes to the production of warm anomalies but with diminishing importance from west to east. The primary cause of anomalous zonal advection is eastward currents that form the equatorial component of Rossby waves excited by westerly wind anomalies. These currents advect warm water eastward. Those same westerly anomalies suppress evaporative heat loss by weakening the trade winds. The anomalous surface heat flux thus reinforces the warming in the west and central Pacific. In the east where wind anomalies are weak or easterly the increased SST creates an anomalous surface heat flux that offsets, but cannot overwhelm, the dynamically induced warming. It is worth noting the contrast between the eastern Pacific and areas to the west. In the east anomalies related to entrainment, and aided to a lesser extent by zonal advection, force the warming and this is partially offset by the anomalous surface heat flux. To the west anomalies in surface heat flux and zonal advection force the warming, with zonal advection dominant, and entrainment acts as a negative feedback to offset this.

Seager et al. (1988) pointed out that the ability of their model to simulate climatological SST was, in part, related to the strength of the negative feedback between SST and surface heat flux which held the model in check. The role of the surface heat flux in the present simulation seems very different from that in the climatological case. It is now either a moderating influence that cannot overcome the effects of dynamically induced temperature changes, or actually adds to anomalies through its dependence on the surface wind field. The different behavior can be attributed to the slow variation of the SST and wind speed in the cli-

matological case, which allows the surface heat flux to operate as a negative feedback on the SST, even though the relaxation time for the effects of surface heat flux perturbations is, according to Seager et al., greater than 100 days. In the case of simulation of variability the winds and the SST can change rapidly. The surface heat flux can then force anomalous heating or cooling on timescales short compared to that of the relaxation time. So to view it as a simple negative feedback on the SST is overly simplistic in this case. The surface heat flux therefore plays a much more active role when the attention is focused on variability of SST, even though it is frequently overwhelmed by dynamically induced variability. This observation is tempered to the extent that variable cloud cover would add a negative feedback on the SSTA by, for example, increasing cloud cover and decreasing incoming solar radiation in regions of positive SSTA.

Major discrepancies were seen between model and observations in the two El Niño events discussed here. In 1982/83 the most striking problem is the failure of the model to cool until around August 1983 whereas the observed temperature anomaly decays much earlier. In 1986/87 the opposite occurs with the model cooling during 1987 but the observed anomaly continues to increase. In the 1982/83 case a possible explanation is the absence of cloud variability in the model, as discussed above. However errors in the wind dataset cannot be ignored. It is difficult to imagine how the equatorial Pacific could have dynamically cooled in 1983 if the trades remained weak and westerly anomalies appeared in the east.

In contrast, the problem in 1987 is more likely a model failing. Observations suggest that the thermocline in the east shallowed through 1987 but remained deeper than normal and at the same time the temperature anomaly in the east and central Pacific increased. However, the model thermocline returned to its climatological level in response to upwelling Kelvin waves and the emergence of easterly wind anomalies in the east. The gradual decline in anomalous warming due to entrainment allowed the zonal advection and surface heat flux to cool the model SSTA.

Some of the error in these two El Niño events can possibly be attributed to shortcomings of the linear approximation in the dynamical model. It is well known that in linear models the zonal equatorial currents are too westward because of neglect of zonal momentum advection by the meridional circulation (Gill 1975). This suggests that our model might underestimate the warming by zonal advection that results from westerly winds in the central and western Pacific. It also means that the cooling when the westerlies disappear might be underestimated. This is possibly what happened in the model during 1982/83. However, this alone does not explain why the model fails to cool in the east where we expect the effects of zonal advection to be

overwhelmed by entrainment. Also, because the model simulates the SSTA through the latter half of 1982 fairly well (and in fact is too warm in the central Pacific), we would have to claim that some of the warming produced in the model by the surface heat flux (central Pacific) or entrainment (eastern Pacific) in reality is produced by anomalous advection. The events of 1986/87 do however provide some support for the possible importance of nonlinear dynamics. In 1987, the westerlies decline but do not actually disappear until June. In that case the cooling in the model that occurred in the central and east Pacific by upwelling Kelvin waves and entrainment could be overwhelmed by zonal advection that is stronger than in the linear dynamical model. Other dynamical problems may include inaccurate representation of wave reflection at the boundaries.

The chances are that all of these factors—cloud cover variability, incorrect wind data, nonlinear dynamics, and mixed layer physics—are important. Much of the advance in understanding of El Niño will necessarily be achieved through intercomparison of models. Even in the presence of limitations imposed by the paucity and unreliability of the observed data useful and informative work modeling SST is waiting to be done with models ranging from simple linear dynamical models, like that presented here, to ocean GCMs. Given that the surface heat flux is not a simple negative feedback on the surface temperature anomaly, but a term of considerable importance in determining the SST and its variability useful comparisons with our results could be made if those models were run with realistic heat flux parameterizations.

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