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On the processes that control seasonal variations of sea surface temperatures in the tropical Pacific Ocean

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ABSTRACT

In the tropical Pacific Ocean, the cold phase of both interannual and seasonal sea surface temperature variations is characterized by cold waters off the coast of South America and a pronounced equatorial tongue of cold surface waters in the eastern Pacific. The warm phase, in both cases, is marked by the weakening or complete absence of these features. Despite these striking similarities, very different physical processes are dominant on seasonal and interannual time scales. Interannually, a horizontal redistribution of warm surface waters, the dynamical response of the ocean to changes in the winds, is of primary importance. What matters most seasonally are two local processes: seasonal upwelling associated with a divergence of surface currents, and the seasonal modulation of mixing processes, by heat fluxes, that control to what extent upwelling induced by the mean winds influences sea surface temperatures. These results shed light on the different requirements that coupled ocean-atmosphere models should meet if they are to reproduce both seasonal and interannual variability. The results also make a case for measurements, along a meridian in the eastern tropical Pacific, that focus on the relations between sea surface temperature changes, heat flux variations and mixing processes.

1. Introduction

The seasonal changes in sea surface temperature in the tropical Pacific, shown in Fig. 1, are remarkably similar to the interannual changes associated with the Southern Oscillation. The cold phase of both the seasonal and interannual fluctuations is characterized by low sea surface temperatures off the coast of South America and a pronounced equatorial tongue of cold surface waters in the eastern Pacific. The warm phase of both fluctuations is marked by the weakening or complete absence of these cold features. The interannual variations have been studied in great detail (see Philander (1990) for a review) and are known to be determined by the dynamical response of the upper ocean to changes in the winds. Given the similarities between the seasonal and interannual sea surface temperature variations it seems reasonable to assume that seasonal changes too are governed by the dynamical response of the ocean to variations in the wind. Some investigators suggest that the response to the meridional component of the wind is of particular importance in the eastern tropical Pacific. Philander and Pacanowski (1981) showed that a model forced with strictly north–south winds can reproduce a realistic sea surface temperature pattern. Mitchell and Wallace (1992), on the basis of data analyses, also conclude that variations in the meridional winds contribute significantly to seasonal sea surface temperature changes. Seager et al. (1988) use an oceanic model to argue that, because of the sensitivity of the equatorial zone to zonal winds, the relatively modest seasonal change in the zonal winds has a substantial effect on sea surface temperatures.

The measurements in Fig. 2, temperature fluctuations on the equator at 110° W as a function of

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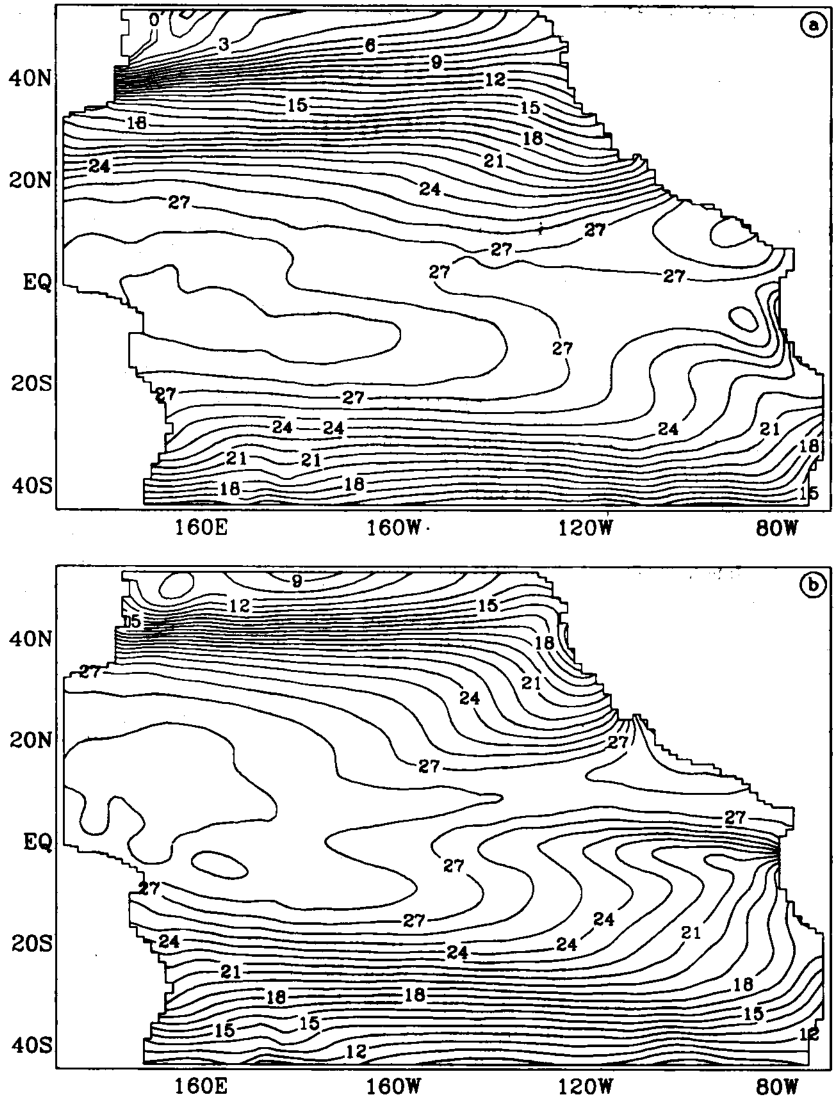

Fig. 1. Climatological sea surface temperature from COADS for March (a) and September (b); contour interval 1 K.

depth and time, indicate that there is more to the story of seasonal sea surface temperature changes than a strictly dynamical response to the winds. In Fig. 2, the high sea surface temperatures in March and April are associated with increased stratification of the upper 50 m, and with an increase in the local flux of heat into the ocean. The low sea surface temperatures in September and October coincide with a minimum in the local heat flux. These results, and the absence of pronounced vertical movements of the isotherms below 50 m, suggest that, at 0° N 100° W at least, seasonal changes in sea surface temperature depend on near-surface processes and heat fluxes, rather than vertical movements of the thermocline as part of the dynamical response of the ocean to changing winds. If heat fluxes were solely responsible for those changes then, in the simplest model,

$$\rho c_p H T_t = Q, \tag{1}$$

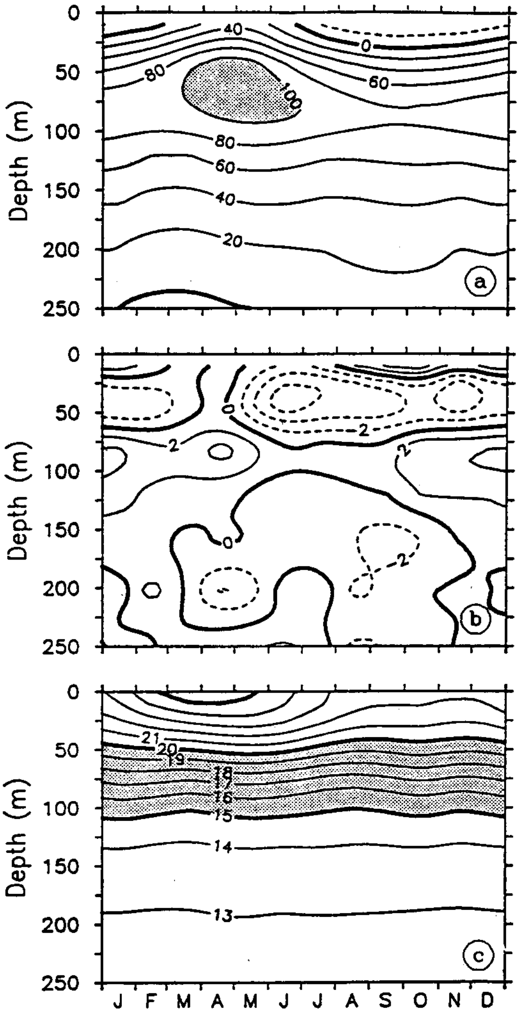

Fig. 2. Climatological seasonal variations at 0° N and 110° W of zonal velocity (a), meridional velocity (b) and temperature (c). Units are °C and cm/s. Dashed contours are for westward or southward flows. Shadings highlight zonal velocities over 100 cm/s and temperatures between 15°C and 20°C. Courtesy M. McPhaden, 1993, unpublished.

there ought to be a 90° phase lag between temperature T and heat flux Q (here t denotes time, ρ the mean density of the ocean, c_p the heat capacity of water and H the depth of the mixed surface layer). That, however, is not observed. Hayes et al. (1991), in their analyses of the data from 0° N 110° W concluded that the heat budget for the upper ocean at that location is a complicated one with the heat flux, upwelling and advection all playing roles. Exactly what is the relative importance of these processes?

The answer to this question depends very much on the region being considered. Over much of the oceans eq. (1) adequately describes changes in sea surface temperatures so that, seasonally, there is a 90° lag between T and Q. Eq. (1) even applies to large parts of the tropical oceans, the Indian Ocean and western tropical Pacific, although, in low latitudes, the depth H of the surface layer of constant temperature depends on the dynamical response of the upper ocean to large scale winds (McPhaden, 1982). In the tropical regions just mentioned, seasonal sea surface temperature variations are very modest, especially in comparison with those in the eastern tropical Pacific. In the latter region the processes that control seasonal sea surface temperature changes are far more complex than those described by eq. (1). We therefore study that region, and the entire tropical Pacific, by means of an Oceanic General Circulation Model capable of realistic simulations of the seasonal cycle, especially of the cycle in sea surface temperature. We use this model to determine to what extent wind variations, in the absence of heat flux variations, can cause the observed sea surface temperature variations. We also address the complementary question: to what extent can heat flux changes, in the absence of wind fluctuations, cause the sea surface temperature variations? The latter question is of special interest when there are steady winds that maintain steady oceanic upwelling, near the equator for example. The upwelling does not affect sea surface temperatures in the absence of mixing processes because the vertical velocity component vanishes at the surface. Mixing processes permit upwelling to manifest itself in sea surface temperature patterns. This argument suggests that a seasonal variation in the intensity of mixing processes, caused by seasonal changes in the heat flux that stratifies the upper ocean, can cause seasonal changes in the sea surface temperatures. In other words, it is possible for the equatorial cold tongue to be enhanced in August and September, or to be diminished in March and April, even when the winds remain perfectly steady but provided enhanced heat fluxes inhibit mixing in the northern spring while diminished heat fluxes do the opposite in late summer.

The paper is organized as follows. The general circulation model is described in Section 2, the heat flux across the ocean surface in Section 3. Section 4 concerns the reference case which, in Section 5, is compared to the 2 other cases. Section 6 summarizes the results. The focus is on sea surface temperature variations. For a discussion of other aspects of the seasonal cycle in a model very similar to this one the reader is referred to Philander et al. (1987).

2. Model

The model is the widely used primitive equation ocean general circulation model (GCM) in its latest edition (Pacanowski et al., 1991). Our version uses constant horizontal diffusion and a horizontal friction that depends on the meridional size of the grid. The vertical diffusion and friction are Richardson number dependent as described by Pacanowski and Philander (1981).

The model configuration for the Pacific Ocean is a modified version of that of Philander et al. (1987) and of Chao (1990). The model domain covers the Pacific Ocean from 130° E to 70° W and from 45° S to 55° N. The realistic topography is derived from the one degree SCRIPPS topography, modified so that the minimum ocean depth is 100 m. All islands and some peninsulas are submerged to a depth of 100 m.

The resolution is a constant 1.5° in longitude. In latitude, resolution is a constant $\frac{1}{3}$° between 10° S and 10° N and is a constant 1° poleward of 20°. Between 10° and 20° latitude, there are 15 grid points with a separation that increases gradually. There are 27 levels with increasing thickness in the vertical (Table 1).

All lateral boundaries are closed walls. The influence of the ocean beyond the northern and southern boundaries (45° S and 55° N) is incorporated by restoring zones. Poleward of 35° S and 45° N temperature and salinity are damped back to values taken from the monthly climatologies of Levitus (1982) with a time scale that decreases with increasing latitude from 40 days to 2 days. The surface boundary condition for salinity is zero flux.

The surface wind stress is calculated from COADS monthly mean wind velocities using the formulation of Pacanowski (1987) which takes into account the oceanic surface currents. In the surface heat flux, to be discussed in more detail in Section 3, the short-wave radiation penetrates below the ocean surface in accord with the formula of Paulson and Simpson (1977).

Table 1. *Vertical resolution of model*

k level	Thickness of box (m)	Depth of w-points (m)	Depth of t-points (m)
1	10	10	5
2	10	20	15
3	10	30	25
4	20	50	40
5	30	80	60
6	40	120	100
7	50	170	145
8	60	230	200
9	70	300	265
10	90	390	345
11	110	500	445
12	130	630	565
13	150	780	705
14	160	940	860
15	170	1110	1025
16	180	1290	1200
17	190	1480	1385
18	200	1680	1580
19	210	1890	1785
20	220	2110	2000
21	230	2340	2225
22	240	2580	2460
23	250	2830	2705
24	250	3080	2955
25	250	3330	3205
26	250	3580	3455
27	250	3830	3705
28			3830

Three sets of calculations were performed with the model. In the first, the reference case, the model is initially at rest and has the temperature and salinity fields that correspond to the climatological fields for January as described by Levitus (1982). The model is then forced for 5 years with the climatological winds and parameterized heat fluxes that require specification of the air temperature, relative humidity, cloud coverage, and wind speed but which depend on the sea surface temperature produced by the model. With this forcing, the model reaches equilibrium and reproduces realistic seasonal variations of the currents and thermal fields. The calculations were continued for two more years and heat fluxes during this time stored for further experiments; mean annual heat flux and wind stress were calculated from the second year of this run. The last 2 years of the calculations were then repeated with a change in surface boundary conditions: specified heat fluxes, to be described in the next section, rather than parameterized heat fluxes. (The final Section 5 of the paper discusses the implications of a specified heat flux). The results to be shown are from year 7 of these calculations. In the next set of calculations, case *cH*, the heat fluxes are assigned their annual mean values while the winds vary seasonally as before. The last 2 years of the reference case are then repeated with this modified forcing. In the third set of calculations, case *cW*, the wind stresses are assigned their annual mean values while the heat fluxes have realistic seasonal variations. The final 2 years of the reference case are then repeated.

In cases *cH* and *cW*, the seasonally varying heat fluxes and wind stresses are regarded as two independent forcing functions. Note that this assumption has an inconsistency because latent heat fluxes depend on the winds. Hence, in the case *cW*, the winds that drive oceanic motion are steady but the imposed variable heat flux takes into account variations in latent heat flux because of variations in the wind speed. There is a similar inconsistency in the case *cH* which assumes that the heat flux is constant even though the winds that drive the ocean (and that cause evaporation) are variable. We proceed in this manner because the processes involved in the oceanic response to heat fluxes and wind stresses are very different. Numerous realistic simulations of the sea surface temperature changes during El Niño establish, along with direct measurements, that variations in heat flux are of secondary importance to interannual sea surface temperature variations in the tropical Pacific. What matters most is the interannual redistribution of warm surface waters because of the dynamical response of the ocean to changes in the wind. Those processes will be captured in our experiment *cH*. (Note that the Philander and Seigel (1987) realistic simulation of El Niño of 1982 is a similar experiment because it assumes zero heat flux into the ocean). The complementary experiment *cW* suppresses these processes and explores the other ones that affect sea surface temperature.

3. The heat flux

The heat fluxes are calculated from bulk formulas as in Rosati and Miyakoda (1988) using COADS data for the various atmospheric fields and sea surface temperatures predicted by the model. The results agree well, both in amplitude and spatial structure, with previous estimates, by Oberhuber (1988), for example.

The principal contributions to the heat flux variations in the topics are short wave radiation and latent heat flux. (Variations in long wave radiation and sensible heat fluxes are negligible). In Fig. 3b, the annual mean short wave radiation that reaches the ocean surface is asymmetrical relative to the equator because of asymmetrical clouds. A region in the central Pacific, just south of the equator is practically cloud free so that the short wave radiation is at a maximum there. To the west, tall cumulus convective clouds form over regions of high sea surface temperatures. To the east, low stratus clouds that form over the cold surface waters significantly reduce the heat flux into the ocean. Whereas the convective cloud cover increases as sea surface temperature increases (Gadgil et al. (1984)), the stratus cloud cover increases as temperatures decrease towards the coast of South America. The sum of the shortwave radiation modified by clouds, and the oceanic heat loss, primarily because of evaporation shown in Fig. 3a, is a net gain that is essentially confined to the neighbourhood of the equator in Fig. 3c.

The seasonal variations of the different components of the heat flux are shown in Fig. 4 for a small region in the eastern equatorial Pacific. The short wave radiation (b) has a clear semi-annual signal but its second peak in September is significantly reduced by the dense stratus cloud cover over the very cold water at the time so that an annual signal is prominent. Evaporation from the ocean (a) is such that it reinforces the annual and practically eliminates the semiannual harmonic so that the heat flux into the ocean (c) has a single maximum during the early months of the year and minimum in June and July.

The results in Fig. 4 are for a region that extends from 4°S to 4°N in the eastern equatorial Pacific. Fig. 5b shows how sharp the latitudinal gradients are in that region. Variations are at a minimum right at the equator. Curiously, the southern

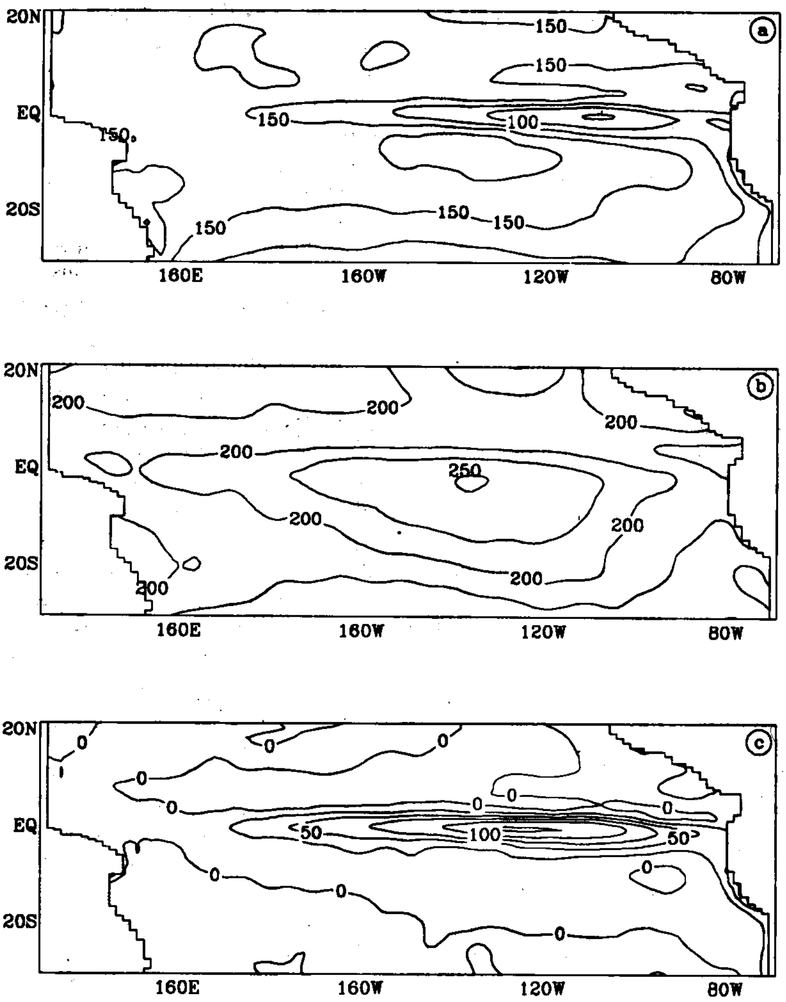

Fig. 3. Annual mean of latent heat flux out of the ocean (a), short wave radiation into the ocean (b), and total net flux into the ocean (c), reference case. Contour interval 25 W/m^2.

seasons of heat loss (centred on June), and heat gain, (centred on December) extend across the equator into the northern hemisphere. This is presumably a reflection of the southeasterly trade winds that penetrate into the northern hemisphere. This asymmetry relative to the equator decreases to the west so that, in Fig. 5a, a semiannual cycle is more prominent along the dateline than along 110°W in Fig. 5b.

The amplitudes of the annual harmonic of heat flux components and of the total heat flux are shown in Fig. 6. The short wave radiation in Fig. 6a has a minimum in the equatorial zone because the sun moves back and forth across the equator and forces only a weak annual harmonic at the equator. The evaporation in Fig. 6b has a maximum around the mean position of the ITCZ where the winds fluctuate the most and where the high sea surface temperatures permit high evaporation. Because this is a feature of the eastern tropical Pacific, latitudinal gradients of the total heat flux into the ocean, in Fig. 6c, are stronger in the eastern than western tropical Pacific. The conspicuous maximum off the coast of Peru in Fig. 6c is due to large variations in both cloudiness and evaporation.

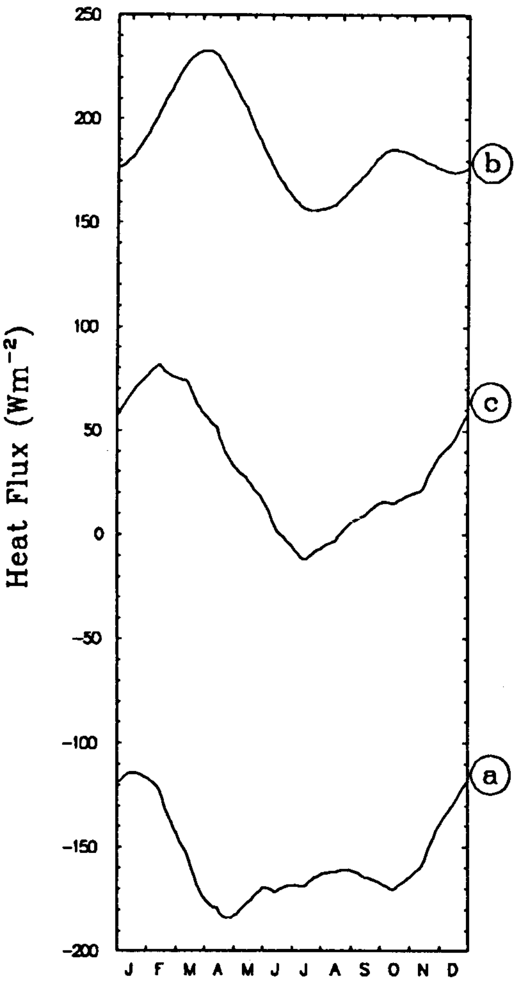

Fig. 4. Seasonal variation of latent heat flux (a), shortwave radiation (b) and total net flux into the ocean (c), reference case. Integrated over a box from 4°S to 4°N, 86°W to 104°W and from surface down to 50 m. All units in W/m^2.

4. Results

The reference case, in which the model is forced with the climatological wind stresses and heat fluxes, reproduces various aspects of the seasonal cycle realistically. The fluctuations of the currents and of the subsurface thermal structure agree with measurements (such as those in Fig. 2). The sea surface temperatures, on which this paper focuses, are also realistic. The difference between the observed and simulated annual mean sea surface temperature is at most 0.5 K. The simulations of the seasonal variations in sea surface temperature are equally accurate as is evident in maps of the amplitude shown in Fig. 7 and even phase of the annual harmonic, not shown here. Note that the model captures the westward phase propagation along the equator of the maximum (and minimum in sea surface temperature). Fig. 8 includes results from case *cH* and *cW* and permits an evaluation of the relative importance of heat fluxes and the dynamical response of the ocean. From case *cH*, it is clear that the importance of the dynamical response is essentially confined to the southeastern equatorial Pacific. The thermocline is shallowest in that region so that variations in wind-induced upwelling readily affect sea surface temperatures in that region. In the west, and far from the equator, the dynamical response is of secondary importance and sea surface temperatures depend primarily on the local flux of heat. It is therefore expected that case *cW*, in which the wind stress is constant while the heat flux varies seasonally, should do well far from the equator. What is surprising in Fig. 8 is how well case *cW* performs in the eastern equatorial Pacific even though it fails to take into account the dynamical response of the ocean. Although for a restricted time (February 1979 to June 1980) and area (150°W to 158°W and 17°S to 21°N), Stevenson and Niiler (1983) found a high correlation between heat flux analyzed from measurements and sea surface temperature measurements.

It is interesting to compare Fig. 6c, the forcing function, and Fig. 8c, the response. Sharp latitudinal gradients and a maximum north of the equator in the eastern Pacific characterize the forcing function but are absent from the response because the mean sea surface temperatures are high in that region where the thermocline is deep. In this narrow band north of the equator in the eastern Pacific, heat fluxes can not affect sea surface temperature very much. However, in the neighbouring regions, to the north and south, off the coast of the Americas, where the mean sea surface temperatures are relatively low and the thermocline is shallow, the seasonal changes in the heat flux strongly influence sea surface temperatures.

Case *cW*, in which the winds are steady, reproduces much of the seasonal changes in sea

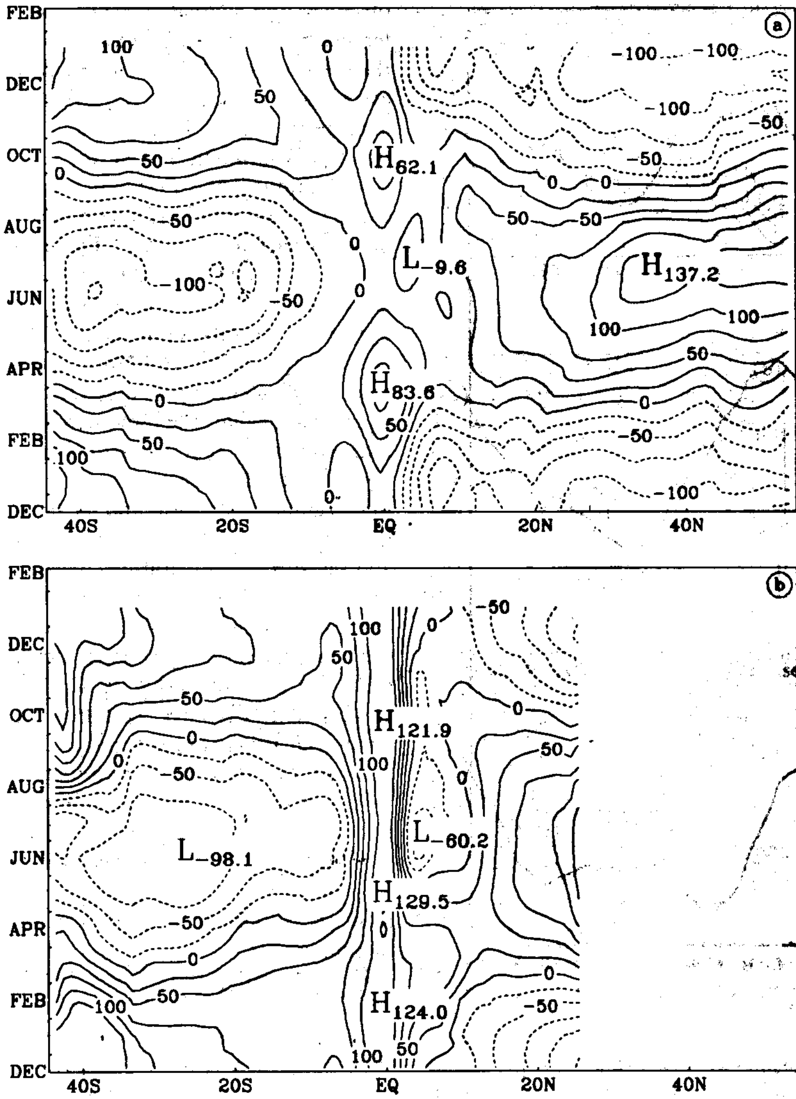

Fig. 5. Net heat flux into the ocean as a function of time and latitude along the dateline (a) and along 110°W (b), reference case. Contour interval 25 W/m^2.

surface temperature even though it has practically no seasonal changes in temperature below a depth of 50 m approximately. This can be seen in Fig. 9 which shows north–south sections of the thermal field, at the extremes of the seasonal cycle, for the reference case (top panels) and for case *cW* (bottom panels).

In case *cW*, enhanced heat fluxes increase the stratification of the upper ocean in the region north of the equator during the northern summer. Reduced heat fluxes during the winter allow intense mixing and isotherms that are practically vertical in the upper ocean as in Fig. 9c. Whereas there is a pronounced seasonal spreading of the equatorial thermocline in the reference case (the upper panels of Fig. 9), such a seasonal change is almost absent from case *cW*. The spreading of the thermocline indicates an intensification of the

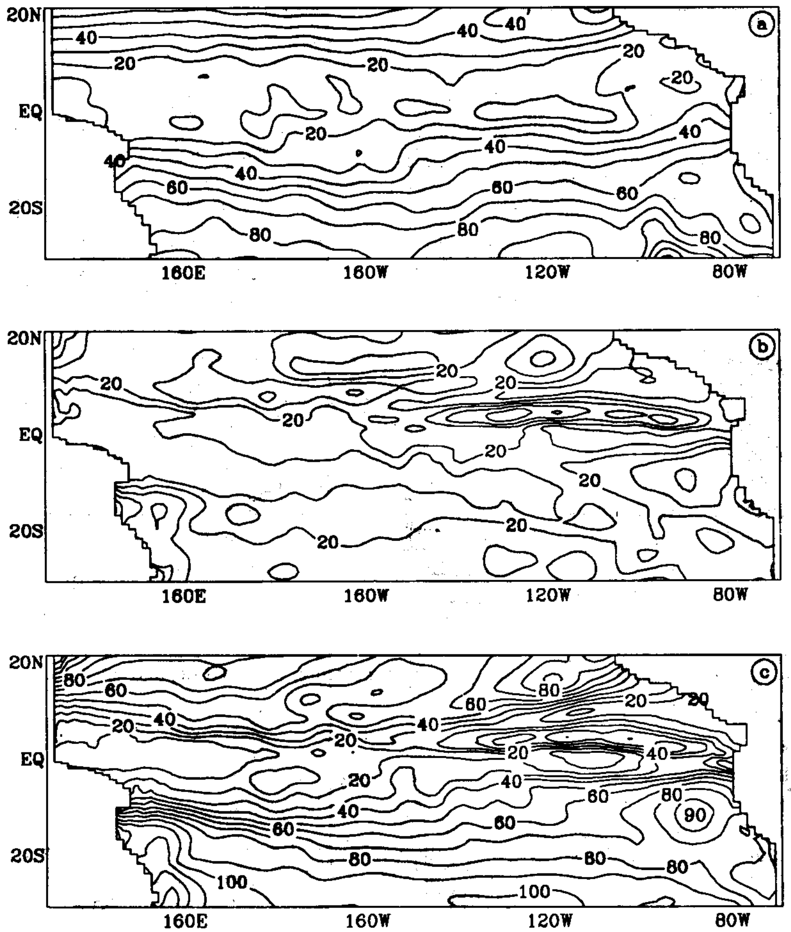

Fig. 6. The amplitude of the annual harmonic of short wave radiation into the ocean (a), latent heat flux out of the ocean (b), and total net heat flux into the ocean (c), reference case. Contour interval 10 W/m^2.

equatorial undercurrent. That current has almost no seasonal change in case *cW*, a very unrealistic feature. Case *cW* nonetheless captures much of the seasonal change in sea surface temperature.

Fig. 10 verifies that, for the location 0°N 110°W, vertical movements of the thermocline play a minimal role in the seasonal cycle. In case *cH*, for which the heat flux is constant, the seasonal changes in temperatures of the surface layer are controlled by seasonal changes in upwelling induced by the divergence of the local surface currents. When this divergence is strong, during the northern summer, then upwelling keeps the surface cold. When the divergence is weak then the steady heat flux stratifies and heats the surface layers. It is evident in Fig. 10 that at 110°W on the equator this case *cH* reproduces the seasonal variations in temperature more accurately than does case *cW*. In the latter case, in which the winds are steady the extremes in sea surface temperature fall short: the maximum in April is too low because the steady winds, which induce upwelling, are more intense than they should be at that time; the minimum in September is too high because the steady winds are less intense than they should be at that time.

The region where seasonal heat flux variations have the smallest amplitude is a narrow equatorial zone in the eastern Pacific. The heat budget for the region is shown in Fig. 11 for a surface box, 50 m

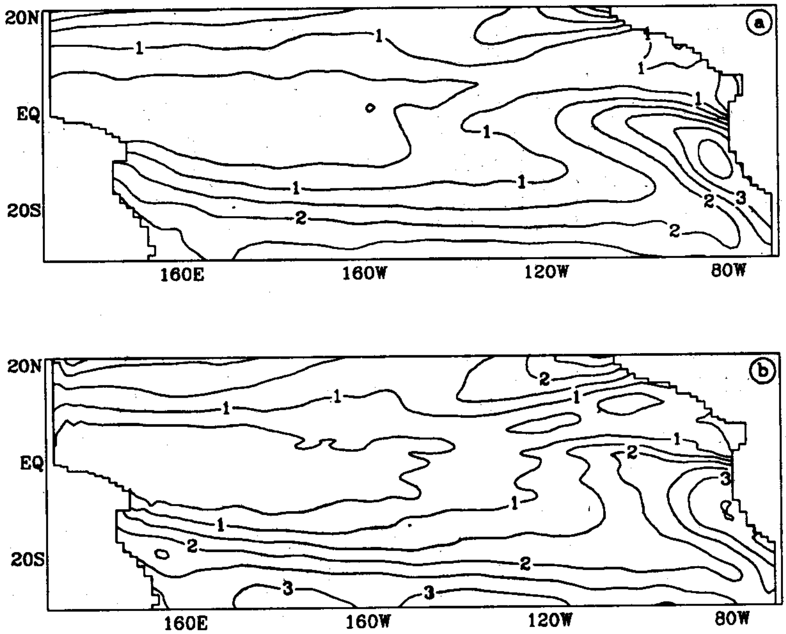

Fig. 7. Amplitude of the annual harmonic of sea surface temperature, observed (a) and reference case (b). Contour interval 0.5 K.

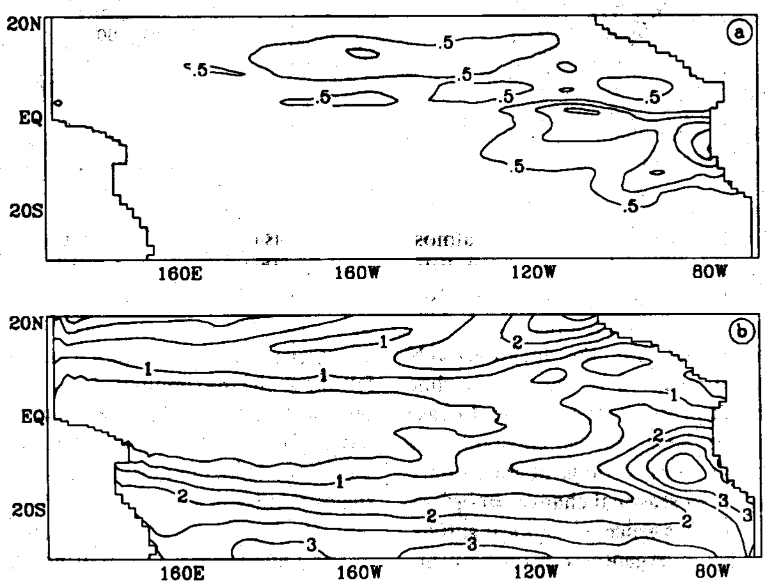

Fig. 8. Amplitude of annual harmonic of sea surface temperature for case cH (a), and case cW (b). Contour interval 0.5 K.

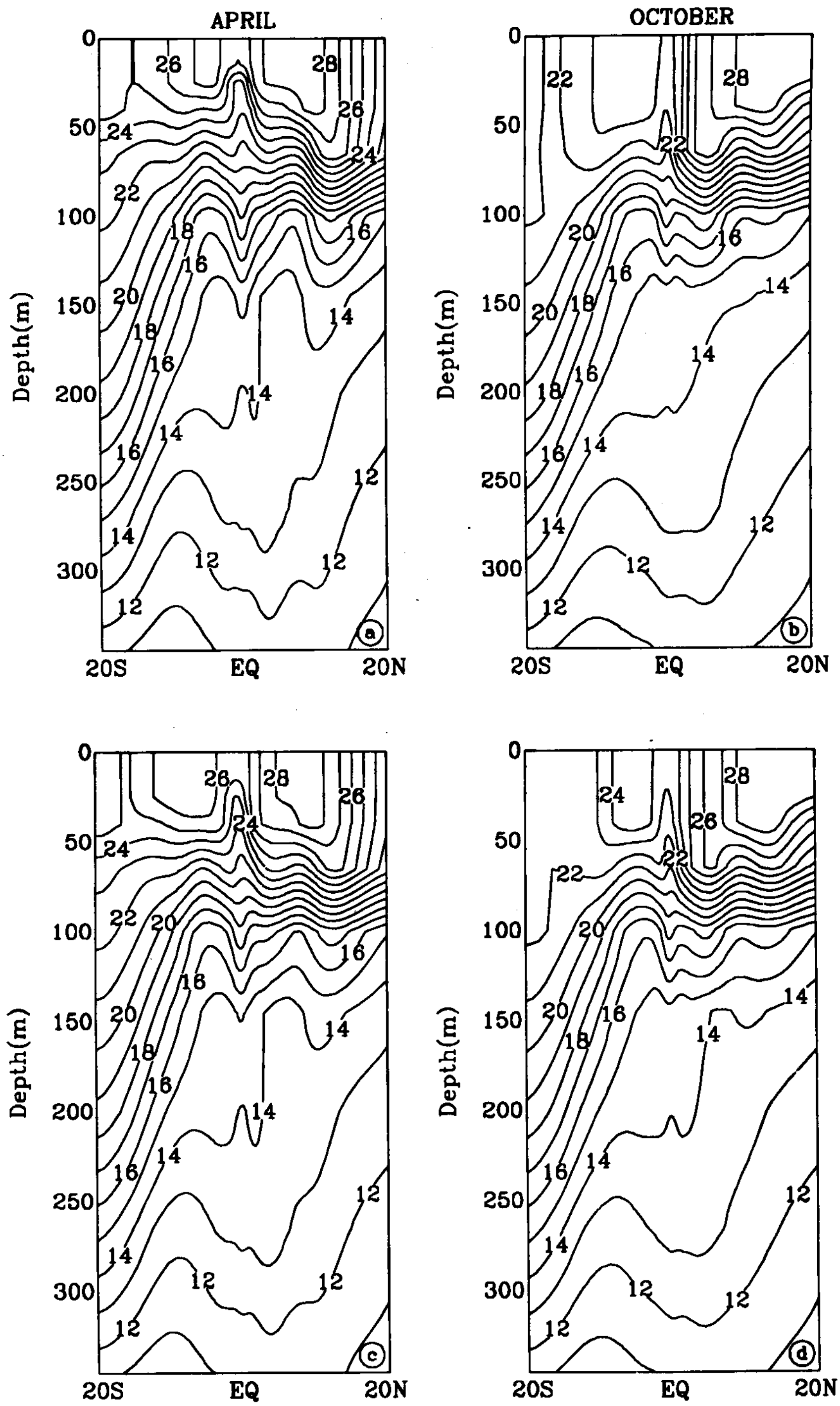

Fig. 9. Meridional temperature section at 110°W in April and October for reference case (a, b) and case *cW* (c, d). Contour interval 1 K.

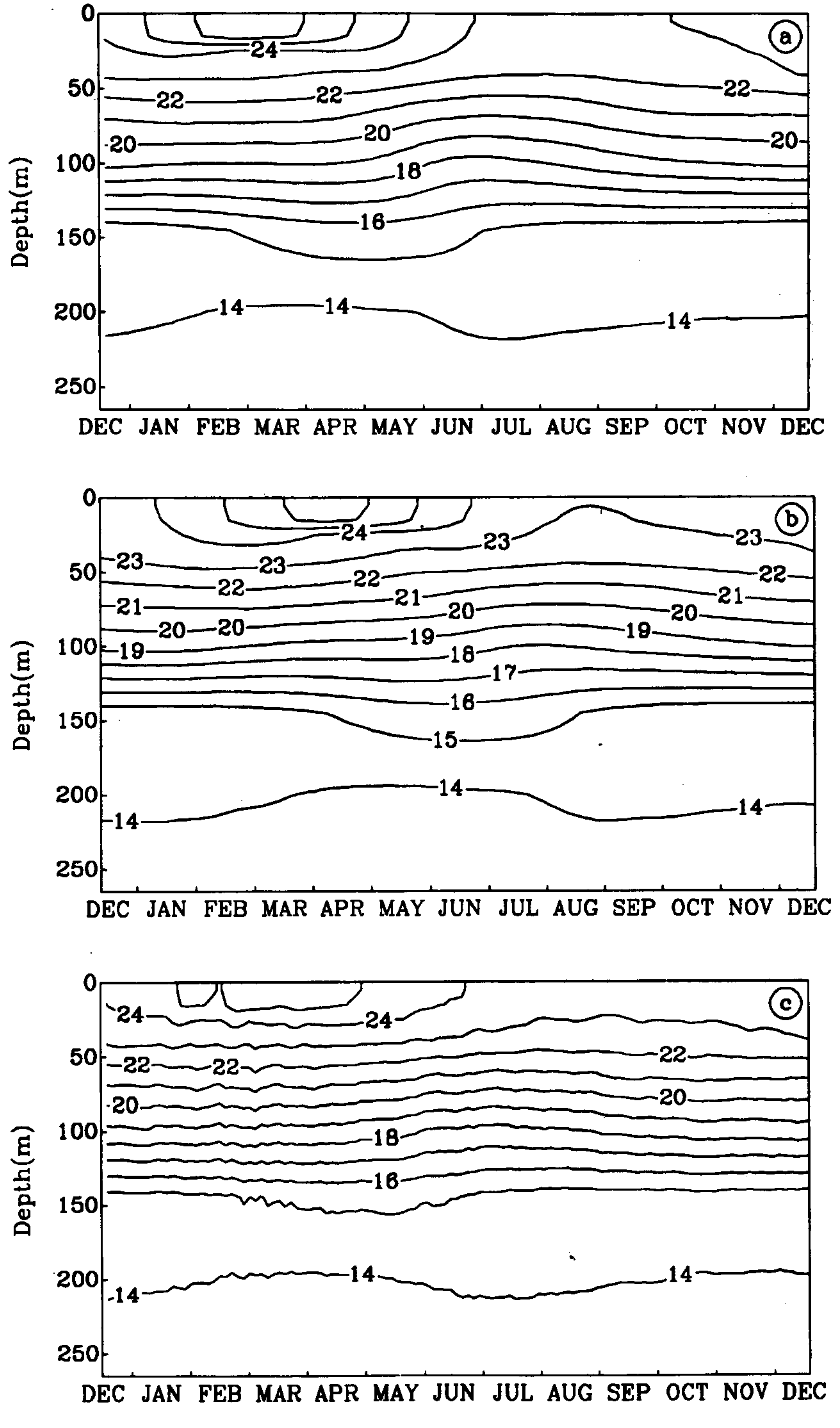

Fig. 10. Temperature as a function of time and depth at 0° N and 110° W for reference case (a), case *cH* (b), and case *cW* (c). Contour interval 1 K.

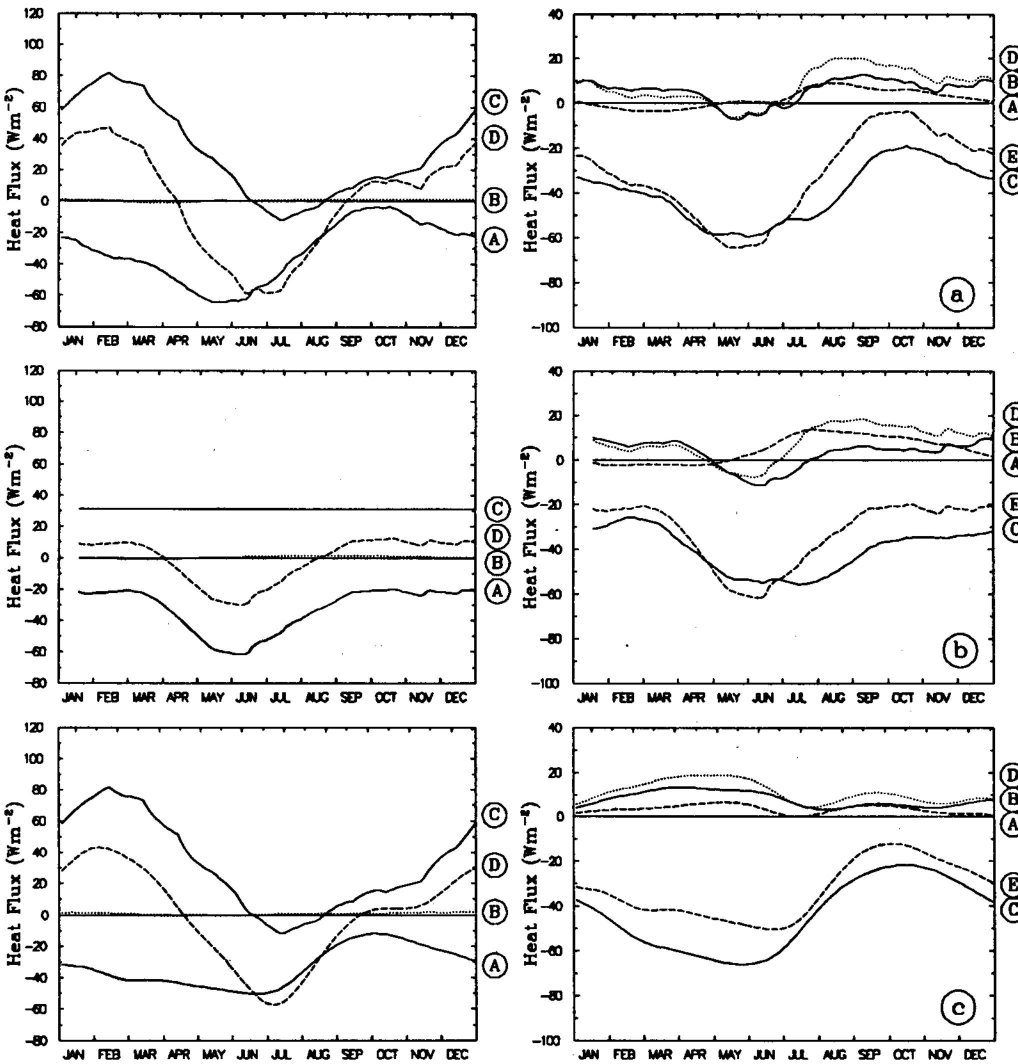

Fig. 11. Left panel: time rate of change of temperature (D) and the individual contributions from advection (A), diffusion (B), and net surface heat flux (C), right panel: time rate of change of temperature due to advection (E) and the individual contributions from zonal advection (A), meridional advection (B), vertical advection (C), and total horizontal advection (D) for reference case (a), case *cH* (b) and case *cW* (c). All quantities are integrated over a 50-m deep box that extends from 4° S to 4° N and from 86° W to 104° W. All results in case *cH* are lowpass filtered. All units are W/m^2.

deep, that extends from 4° S to 4° N and from 86° W to 104° W. The various curves in each panel of this figure correspond to different terms in the following equation:

$$T_z = Q - (uT_x + vT_y + wT_z) + (\nabla A_H(\nabla T) + (A_v(z)\, T_z)_z).$$

This is the temperature equation after it has been integrated over the volume of the box just mentioned. Q denotes the total net heat flux, the last two terms the horizontal and vertical diffusion (shown together in Fig. 11) with horizontal and vertical diffusion coefficients A_H and $A_v(z)$. In the advective terms, the three terms in parentheses, x,

y and z denote eastward, northward and upward directions, u, v and w denote the corresponding velocity components. In the reference case, Fig. 11a there is a positive heat flux into the box at practically all times but temperatures nonetheless decrease between April and September. This cooling is caused by the advective terms, primarily by upwelling as is evident from the importance of the term $wT_z(C)$ in Fig. 11a, right panel. This term can be decomposed into an annual mean value (denoted by a bar) plus a departure from the mean value (denoted by a prime) so that

$$wT_z = \bar{w}T'_z + w'\bar{T}_z + w'T'_z + \overline{wT}_z.$$

Seasonal variations depend primarily on the first two terms on the right hand side (the 3rd term has a numerical value that is much smaller). In case cW, the wind stress is constant and w' is essentially zero so that seasonal variations in wT_z depend mostly on $\bar{w}T'_z$, steady upwelling of a variable stratification. In case cH, on the other hand, the heat flux is constant and the term wT_z varies seasonally primarily because of variations in $w'\bar{T}_z$, variable upwelling of a steady stratification.

In the eastern equatorial Pacific the heat budget of the upper ocean is essentially a balance between the terms (b), (c), and (d) in the following equation:

$$\underset{(a)}{T_t} + \underset{(b)}{\bar{w}T'_z} + \underset{(c)}{w'\bar{T}_z} = \underset{(d)}{Q}$$

In words, the seasonal cycle variations in heat flux are balanced by seasonal variations in the upwelling of a steady stratification plus the steady upwelling of a seasonally varying stratification. Away from this region in the eastern Pacific the seasonal cycle is essentially a balance between terms (a) and (d). Neither of these regimes describes the interannual fluctuations which are associated with vertical movements of the thermocline and with a horizontal redistribution of warm surface waters. Although an experiment similar to cH can reproduce interannual sea surface temperature variations, the processes on interannual and seasonal time scales are different, because in the one case the wind variations induce large changes in the depth of the thermocline, in the other case they do not. For the interannual heat budget term (b) in the last equation should be written:

$$\bar{w}T'_z = \bar{w}(T_{\text{surface}} - \alpha\,\delta h)/H,$$

where H is the mean depth of the thermocline, δh represents thermocline displacements, T_{surface} is the sea surface temperature and α is a constant. On interannual time scales this term $\bar{w}T'z$ is the dominant one in the heat equation so that sea surface temperatures are controlled by variation in the depth of the thermocline: $T_{\text{surface}} \sim \alpha\,\delta h$. On seasonal time scales the depth of the thermocline hardly varies so that $\delta h \sim 0$ and the appropriate balance is the one discussed before.

In summary, these calculations indicate that seasonal variations in sea surface temperatures depend (i) on the variations in heat flux which modulate the effects of time-mean upwelling and (ii) on variations in upwelling that entrains cold water across the thermocline while leaving the depth of the thermocline essentially unchanged. Both these processes depend critically on the time-mean response of the ocean to steady winds.

5. Discussion

Certain aspects of the calculations described here can be justified only if the motivation for these studies is taken into account. The justification of the heat flux as a surface boundary condition is problematic. If, in the reference case, the calculations were continued for an extended period, several decades, then it is likely that sea surface temperatures would have drifted away, towards unrealistic values. The parameter that constrains sea surface temperature the most is the air temperature. But that parameter, in turn, depends on sea surface temperatures. In other words, it is possible to cope with the factors that control sea surface and air temperatures only in a coupled ocean-atmosphere model. Attempts to develop such models have recently run into difficulties. Some, the one described by Philander et al. (1992) for example, succeed in simulating a realistic interannual Southern Oscillation when forced with annual mean solar radiation but have difficulty coping with the seasonal cycle. Others, the one developed at the University of California at Los Angeles (Mechoso, private communication, 1993), can simulate the seasonal cycle but have no Southern Oscillation. As yet, there is no coupled model that reproduces both a seasonal cycle and the Southern Oscillation (although there are models of the Southern Oscillation in which the seasonal cycle is simply specified as part of the background state.) The results presented in this paper suggest that the problem with the coupled GCMs is the following. Interannual sea surface temperature changes depend on the ocean's dynamical response to winds so that the atmospheric component of a coupled model should reproduce accurate large scale winds but not necessarily accurate heat fluxes to succeed with the Southern Oscillation. To reproduce the seasonal cycle, on the other hand, requires accurate heat fluxes, mean winds, and wind fluctuations that control local upwelling in the eastern Pacific. Large-scale wind fluctuations that redistribute surface waters are not particularly important.

In the results presented here, we bypassed some of the problems faced in coupled models by specifying both the winds and heat fluxes. In a coupled model only the incoming solar radiation is specified. At the equator, it has predominantly a semiannual, not annual cycle. In the eastern equatorial Pacific, however, the heat flux into the ocean has predominantly an annual cycle. Section 3 of this paper identifies part of the reason as reflective low stratus clouds over cold surface waters late in the northern summer. Hence it is possible that one root of the problem with some coupled models is poor parameterization of clouds in their atmospheric component.

The principal result of this paper is that, in the tropical Pacific, seasonal variations in sea surface temperature are crucially dependent on heat fluxes that modulate mixing processes in the ocean. Not only is it important that the ocean model be forced with accurate heat fluxes, it should also have an accurate parameterization of mixing processes. This particular model uses the Richardson-number-dependent scheme of Pacanowski and Philander (1981). There is a need for studies similar to this one with other parameterizations. Of particular interest are the recent calculations by Chang (1994), who used a completely different model of the ocean, one that has a Kraus–Turner mixing parameterization, to obtain similar results.

Finally, there is a need for observational studies in the eastern equatorial Pacific to establish how sea surface temperatures changes, heat flux variations and mixing processes are related. The valuable study of Hayes et al. (1991) concerns measurements at a single point, 110°W on the equator, a point where latitudinal gradients in heat flux are enormous (see Section 3). There is a need for a similar study over a band of latitudes, from 10°S to 10°N, for example.

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