Interannual Variability of Meridional Heat Transport in a Numerical Model of the Upper Equatorial Pacific Ocean

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ABSTRACT

The interannual heat budget of the Pacific equatorial upwelling zone is studied using a primitive equation, a reduced gravity model of the upper Pacific equatorial ocean. The model is forced with monthly mean FSU winds from 1971 to 1990. A meridional overturning cell transports heat poleward with poleward flowing warm surface water compensated by colder, deeper equatorward flow. A horizontal cell transports heat equatorward as warmer equatorward flow, which enters the equatorial zone near the western boundary to feed the Equatorial Undercurrent, is compensated by colder poleward flow in the eastern basin. The heat transported by transient eddies is equatorward.

Heat transport anomalies from the 20-year mean balance the time rate of change of heat content on interannual and seasonal timescales. Anomalies of surface heat flux do not contribute significantly to the interannual heat budget.

Although each simulated ENSO event develops differently, similarities in the interannual heat budget emerge. Heat content increases due to an anomalous net equatorward heat transport during the initial stages when the easterlies are anomalously strong. This increase is associated first with an increase in the equatorward heat transport by the horizontal cell due to the increase of the zonal temperature gradient and then with a reduction of the poleward heat transported by meridional overturning as the easterlies slacken. Following the appearance of the maximum SST anomaly, anomalous net poleward heat transport compensates for the subsequent decrease in heat content. The anomalous poleward heat transport is associated first with a reduction of the equatorward heat transport by the horizontal cell as the Equatorial Undercurrent transport decreases, and later with an increase in the meridional overturning as the easterly wind stress returns to normal and the poleward surface flow is anomalously warm.

Northward heat transport occurs during the warm phase when there is an export of heat; southward heat transport occurs during the preconditioning phase when there is an import of heat into the equatorial region. This result indicates that a cross-equatorial exchange of warm water occurs on the interannual timescale as first suggested by the sea level analysis of Wyrtki and Wenzel.

1. Introduction

With an analysis of Pacific island sea level measurements, Wyrtki (1985) demonstrated that a redistribution of warm water occurs in association with ENSO events. The sea level analysis revealed a dramatic zonal redistribution, with sea level building up gradually in the western Pacific, then discharging eastward. At the end of the event, the warm water volume in the near-equatorial Pacific is depleted, suggesting a meridional redistribution to higher latitudes. The sea level data also suggest that warm water is interhemispherically redistributed as well (Wyrtki and Wenzel 1984). They found that, in conjunction with El Niño, the sea level pressure difference between Pago Pago in the Southern Hemisphere and Hawaii reverses in such a way as to suggest that warm upper-thermocline water is transferred to the Northern Hemisphere during El Niño.

Not much is understood about these redistribution processes.

The heat content of the upper equatorial ocean (of which sea level is a proxy) may be increased either by diabatically warming the region or by dynamically deepening the thermocline. Philander and Hurlin (1988) suggest the importance of the diabatic heating in the evolution of equatorial heat content during the 1982–83 ENSO. They conclude that heat is exported during the warm phase by an increase in the net poleward transport with little net heat lost or gained through diabatic heating. During the following cold phase, the equatorial heat content is restored by a large diabatic heating through the surface. Only a small amount of heat is exported poleward by heat transport during this period. The heat budgets for the warm and cold phases average to the normal annual cycle with the ocean exporting heat poleward to balance the gain from the surface.

Their result, that the variations of upper ocean heat content during ENSO depend greatly on the diabatic
surface heat flux, is at odds with the results of simple layer models that successfully simulate the event adiabatically (Busalacchi and O'Brien 1981). Cane and Zebiak (1985), using a simple coupled linear model to simulate the coupled interaction during El Niño, emphasize the importance of the linear wave dynamics to both timing and duration of the events. It may be that the diabatic heating is important to the evolution of SST during ENSO events. However, wave dynamics influence the motion of the thermocline, which has a more dramatic effect on the variations in upper ocean heat content especially in the western Pacific. The Gent and Cane (1989) model used here is a fairly simple seven-layer primitive equation model with thermodynamics. The model has been forced with observed winds for the period 1971–1990. One of the goals of this analysis is to investigate the importance of diabatic heating on the interannual heat budget of the upper equatorial ocean.

The second goal is to understand the meridional heat redistribution process better and to determine the mechanisms important at different phases of the warm and cold events. It has been suggested by Giese and Harrison (1991) that the heat transport by instability waves may be important in both preconditioning the ocean before the onset of the event and during the warming in the central Pacific. They find that, as Kelvin waves forced by westerly wind stress anomalies propagate eastward, the meridional shear in the zonal currents near the equator is enhanced. The enhanced shear leads to an amplification of the tropical instability waves and thus an enhancement of the equatorward heat transport by the waves. They find that this contributes about half of the warming observed in the model. They also speculate that a preconditioning of the ocean with strong easterlies may increase the tropical instability wave activity, thereby increasing the equatorward heat transport, which may lead to a buildup of heat in the equatorial region and make it easier to produce a warming in the central Pacific. Since the Gent and Cane (1989) model has sufficient resolution to simulate the tropical instability waves, these hypotheses are examined in this model analysis.

The other meridional heat transport mechanisms are also investigated with the model. The heat transport by the meridional vertical circulation, which is poleward in the long-term mean, may vary with the strength of the zonal wind stress due to variations in the poleward Ekman flow. Brady and Gent (1994) found that on the seasonal timescale the heat transport by meridional overturning varied more with the variations of the vertical temperature distribution. In the boreal fall, when the zonally averaged surface temperature is colder than average and the zonally averaged temperature of the equatorward interior flow is warmer than average, the heat transported by the zonally averaged meridional overturning cell is weaker than the annual average. This is the opposite of what is expected since the zonal wind stress is stronger than average in the boreal fall leading to a stronger than average meridional circulation. The heat transport by the seasonally varying horizontal circulation or gyre exchange is equatorward and shown to be related to the strength of the Equatorial Undercurrent near the equator (Brady and Gent 1994). Horizontal zonally oriented circulation cells are formed by the undercurrent since warm water enters the EUC in the western Pacific by equatorward boundary currents, primarily originating in the Southern Hemisphere. Colder water flows poleward to compensate in the eastern Pacific. It is hypothesized that this equatorward heat transport mechanism may vary substantially on the interannual timescale. The undercurrent has been observed to disappear in the central Pacific during the 1982–83 ENSO (Firing et al. 1983). As a result the horizontal circulation may also weaken. Or, if the western boundary currents do not weaken to compensate for the disappearance of the undercurrent, the poleward compensating meridional currents may flow back across the section in the western Pacific region. This would decrease the equatorward heat transport by this mechanism because the zonal temperature difference would not be as large.

The Gent and Cane model is ideal for addressing these hypotheses. The model run analyzed here is 20 years long and thus simulates four El Niño events observed over the period 1971–1990. Many more realizations are desirable for statistical analysis but this is a start. Because it is forced with observed winds, model simulation of SST and currents can be compared to the observed SST and currents to assess how well the model simulates the observed events.

This paper is organized into the following sections. The model and the model circulation are described in section 2. A comparison to observed circulation and to observed sea level is used for model validation. The model simulation of interannual variability is discussed in section 3. The mean and interannual heat budget and mechanisms of heat transport are discussed in section 4. A summary and the conclusions follow in section 5.

2. Model description

The primitive equation reduced-gravity model developed by Gent and Cane (1989) and Gent (1991) is used to simulate the evolution of the heat content in the equatorial Pacific over the past two decades with observed wind forcing. The model run analyzed here is forced with a 20-year time series of monthly mean wind stress obtained from the Florida State University (FSU) pseudostress observations from the years 1971–1990.

Spinup of the model took place over many model years. Starting from rest the model was forced with a repeat annual cycle of wind stresses derived from the FSU database until an equilibrium was achieved. Then,
the forcing was changed to the monthly mean FSU winds from 1965 to 1988 until a new equilibrium was achieved. Using initial conditions at the beginning of 1980, a normal year, the model was forced with the wind stress time series starting at the beginning of 1968, another normal year, because the wind field near the equator during 1968 was comparable to the winds in 1980. The model was then run with the FSU forcing until the end of 1990. The last 20 years are analyzed starting with the beginning of 1971 to allow the initial transients to die off.

3. Method

Diagnosing the heat budget for a model requires estimating the terms in the evolution equation for heat content. Integrating over a volume in the Gent and Cane model by summing over the layer index \( k \), which ranges from 1 to \( n_z \), equal to 7, and integrating over a surface area \( A \) bounded by the surface \( S \), the equation for total heat content is given by

\[
\rho c_p \left( \int \int \sum_{k=1}^{n_z} \frac{\partial}{\partial t} \left[ h_k (T_k - T_{\text{deep}}) \right] dA \right) + \oint \sum_{k=1}^{n_z} h_k (T_k - T_{\text{deep}}) \mathbf{u}_k \cdot \mathbf{n} dS = \int \int (Q_s + Q_{ne}) dA + F. \tag{1}
\]

The temperature of the deep quiescent layer is \( T_{\text{deep}} \). The unit vector \( \mathbf{n} \) is normal to the lateral boundary \( S \); \( Q_{ne} \) is vertical dissipation at the bottom interface, which is proportional to the difference in temperature between the deepest moving layer and the infinitely deep quiescent layer; \( F \) is horizontal diffusion due to filtering; and \( \rho c_p \) is taken equal to 4.12 \times 10^9 J m^{-3}. Here \( Q_s \) is the surface heat flux, which is given by the following formula developed by Seager et al. (1988) with coefficients from Blumenthal and Cane (1989):

\[
Q_s = (1 - A)(1 - 0.75 C + 0.002 \theta) Q_0 - \rho_a C_E L \| \mathbf{u}_w \| (1 - \delta) q(T_1) - \alpha (T_1 - T^*), \tag{2}
\]

where \( A \) is the albedo, equal to a constant 0.06; \( C \), the cloud cover fraction, is interpolated from the monthly averages of Esbensen and Kushnir (1981); and both \( \theta \), the solar altitude, and \( Q_s \), the clear-sky flux, vary seasonally. The parameterization of the latent heat of evaporation, \( \rho_a C_E L \| \mathbf{u}_w \| (1 - \delta) q(T_1) \), depends only on the wind speed \( \| \mathbf{u}_w \| \) and the model sea surface temperature \( T_1 \) with \( \delta \) equal to 0.78. The term parameterizing the longwave and sensible heat transfer is \( \alpha (T_1 - T^*) \), with \( \alpha \) equal to 1.667 W m^{-2} and \( T^* \) equal to \(-3.0^\circ\text{C}\). The annually averaged surface heat flux typical for one of the model years was shown in Brady and Gent (1994) and is well within the range of estimates from observations using bulk formulas.

The meridional heat transport integrated across an ocean basin from the western boundary at \( X_w \) to the eastern boundary at \( X_e \) is

\[
\rho c_p \int_{X_w}^{X_e} \sum_{k=1}^{n_z} h_k v_k (T_k - T_{\text{deep}}) dx. \tag{3}
\]

As shown in Brady and Gent (1994), the time-mean meridional heat transport can be broken down into a sum of three mechanisms: the transient eddy heat transport, the heat transport by the time-mean horizontal flow or gyre exchange, and the heat transport by the time-mean meridional overturning. Direct estimates of the meridional heat transport have been made from geostrophic and Ekman transport estimates using hydrographic sections along latitude lines (Bryden and Hall 1980; Bryden et al. 1991).

4. Model validation

a. Comparison with sea level observations

The model is quite successful at simulating the interannual evolution of ocean heat content anomaly near the equator (Fig. 1). The anomaly of upper-layer volume calculated by Wyrtki (1985) from sea level measurements from 1975 to 1983 in the region from 15°N to 15°S is compared to the model heat content anomaly (from the 20-year mean) integrated over an area from 10°N to 10°S for the same time period. All of the observed low-frequency features are well simulated by the model. These include the positive anomaly observed in 1976, the gradual accumulation of upper-layer volume (heat content) from 1977 to 1982, and the relatively short discharge occurring after the peak observed in late 1982. The good agreement between the interannual anomalies of the observed upper-layer volume and model heat content and the work of Brady and Gent (1994) give confidence that the important processes in the interannual heat budget can be realistically diagnosed using the Gent and Cane model.

b. The undercurrent during the 1982–83 ENSO

A good validation of the Gent and Cane model is the successful simulation of the response of the equatorial ocean zonal velocity field to both local and remote changes in the wind stress field as compared to the observations. A time series of zonal velocity on the equator at 159°W as a function of depth is shown in Fig. 2a to compare with the measurements of Firing et al. (1983) as shown in Fig. 2b. Over the last half of 1982, the zonal velocity core of the undercurrent weakens and eventually disappears. At the end of 1982, a strong eastward jet appears at the surface for 3 months. The EUC is not at normal strength again until near the end of 1983. The time series of equatorial zonal velocity of the model (Fig. 2a) compares quite well to the observations (Fig. 2b).
These simulated and observed variations in the zonal velocity field on the equator are directly attributable to changes in the zonal wind stress forcing. When the easterly winds started to relax and turn westerly in mid-1982 in the western Pacific, an eastward surface jet was forced. This caused an eastward surge of warm water that reduced the slope of the thermocline. The relaxation of the thermocline caused a reduction of the eastward pressure gradient force that weakened the zonal flow at the core of the undercurrent and it subsequently disappeared. The disappearance of the undercurrent was observed to occur later at progressively eastward locations, (Firing et al. 1983; Halpern 1987). The eastward surface jet observed at 159°W at the end of 1982 in both panels is forced by the local westerly wind anomaly that was observed to penetrate progressively eastward into the central Pacific after originating in the western region.

c. Model simulation of ENSO

The panels in Fig. 3 show longitude versus time plots of (a) heat content anomaly averaged between 2.35°N and 2.35°S, (b) between 5° and 10°N, (c) model SST anomaly between 2.35°N and 2.35°S, (d) zonal wind stress anomaly between 2.35°N and 2.35°S, (e) NMC SST anomaly between 2.35°N and 2.35°S, and (f) latent heat anomaly between 2.35°N and 2.35°S. The model anomalies were calculated by subtracting the 20-year mean of the time series of monthly averages over 1971 to 1990 and then smoothing with a 13-month running filter centered on the month and tapered at the ends. The National Meteorological Center SST anomaly was calculated by subtracting the time-averaged SST computed from 1970 to 1991 and then smoothing with the 13-month running filter tapered at the ends.

1) Heat Content

Four significant positive events are observed in the time history of equatorial heat content anomaly along the equator (Fig. 3a). All the positive equatorial heat content anomalies simulated here evolve in a similar way by a buildup of positive heat content anomaly in the western Pacific but with varying duration over varying degrees of longitude. With a relaxation in the zonal winds, eastward propagation across the basin is observed with the anomaly arriving in the far eastern Pacific about one year later. Negative equatorial anomalies in Fig. 3a propagate eastward following the positive anomalies. The only event with an anomalous character in this regard is the 1982–83 event. While eastward propagation into the eastern basin is clear in the 1982–83 simulated event, it is preceded by a weak positive heat content anomaly that propagates westward from the central Pacific into the western Pacific.

Westward propagation of about 0.3 m s⁻¹ is suggested in the plot of off-equatorial heat content anomaly, which is integrated latitudinally from 5°N to 10°N. The behavior observed in Fig. 3b can be compared to the behavior of a Rossby wave basin mode as discussed by Cane and Moore (1981). The off-equatorial anomalies of heat content appear to behave like standing waves with a node at midbasin.

It is possible to interpret the evolution of certain events simulated here within the context of remotely
forced equatorial waves according to the theory of Schopf and Suarez (1988). Consider the event of 1972 for example, (Fig. 3a). It begins as a positive anomaly in equatorial heat content in 1971, which builds up from the stronger than average easterlies observed in 1971. After the onset of the wind relaxation event, the positive anomaly propagates eastward. In phase with the relaxation of the wind field, a negative anomaly of off-equatorial heat content develops to the west of the wind relaxation event (Fig. 3b). After arriving at the west coast, the negative anomaly reflects as an equatorial Kelvin wave and propagates eastward eventually arriving in the eastern basin in 1973 to end the event (Fig. 3a). Meanwhile, a positive anomaly in the off-equatorial heat content on the eastern side of the basin, a product of the reflected Kelvin wave, propagates back to the west to arrive at the end of 1973 on the western side of the basin. This off-equatorial anomaly appears to be related to the positive equatorial heat content anomaly at the start of the next event in the western Pacific. This shows how the earlier event cycles into the next. Although the behavior is not as clearly indicated in zonal wind stress anomalies shown in Fig. 3d as in the simpler model of Schopf and Suarez (1988), similarities are striking. However, such a simple application of the delayed oscillator theory does not explain the other events well. The later events are far more complicated although there appears to be some cyclic behavior.

2) Sea surface temperature

Only three events of very large positive or warm model SST anomalies are observed in Fig. 3c. In addition, there is a westward propagating weak warm SST anomaly associated with the event of 1976–77. This event is rated as moderate by Quinn et al. (1987) but appears to be very weak in both the model and NMC observed SST anomalies (Fig. 3c). Typical amplitudes of the very large warm anomalies are in the range of 2–3.5°C. Note the very large amplitude of the 1982–83 warm anomaly, which is classified as a very strong ENSO event (Quinn et al. 1987). Cold or negative SST anomalies follow all warm events. The cold
anomalies of the last decade appear to cover the entire Pacific domain, whereas the warm anomalies are more concentrated in the central to eastern Pacific.

Note that the equatorial heat content anomalies and the equatorial SST anomalies appear very different in both evolution and spatial correspondence, but it is clear that they are part of the same event. The SST anomaly of 1982–83 appears very similar to the equatorial heat content anomaly in the central to eastern Pacific. In general, the heat content anomaly in the central to eastern Pacific is related more to the anomaly of SST than in anomalies of thermocline depth. In the western Pacific, the heat content anomaly is related more to the anomalies of thermocline depth.

The SST anomalies produced by the model differ from the observations, Fig. 3e in quantitative detail only. Most of the qualitative features of the observed equatorial SST field are reproduced in the model SST field. However, the near-equatorial SST anomalies evolve quite differently from the near-equatorial heat content anomalies. The heat content anomalies show primarily eastward propagation, while the SST anomalies show either westward or eastward propagation. The strongest model SST anomalies either develop or strengthen near the date line and extend longitudinally into the eastern basin, although they may develop there first, as in the case of the events of 1972 and 1986–87. There is no consistency to the absolute location of the maximum anomaly. The 1982 event has a maximum located near 135°W, whereas the other two strong events have maxima near the date line. Note the difference between the evolution of SST and the evolution of equatorial heat content for the 1976 event. The weak warm SST anomaly associated with the 1976 event propagates to the west, whereas the anomaly in equatorial heat content (Fig. 3a) clearly propagates east-
ward. There is some indication of weak (0.5°C) anomalies in the western Pacific associated with the anomalies of positive equatorial heat content. In the eastern and central Pacific, the strong warm SST anomalies clearly develop after the arrival of the heat content anomaly.

3) ZONAL WIND STRESS

The strong zonal wind stress anomalies, Fig. 3d, are confined to midbasin. Positive zonal wind stress anomalies denote westerly anomalies or weakening of the normal easterly winds, and negative zonal wind stress anomalies denote easterly anomalies or strengthening of the easterly winds. Positive equatorial heat content anomalies develop in the western Pacific when there are negative (easterly) wind stress anomalies. Note the long and unusually strong buildup of equatorial heat content that occurred from mid-1973 to mid-1976 associated with the long easterly anomaly near the date line. However, there is no particular correspondence between strong easterly anomalies and strength of the SST anomalies. Although the 1976 event was preceded by a long duration of stronger than average easterlies, the associated SST anomalies are observed to be weak. In contrast to other events, the anomaly in heat content is larger in the west than in the central to eastern Pacific, probably because the SST anomalies, which contribute to the heat content anomaly more significantly in that region, are rather weak. In the strong 1982–83 SST event, the preceding years showed a fairly weak pre-conditioning of easterly anomalies and a weak buildup of equatorial heat content in the west. There is better correspondence between strong westerly (positive) wind stress events and strong warm SST anomalies, as the event of 1982–83 dem-
onstrates. For all events, the SST anomalies and the zonal wind stress anomalies correspond well temporally. For the strongest three events, however, the maximum SST anomaly is located to the east of the maximum zonal wind stress anomaly.

4) **TIME SERIES OF SST, ZONAL WIND STRESS, AND ZONAL TRANSPORT**

Plots of the spatially averaged zonal wind stress anomaly, the spatially averaged SST anomaly and the zonal transport at 150°W, meridionally integrated from 2.35°N to 2.35°S, Fig. 4, all demonstrate large variability on the interannual timescale. All time series of monthly means have the 20-year mean removed and are smoothed to remove the annual cycle with a 13-month weighted filter.

The 20-year mean zonal transport in the undercurrent from 2.35°N to 2.35°S, found to be 31.7 Sv (1 Sv = 10^6 m^3 s^-1), is larger than the 26.0 Sv between ±2° latitude integrated to 500-m depth found at 150°W in the diagnostic model of Bryden and Brady (1985). However, averaged over the same time period as the diagnostic model, 1979–81, the mean zonal transport at 150°W in the model compares very favorably at 26.4 Sv.

Figure 4c shows three significant negative (westward) zonal transport anomalies on the equator. The minima occur at the end of 1972 and in the middle of 1983 and 1987. The negative anomalies are associated with positive anomalies of SST (Fig. 4b) and positive anomalies of wind stress (Fig. 4a). Hence, the negative anomalies of zonal transport are associated with the warm phase of ENSO. With a
mean of 31.7 Sv, at the minimum of each negative transport anomaly of nearly 25 Sv, the undercurrent has essentially disappeared. It is interesting to note that following all three events, a large (20–30 Sv) positive (eastward) transport anomaly occurs, which is associated with both cold SST anomalies and negative wind stress anomalies (or the reestablishment of the easterlies during the cold event). From the warm event to the cold event, this is a dramatic change in the near-equatorial zonal transport of nearly 50 Sv over the time period of a year.

The model results do not concur with the observations, suggesting that an anomalous eastward transport of the equatorial currents accompanies the
appearance of warm surface water. The anomaly of eastward transport from $10^\circ$N to $10^\circ$S (Fig. 4d) shows only a modest 10–20 Sv increase in the eastward transport at the onset of the warm events, whereas Wyrtki (1985) and Kessler and Taft (1987) suggest an increase of at least 40 Sv. However, an eastward increase in the warm surface flow, coincident with the decrease in the colder undercurrent velocity observed in Fig. 2, is coincident with a decrease in the eastward transport of the undercurrent. This redistribution of volume flux to warmer surface temperatures can account for the observed increase in surface warming.

5. Results

a. Mean annual heat budget

The equatorial oceans gain heat through the surface all year round. Thus, to maintain a long-term equilibrium, the equatorial oceans must transport heat poleward. The poleward export of heat is accomplished by a meridional overturning of the time mean flow. Through Ekman dynamics, the mean easterlies drive the warm surface water poleward. The warm surface water is replaced through upwelling by colder water from below. The zonally averaged overturning cell actually exports more heat than is required and is partially compensated by equatorward transport of heat by both the time mean horizontal cell or “gyre exchange” and the transient eddies. The horizontal cell transports heat equatorward as warmer equatorward flow, entering the equatorial region near the western boundary to feed the undercurrent, is compensated by colder poleward flow in the eastern basin. 

Brady and Gent (1994) have shown that seasonal variations in the heat transport by gyre exchange are related to seasonal variations in the strength of the undercurrent. When the undercurrent strengthens, the horizontal gyrelike circulation is strengthened and hence the equatorward heat transport by gyre exchange increases. However, it is shown that seasonal variations in the zonal temperature gradient do not affect seasonal variations in the transport by gyre exchange. The transient eddies are generated by an instability of the zonal currents (Philander et al. 1985) and are observed to transport heat down the mean meridional temperature gradient toward the colder equatorial water (Hansen and Paul 1984; Bryden and Brady 1989; Brady 1990).

The following 20-year mean heat budget pertains to the near-equatorial upwelling region, defined as the domain from 2.35$^\circ$N to 2.35$^\circ$S from the west to east coast of the model. There is a negligible long-term trend in heat content over the 20-year run. To maintain this equilibrium, the circulation within the upwelling zone exports heat at an average rate of 0.30 PW (1 PW is $10^{15}$ W), which nearly balances a net average surface heat gain of 0.36 PW. The residual deficit of 0.06 PW is due to heat lost by the diffusive processes $F$ and $Q_{av}$ in Eq. (1). Of this residual, about 0.04 PW is due to interfacial diffusion between the lowest layer and the deep quiescent layer $\int Q_{av} dA$, while the remainder is due to the effect of horizontal filtering $F$. The 20-year mean heat budget discussed here compares well with the 3-year mean heat budget discussed in Brady and Gent (1994), from a model run forced with the repeat monthly mean wind stress over the annual cycle derived from the same wind stress field. They found that for the same upwelling region, an ocean heat export of 0.35 PW nearly balances the surface heat flux of 0.37 PW.

The net heat export is broken down into the sum of three heat transport mechanisms, as in Brady and Gent (1994), by defining the time-mean to be over one year. The heat transport mechanisms are computed each year, and then the ensemble is averaged. It is found that the region loses heat poleward by meridional overturning at an average rate of 1.99 PW. This compares to 1.96 PW found by Brady and Gent (1994). This loss is partially offset by a net equatorward heat transport due to gyre exchange, at a rate of 1.29 PW, and transient eddy heat transport, at a rate of 0.40 PW. These estimates agree fairly well with the estimates found by Brady and Gent (1994), which were 1.37 and 0.24 PW, respectively. The larger net heat transport by transient eddies found here is due to the richer frequency content of the variability of the observed forcing used here, which generates stronger variability of the equatorial currents.

b. Interannual heat budget

The terms in the interannual heat budget of the equatorial upwelling region are shown in the three panels of Fig. 5. It shows that the net heat transport anomaly nearly balances the time rate of change of the heat content. Over the entire time series, the anomalies in the net surface heat flux $Q_s$ are negligible on the interannual timescale in the model heat budget. The surface heat flux anomaly is primarily due to the anomaly in latent heat loss shown in Fig. 3f. The maximum negative anomalies in the surface latent heat flux are no more than 15–25 W m$^{-2}$. This result contradicts the results from the heat budget of the GFDL model-simulated 1982–83 ENSO as discussed by Philander and Hurlin (1988). These two studies concur that during warm events heat is lost through the ocean surface primarily by anomalous latent heat loss due to anomalously warm SST as is shown in Fig. 3f. However, this model suggests that the latent heat loss is insignificant in the interannual budget compared to heat exported by the ocean circulation.

Four events of significant heat export are simulated in the 20 years of model run. Only three of the events show significant heat export greater than 0.5 PW pole-
ward. The simulated 1976 ENSO is much weaker. A particularly large poleward export of heat of nearly 1 PW occurs during the 1982–83 event. The maximum export of heat occurs at the time of maximum SST anomaly as shown by comparison with Fig. 6b. This is to be contrasted with the time series of time rate of change of heat content (Fig. 5a), which indicates that the large exports of heat are associated with the large decreases in heat content. This shows that changes in the model heat content are not associated with changes in SST primarily. The exception is the eastern Pacific, where changes in heat content are caused primarily by changes in SST, as is shown in Figs. 5a and 5c. Here anomalies in the heat export are related to anomalies in SST.

Before each large heat export event, there is a period where heat is imported into the equatorial region. These periods occur over a longer period of time as shown for the 1976, 1982–83, and 1986 events. The time series starts too close to the beginning of the 1972 event to show this. This import first occurs in the western basin and subsequently in the eastern basin as discussed in section 5d.

c. Mechanisms of interannual meridional heat transport

Shown in the upper panel of Fig. 6 is a breakdown by mechanism of the net heat transport in PW for the upwelling zone as defined above. The poleward heat lost by meridional overturning (MO) is shown by the dashed line, the equatorward heat gained by the horizontal cell or gyre exchange (GE) is shown by the dotted line, and the equatorward heat gained by transient eddies (TE) is given by the dot-dashed line. The sum, or total net heat transport into the upwelling zone, is shown by the solid line. Each of the terms is estimated with a mean defined over a year. Values are plotted every three months with no subsequent smoothing. In the lower panel of Fig. 6, the 20-year mean is subtracted to show the interannual anomalies.
A poleward anomaly of GE, which is produced by a decrease in the normally equatorward GE, is responsible for the initial heat loss during the warm event. A poleward anomaly of GE can be related to either a decrease in the strength of the undercurrent or a decrease in the zonal temperature gradient at depth. Comparing the anomalies of GE to zonal wind stress anomalies, it appears that the two are almost 180 degrees out of phase on the interannual timescale. The reason is that changes in the basin-scale zonal wind stress produce changes in the zonal slope of the thermocline, which are nearly in phase on the interannual timescale. The slope of the thermocline is proportional to the east–west zonal temperature difference. Because the eastward pressure gradient force will change as a result of changes in the zonal wind stress, the strength of the undercurrent may also change. Both the undercurrent and the zonal slope of the thermocline will decrease as a result of a decrease in the strength of the easterly winds. They will decrease the equatorward heat transport by GE resulting in a poleward anomaly. This happens during the beginning of the event when a large heat export is observed in the model.

It has been observed that, between 110° and 140°W at the annual timescale, the local zonal wind stress and zonal pressure gradient are balanced to within the errors of the analysis (McPhaden and Taft 1988). They concluded that, at the annual timescale, the zonal pressure gradient balanced in a quasi-steady way the local wind stress forcing. This would be expected to hold at longer timescales also. McPhaden and Taft (1988) observed a doubling of the easterly zonal wind stress from 1984 to 1986, and likewise a doubling of the zonal pressure gradient, with only a negligible increase in the eastward transport of the undercurrent. This was the period before the onset of the 1986 ENSO, when the model indicates a buildup of heat content in the equatorial Pacific.

During this early buildup period the winds are anomalously strong, which will tend to increase the slope of the thermocline. This will tend to increase the equatorward GE contribution to the initial equatorward heat transport anomaly by increasing the east–west temperature difference. It is interesting to note before the 1976 event the equatorward anomaly of heat transport by GE was both comparatively large and of long duration, whereas the event was comparatively weak. The buildup of heat content in the western Pacific was also comparatively large for this event, as shown in Fig. 3a. In the model, contrary to observations, there is a large increase in the undercurrent transport during the cold event simultaneously with the increase in the heat transport by GE.

An equatorward anomaly of heat transport by meridional overturning (MO) usually offsets the poleward anomaly of GE when the winds begin to slacken at the onset of the warm event, as can be seen in the 1982–83 event. The maximum equatorward MO anomaly occurs before the maximum westerly wind stress anomaly. This can occur for two reasons. The first is that the zonally averaged temperature of the surface water becomes relatively cool due to increased upwelling resulting from preconditioning by the previ-
ously stronger easterlies. The second is that the zonally averaged transport-weighted temperature of the interior flow becomes warmer because the bulk of the equatorward flow is shifted to shallower, warmer water due to the increase of the zonal pressure gradient near the surface. Both can reduce the overturning temperature difference, which results in an equatorward anomaly of heat transport by MO.

The heat transported by MO can be written as follows:

$$MO = \langle \nu_{surf} \rangle \left( \langle T_{surf} \rangle - \langle T_{int} \rangle \right)$$

$$+ \left( \sum_{k=1}^{nz} \langle \nu_{knet} \rangle \right) \left( \langle T_{int} \rangle - T_{deep} \right),$$

(4)

where $\langle \nu_{surf} \rangle$ is the net volume transport in the first model layer, which is equal to the net upwelling transport into the surface layer, and $\langle T_{surf} \rangle$ and $\langle T_{int} \rangle$ are the transport-weighted average temperature in the surface ($k=1$) and interior (summed over $k=2$ to $nz$), respectively. The term $\sum_{k=1}^{nz} \langle \nu_{knet} \rangle$ is the net volume transport in the equatorial zone. The transports are zonally integrated across the width of the basin as denoted by the symbols $\langle \rangle$ and averaged in time to remove the annual cycle. The temperature is zonally averaged and averaged in time over the same interval. The above expression can be linearized about a 20-year mean state. The anomaly of MO with respect to the 20-year mean is written approximately as follows:

$$MO' \approx \langle \nu_{surf} \rangle' \left( \langle T_{surf} \rangle - T_{int} \right)$$

$$+ \langle \nu_{surf} \rangle \left( \langle T_{surf} \rangle - T_{int} \right)'$$

$$+ \left( \sum_{k=1}^{nz} \langle \nu_{knet} \rangle \right)' \left( \langle T_{int} \rangle - T_{deep} \right)$$

$$+ \left( \sum_{k=1}^{nz} \langle \nu_{knet} \rangle \right) \left( \langle T_{int} \rangle - T_{deep} \right)'$$

(5)

where the overbar is the 20-year mean and the prime is the deviation from the mean of the smoothed time series. Equation (5) is not exact due to nonlinearities, but the inequality is very small over the entire time series. Figure 7a shows the contour field of smoothed volume transport as a function of zonally averaged temperature. The dashed line indicates the transport-weighted temperature of the equatorward interior flow; the anomaly of the interior temperature is shown by the dashed line in Fig. 7b. Note that the transport-weighted temperature decreases because of an increase in transport in the deeper levels. It increases as the bulk of the interior transport moves to warmer, shallower levels. Thus, the vertical distribution of the bulk of the equatorward interior flow varies on the interannual timescale. The anomalies in the sum of the volume transport in layers 2–7 compensate for the anomalies in the transport in the surface layer as shown in Fig. 7c. The net transport is also shown and indicates a net loss of volume during ENSO events particularly during the 1982–1983 event.

Figure 8 shows the contribution of the individual terms in (5) to MO'. In Fig. 8a, the dashed line indicates the contribution due to variations in upwelling. These anomalies are related to anomalies in the zonal wind stress. The time-averaged surface layer temperature is equal to 27.1°C, and the time-averaged interior temperature is equal to 21.4°C. Figure 8b shows the contribution due to variations in the overturning temperature difference, the second term in Eq. (5). The average net surface transport is 85.9 Sv poleward. These two terms account for most of the heat transported by MO since the last two terms in (5) are small (Figs. 8c and 8d), although the contribution of the third term can be significant at times. The last term is negligible small because the mean net volume transport is 0.2 Sv for the 20-year time series. Comparing Figs. 8a and 8b suggests that most of the poleward anomalies in MO, except the poleward anomaly observed from 1988 to 1989, are due to positive anomalies in the overturning temperature difference. Most of the equatorward anomalies in MO are due to a decrease of poleward transport in the surface layer. As the surface water in the equatorial zone warms or the interior flow becomes colder, the poleward MO can become stronger even though the winds are relatively weak. During the 1982–83 event, when the winds are still slack, there is a poleward increase in MO due to the warm SST anomaly observed at that time in the Central Pacific.

The anomalies of heat transport by the transient eddies (TE) are generally negligible over the 20 years simulated except for the unusually large equatorward anomaly during 1982 and 1983. In this event, the equatorward anomaly of heat transport by TE is as large as the heat transport anomalies by the other mechanisms. Because the mean was defined over a year for this calculation, the timescale of the transient motions that contribute to this heat transport mechanism can be larger than a month, which is the timescale of the instability waves. Thus, the large anomaly in the transient eddy heat transport observed during this event may be due to transient motions other than the instability waves. This model does not agree with the hypothesis put forth by Giese and Harrison (1991) that, prior to a warming in the central Pacific, the amplitude of the instability waves increases due to preconditioning by the stronger than normal easterlies. This model does not show a significant increase in the transient eddy heat transport during the periods of stronger than normal winds, which precede the warm events. Giese and Harrison (1991) find that the warm anomalies are produced in their model by zonal advection of heat by remotely forced Kelvin waves and also by meridional advection of heat due to the enhanced instability waves.
They suggest that, after the passage of the Kelvin wave, the local meridional shear between the South Equatorial Current and undercurrent will increase leading to the amplification of the instability waves and an enhanced equatorward heat transport. This might have happened in 1982 but cannot be addressed in this analysis because one year was chosen as the period over which to define the time mean.

**d. Interhemispheric and zonal redistribution of heat**

The panels of Fig. 9 show the meridional heat transport anomalies in PW at ±10°, ±5°, and ±2.35° latitude. The 20-year mean and annual cycle have been removed. Note that the meridional heat transport is northward when there is a net heat export from the box and southward when there is a net heat import. This tendency for cross-equatorial exchange is consistent in all four simulated ENSO events in the 20-year time series, which agrees with the sea level study of Wyrski and Wenzel (1984). They showed that a cross-hemispheric redistribution of warm water takes place on a timescale of approximately 4 years. They found that, in conjunction with El Niño events, the sea level difference between Hawaii (21°N, 158°W) and Pago Pago (14°S, 171°W) reverses with an amplitude of about 6 cm. This results in a migration of about 4 Sv of warm water northward across the equator during warm events. They hypothesized that ENSO events trigger an interhemispheric mode of oscillation of the subtropical gyres forced by the interannual changes in the wind torque.
over the subtropical gyres. This model has no well-formed subtropical gyres, because the north and south boundaries are at 30° latitude, but still shows a cross-equatorial redistribution occurring on the interannual timescale with the same tendency of northward cross-equatorial transport during the warm phase. Since the wind stress forcing is derived from observations, this redistribution is probably due to changes in the wind stress curl.

The meridional volume transport across the equator, associated with low-frequency quasi-steady motions of the interior circulation, can be estimated by the zonal
The integral of the equatorial wind stress curl divided by $\beta$ from the Sverdrup relation (Sverdrup 1947). The smoothed anomaly of Sverdrup transport, taken with respect to the 20-year mean, equal to 0.87 ($\pm$0.25) Sv, is shown in Fig. 10b. For comparison, the model cross-equatorial volume transport anomaly from the 20-year mean of 1.37 ($\pm$1.00) Sv is shown in Fig. 10a. Note that although there are first-order differences, the two time series are reasonably correlated, with a correlation coefficient of 0.39, which is just significant at the 95% confidence level. The confidence level is estimated using 28 degrees of freedom resulting from an integral timescale of 8.1 months as estimated from the integral of the product of the autocorrelation functions as discussed by Sciremammano (1979).

The big events may be due to the response of the interior transport to changes in the zonally integrated wind stress curl. The differences noted between the two time series may be due to waves or changes in the strength of the western boundary current. The volume transport of the western boundary current is included in Fig. 10a, whereas the Sverdrup theory is only valid for the interior quasi-steady circulation. This requires further investigation.

Joyce (1988) discusses wind-driven cross-equatorial flow and calculates a mean net cross-equatorial heat transport in the range of 0.5–1.1 PW using the wind stress curl field from Hellerman and Rosenstein (1983). In the model, the 20-year mean cross-equatorial meridional heat transport is 0.28 PW northward, which is smaller than the Joyce estimate but close to the indirect estimate of Hsiung (1985) of about 0.32 PW northward. Talley (1984) calculates a net southward meridional heat transport of 0.53 PW across the
equator in the Pacific but cautions that large errors in the bulk formulas for the surface heat fluxes can question the sign of the meridional heat transport at the equator. This is also true of the Hsiung (1985) estimate. Moreover, ignoring the heat transport through the Indonesian Archipelago as Talley (1984) and Hsiung (1985) have done, would affect the estimate of cross-equatorial heat transport by these indirect methods. Talley estimates it to be as large as 0.6 PW.

Figure 5a shows eastward propagation of heat content along the equator. The panels of Fig. 11 show the zonal redistribution of heat in PW. Each latitude band is partitioned into an eastern and western box at 150°W. The largest export of heat occurs in the equatorial bands and the bands from 5° to 10° latitude. Overall, the east and west heat transport anomalies are nearly 180 degrees out of phase in the higher latitude bands indicating that a zonal redistribution occurs there. In the equatorial band, the time series are shifted by approximately one year or half of ENSO's duration. In the near equatorial band, the largest import into the eastern box occurs before the largest export out of the western box. In addition, note that heat is exported first in the poleward boxes. The box from 5° to 10°N exports heat at the same time that heat is imported into the equatorial box. This is particularly noticeable in the 1982–83 event.

When there is an import in the eastern box and an export in the western box such as occurs in 1982, there is not a noticeable increase in the undercurrent transport. In the equatorial bands the eastern box exports more heat than is imported into the western box. This occurs at the time of northward heat transport across 2.35°N, suggesting that heat is removed from the eastern box and transported northward. There is a weakening of the undercurrent at this time. The large increase in the eastward transport of the model undercurrent at 150°W occurs after the period of large heat export anomaly and does not appear to be associated with a change in the east–west redistribution of heat.

6. Conclusions

The Gent and Cane seven-layer primitive equation model is quite successful at simulating warm and cold ENSO events using observed wind stress. The evolution of model heat content on the interannual timescale compares favorably to the estimate of upper-layer volume anomaly using sea level measurements (Wyrtki 1985). The response of the equatorial zonal velocity field in the model compares favorably to observations in the central Pacific (Firing et al. 1983) of the disappearance of the undercurrent during the 1982–83 ENSO. The annual mean model heat budget for a region near the central equatorial Pacific compares well with the heat budget from a diagnostic model (Bryden and Brady 1985). Thus, it is concluded that using the model to investigate the processes important in the evolution of interannual heat content anomalies will extend to the observed ENSO phenomenon.

The results of this investigation show that, on the interannual timescale, the anomalies in model heat content of the upper thermocline are balanced by anomalies in the net heat transport near the equator. Anomalies on the interannual timescale in the net surface heat flux are much smaller than anomalies in the net heat transport. This result disagrees with the heat budget study of the 1982–83 ENSO event using the GFDL model (Philander and Hurlin 1988). However, Springer et al. (1990) show a good correspondence between their model, an adiabatic linear 1½-layer model forced with observed winds, and sea level observations of the 1982–83 El Niño. This suggests that the surface heat flux forcing is unimportant in the development of heat content anomalies. An analysis of local heat storage in the mixed layer using XBT data (Meyers et
al. 1986) indicates that cooling due to evaporation is the dominant process in the cooling of SST in the western Pacific during ENSO. The model study presented here shows that the anomalies of latent heat of evaporation, which account for nearly all of the anomalies in the total surface heat flux, do not affect the anomalies of heat content in the upper thermocline. However, the heat budget of the mixed layer was not examined here. It was noted earlier that anomalies of SST and heat content are very different especially in the western Pacific where heat content anomalies depend more on variations in the depth of the thermocline than variations in SST.

A model with linear dynamics and nonlinear thermodynamics was used by Seager (1989) to examine the development of SST anomalies near the equator from 1970 to 1987. He showed that in the western Pacific, SST anomalies are forced by zonal advection while the surface heat flux anomalies represent a small negative feedback effect. In the eastern Pacific, his model showed that SST anomalies are forced mostly by entrainment anomalies and to a lesser extent by anomalies in zonal advection. Thus, in both regions the forcing by the surface heat flux anomalies was smaller than the dynamical forcing, as is the case in this model heat budget analysis.

The magnitude of this model's surface heat flux anomalies compares well with that found by Weare (1983). His analysis of net surface heat fluxes from observations using bulk formulas over the period from 1957 to 1976 showed that the interannual anomalies of net surface heating were of the order of 25 W m\(^{-2}\).
He suggested that anomalies of net surface heating may contribute to the warming observed during the buildup stage. However, in the mature phase, the anomalies were noisy and small and would probably have little effect on the evolution of SST.

One aspect of the net surface heat flux that could be improved is the parameterization for cloud cover amount $C$ in (2). For this model run, the cloud cover was taken to be the annually repeating monthly mean observed cloud cover of Esbensen and Kushnir (1981). During ENSO the observed clouds are quite different due to anomalous movements of the major centers of convection. The intertropical convergence zone, for example, moves south of its normal position and the convective regime normally located in the Western Pacific moves eastward. The associated cloud cover anomalies alter the incoming solar flux significantly. To test whether this would have a sizable effect on either SST or the model heat budget, it would be interesting to run another model experiment using an observed cloud dataset during one or more of the last few ENSO events. The COADS database may be one source for this cloud data. Alternatively, a simple cloud–SST feedback mechanism may be included in the formula for $Q_0$, such that $C$ increases to a particular value at some threshold of SST.

This work has shown that interannual variations in the heat transport by meridional overturning are not wholly due to changes in the overturning circulation driven by the variations in the zonal wind stress. The temperature of the surface layer and temperature of the equatorward interior flow also vary on the interannual timescale and contribute significantly to variations in the heat transport by meridional overturning. The heat transport by gyre exchange also varies significantly on the interannual timescale, depending on changes in the undercurrent transport and changes in the zonal temperature gradient. The disappearance of the undercurrent produces a large poleward heat transport anomaly by this mechanism. The net heat export observed during ENSO events is begun by an export due to this mechanism. The variations of net heat transport by the transient eddies are generally much smaller than the variations in the other mechanisms. This result is at odds with the hypothesis of Giese and Harrison (1991), which maintains that the transient eddy heat transport increases before ENSO events due to increased easterly wind stress.

An interhemispheric exchange of heat into the Northern Hemisphere occurs during periods of massive poleward heat export accompanying the warm phase of ENSO. Heat is transported into the Southern Hemisphere during periods of net heat import to the equator. A smoothed time series of cross-equatorial volume transport in the model indicates the northward volume transport during the warm phase and southward volume transport during the cold phase of ENSO, which agrees with the sea level study of Wyrtki and Wenzel (1984). It is suggested that interannual variations in the zonally integrated wind stress curl account in part for the variations in the cross-equatorial transport.

A similar interhemispheric heat exchange also occurs on the seasonal timescale as discussed in Brady and Gent (1994). However, there is a symmetry in the seasonal exchange in that heat is transported across the equator into the winter hemisphere and there is a net export from the equatorial region. There is a small net import of heat into the equatorial region during the transition seasons and also a cross-equatorial transport into the spring hemisphere. Based on the results presented here and the work of Wyrtki and Wenzel (1984), the same symmetry in the pattern of cross-equatorial heat transport does not hold for interannual phenomenon. Instead, heat is transported northward during episodes of near-equatorial heat export and southward during episodes when there is an import of heat in the near-equatorial region. Northward transport during export occurs in the last four decades, as shown by Wyrtki and Wenzel (1984), but this could also be due to an enhancement of the seasonal cycle during the northern fall and winter.

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