The Seasonal Cycle and Its Modulation in the Eastern Tropical Pacific Ocean

DAIFANG GU AND S. G. H. PHILANDER

Program in Atmospheric and Oceanic Sciences, Princeton University, Princeton, New Jersey

MICHAEL J. MCPHADEN

Pacific Marine Environmental Laboratory, National Oceanic and Atmospheric Administration, Seattle, Washington

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ABSTRACT

Data for the period from 1985 to 1993 from TAO moorings along 110°W (5°S–5°N) and 140°W (2°S–9°N) describe the vertical, meridional, and temporal structure of the seasonal cycle of several variables. The results have a number of interesting features. The amplitude of the seasonal cycle is relatively constant in the surface layers but varies considerably at the depth of the equatorial thermocline where it was small before 1989, large thereafter. Also, vertical seasonal movements of the thermocline have little effect on sea surface temperatures. These seasonal variations are consistent with a westward propagating coupled ocean–atmosphere mode in the surface layers. Conversely, the low-frequency modulation of the seasonal cycle in the thermocline is associated with changes in the seasonal cycle of the zonal wind in the central and western tropical Pacific and might be attributable to equatorial Kelvin waves forced resonantly by the surface winds.

1. Introduction

Interactions between the tropical Pacific Ocean and the atmosphere produce interannual El Niño events and also influence the seasonal cycle, which, in sea surface temperature (SST) and in the zonal component of the surface wind, is characterized by an annual harmonic with westward phase propagation at a speed near 50 cm s^{-1} (Horel 1982). Data analyzed by Gu and Philander (1995) indicate that in the surface layers of the equatorial Pacific this cycle corresponds to the oceanic aspects of a coupled ocean–atmosphere mode that has westerly wind anomalies to the west of the warm SST anomalies, a mode may be of the type investigated by Neelin (1991). Those winds cause a westward displacement of the warm surface waters, which in turn displace the wind anomalies farther westward. This mode propagates far westward when the thermocline is shallow during La Niña, but slows down and does not extend as far to the west when the thermocline is deep during El Niño. Near the equator this mode is confined to the upper ocean; measurements indicate that the annual harmonic has westward phase propagation only near the surface but propagates eastward in the equatorial thermocline at the depth of the Equatorial Undercurrent (Meyers 1979; Halpern 1987; McPhaden and Taft 1988; Halpern and Weisberg 1989). Off the equator the annual harmonic propagates westward at the surface and also in the thermocline (White et al. 1985; Kessler 1990).

This paper reexamines the seasonal cycle on the basis of data from the Tropical Atmosphere Ocean array along 110°W (5°S–5°N) and 140°W (2°S–9°N) for the period 1985–1993. The data confirm the previous results described above and, in addition, provide a detailed documentation, for a specific period, of the relation between variations in the surface winds, currents, and thermal field; of the latitudinal and vertical structure of the seasonal cycle; and of the low-frequency modulation of its amplitude. A particularly remarkable result is that the annual cycle of the thermocline depth in the eastern tropical Pacific strengthens considerably after 1988; this property motivates us to present a possible mechanism for generating that phenomenon that involves resonance with the driving wind field. Hopefully the results presented here will prove helpful to developers of coupled ocean–atmosphere GCMs that attempt to simulate climate fluctuations in the Tropics. At present those models fail to reproduce a realistic seasonal cycle in the eastern equatorial Pacific (Mechoso et al. 1995).

The paper is organized as follows. Section 2 briefly discusses the data; section 3 describes the meridional structures of the seasonal cycles of winds, temperatures, and currents; and section 4 describes the modulation of the seasonal cycle of subsurface temperatures and proposes the resonant mechanism to account for the modulation. Section 5 discusses the results.

Corresponding author address: Dr. S. George Philander, Program in Atmospheric and Oceanic Sciences, Princeton University, P.O. Box CN710, Sayre Hall, Princeton, NJ 08544-0710.
E-mail: daifang@splash.princeton.edu

2. Data

The Tropical Atmosphere Ocean TAO Array of ATLAS (Autonomous Temperature Line Acquisition System) and PROTEUS (Profile Telemetry of Upper Ocean Currents) moorings, by the end of 1993, consisted of moorings at 63 locations spanning the domain 8°N–8°S, 137°E–95°W. The longest time series in the eastern Pacific are available at 110°W (5°S, 2°S, 0°, 2°N, 5°N) and 140°W (2°S, 0°, 2°N, 5°N, 9°N). Long time series are also available in the western Pacific along 165°E. The off-equatorial ATLAS moorings provide daily averaged temperature and wind speeds, the equatorial PROTEUS moorings provide those variables as well as horizontal currents. See McPhaden (1993) for a more detailed description of the data.

The data used in this investigation are from nine levels in the upper 200 m. Following procedures outlined in McPhaden and McCarty (1992), gaps in the data are filled by regression using the data from one layer above, or one layer below, or both whenever it is possible. A gap remains if the entire mooring is missing. The monthly mean time series are next obtained from these data; application of a 1–2–1 filter in time eliminates the residual effects of intraseasonal fluctuations that are dominated by instability waves with a period of 20 days (McPhaden and Taft 1988; Kessler et al. 1995). The errors are generally less than 0.5°C (McPhaden and McCarty 1992). These smoothed monthly mean time series, whose availability is shown in Fig. 1, are used to discuss the seasonal cycles in section 3. Further interpolation using optimal objective analysis (Knox 1976) yields data for the study of modulations of the seasonal cycles in section 4. A similar approach has been applied to the data at 0°, 165°E, which is discussed in section 5.

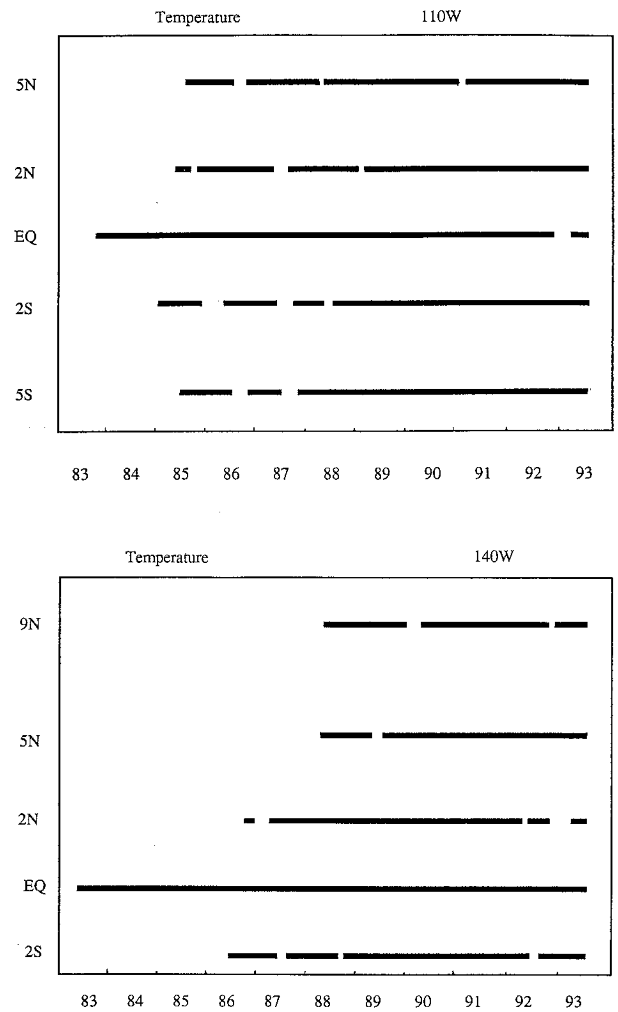

FIG. 1. The available subsurface temperature data at 110° and 140°W after vertical gaps have been filled.

3. Seasonal cycles

The northeasterly and southeasterly trade winds that prevail over the tropical Pacific converge in the intertropical convergence zone (ITCZ), which migrates seasonally between the neighborhood of the equator and approximately 12°N. The mean position of the ITCZ is near 9°N so that the meridional component of the wind reverses direction seasonally at that latitude. This northerly position of the ITCZ relative to the equator is the reason why seasonal changes in the surface winds have their largest amplitude north of the equator (Fig. 2). The phase of the seasonal cycle in the zonal winds has a 180° change somewhere between 2°N and 5°N at both longitudes; such a phase change is not clearly presented in the previous studies (e.g., Horel 1982). At the equator the winds are weak in March and April when the ITCZ is closest to the equator, and are intense in September when the ITCZ is farthest north. The seasonal changes at the equator propagate westward; the minimum appears first at 110°W, at 140°W a month later.

The winds force the ocean so that near the equator the oceanic currents too have westward phase propagation. In Fig. 3 it is evident that the phase of the seasonal variations of the zonal current increases by about a month between 110°W and 140°W. The maximum variability of the zonal currents is at a depth of 40 m at 0°, 110°W and 80 m at 0°, 140°W at the upper part of the thermocline. The amplitude of the seasonal cycle in the zonal current is about 40 cm s^{-1} at both longitudes at the depth of maximum variability. Sea surface temperature variations behave similarly. In analyzing Comprehensive Ocean-Atmosphere Data Sets (COADS), Gu and Philander (1995) found that the westward propagation of seasonal cycles of zonal wind and SST is such that the seasonal anomaly of the zonal wind is to the west of that of SST. This suggests that the anomalous wind, a response to the SST anomaly to its east, in turn influences the underlying SST through anomalous heat flux and upwelling, causing westward displacement of SST anomaly. This circular argument—the winds both

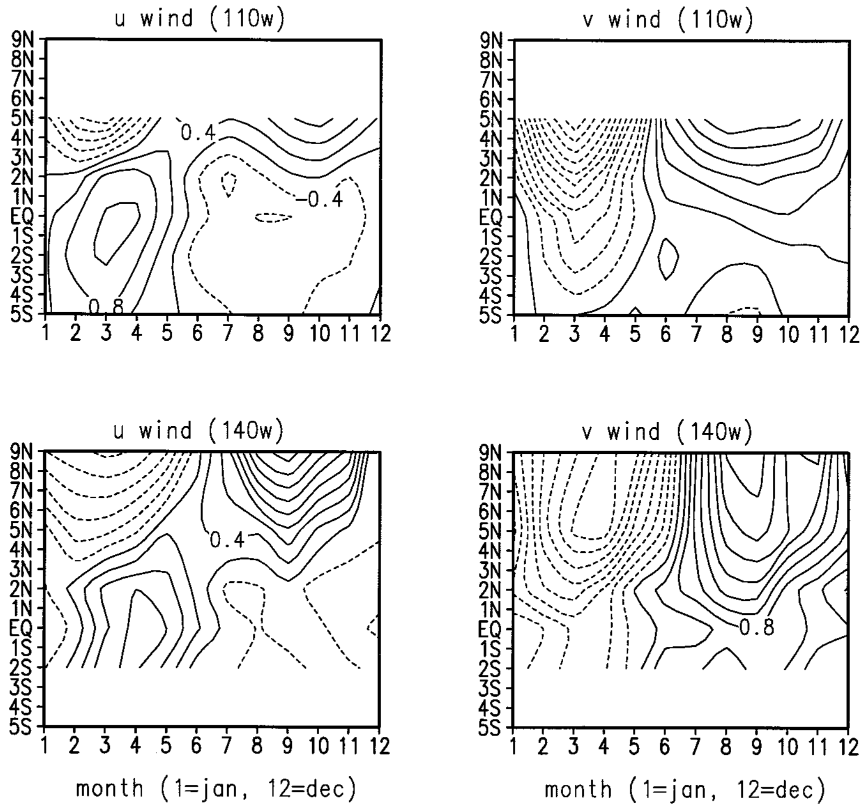

FIG. 2. Latitude–time plots of the seasonal cycle in zonal and meridional winds along 110° and 140°W.

influence and depend on the oceanic conditions—indicates that the westward phase propagation is a property, not of either medium by itself, but of the coupled ocean–atmosphere system. The mode under discussion is similar to that of the ''slow SST'' mode described by Neelin (1991) (also see Xie 1994). From a comparison of Figs. 2 and 3 it is evident that, at both 110°W and 140°W, the zonal currents lag the zonal winds by about a month, which is consistent with such a mode.

The contours in Fig. 4 show the time-averaged meridional thermal structures in the upper ocean along 110°W and 140°W; the arrows indicate the amplitude and phase of the annual harmonic. The phase in the upper ocean, at depths shallower than approximately 50 m, is very different from that at greater depths. [This aspect of the phase of the seasonal cycle is in sharp contrast to that of interannual variations, which have a phase that is nearly independent of depth from the surface layers through the thermocline. See Figs. 6c and 7c below and Koberle and Philander (1994).] An important implication for the theoretical study and modeling of the seasonal cycle in the eastern tropical Pacific is that vertical excursions of the thermocline, while of crucial importance to interannual variability, play a secondary role in the seasonal cycles of SST and zonal wind. The phase of the seasonal cycle of temperature in the surface layer is more or less uniform from 5°S to 5°N, with maximum temperatures in April along 110°W, and between May and June along 140°W. At depths greater than 50 m, in and below the thermocline, the phase varies from latitude to latitude but is practically independent of depth. These results indicate that thermal changes at depth are caused by vertical movements of the thermocline. Figure 5 shows that the associated phase propagation in the thermocline is eastward near the equator, westward in higher latitudes, a result consistent with the analyses of Meyers (1979). It is clear that the seasonal cycles near the surface and in the thermocline are governed by different processes. Whereas coupled ocean–atmosphere modes are important near the surface, the seasonal variations in the thermocline appear to correspond to Kelvin waves near the equator and Rossby waves off the equator. We return to this matter shortly.

4. Modulation of the seasonal cycle

We next turn our attention to interannual variations of the seasonal cycle. Figures 6 and 7 depict the bandpass filtered monthly mean time series of temperature, at 0°, 110°W and 0°, 140°W, obtained by using a Bartlett filter (Jenkins and Watts 1968) centered at the annual period. This filter allows us to separate seasonal from interannual variability. A striking feature at 110°W, but

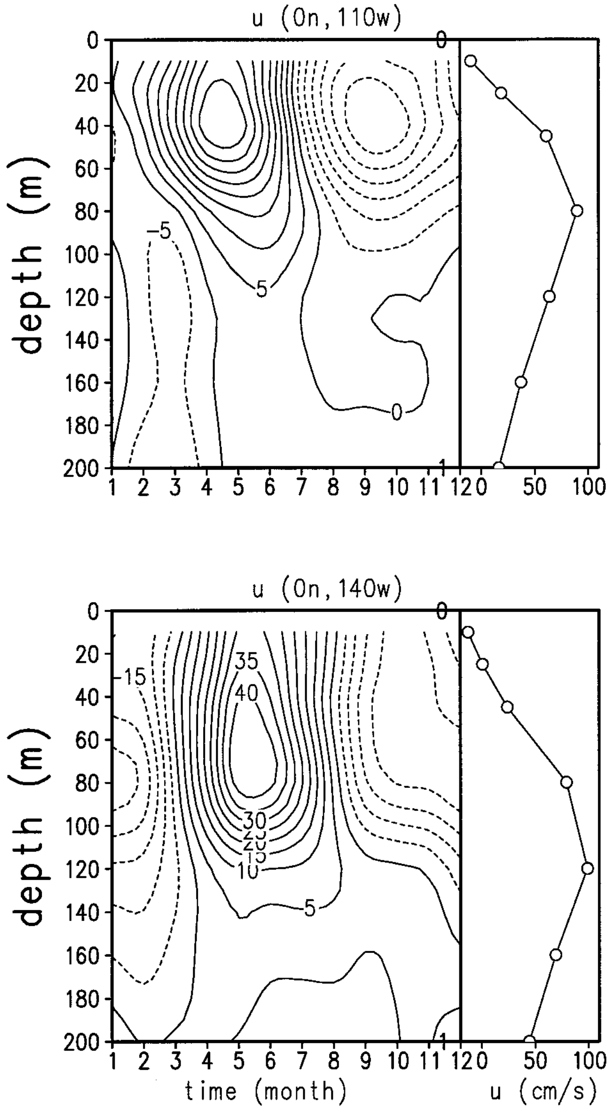

FIG. 3. The mean (on the right of the panel) and seasonal variations (on the left of the panel) of zonal currents at 110°W (upper panel) and 140°W (lower panel) with contour interval 5 cm s^{-1}. Negative values indicate westward flow.

less prominent at 140°W, is the change in the seasonal variability of the thermocline around 1989. At first there are almost no seasonal excursions of thermocline, but after 1989 the amplitude of those excursions increases significantly. Note that these large modulations of the amplitude of the seasonal cycles at depth are not evident at the surface, as the amplitude of the seasonal cycle of SST remains relatively unchanged throughout the record (Figs. 6a and 7a). In addition, after 1989 the phase of the seasonal cycles of sea surface temperatures changes by about one month at 140°W, with the maximum arriving earlier in the year. A similar change is absent at 0°, 110°W so that the westward propagation speed of the annual SST signal increased after 1989.

The upper and middle panels in Fig. 8 contrast the time mean and annual harmonics across 110°W for the years 1985–87 and 1990–92 respectively, and the lower panel shows the difference in the mean and annual cycle of temperature between the earlier and the later period. The ocean is seen to be warmer and the thermocline deeper in the later period. The largest change occurs at the depth of the thermocline, especially south of the equator (>1°C). This warming is intriguingly different from that associated with El Niño. Whereas the interannual deepening of the thermocline in the east during El Niño is accompanied by a shoaling of the thermocline in the west so that in effect there is a horizontal redistribution of warm surface waters, the deepening of the thermocline in the east, evident in Fig. 8, was apparently not accompanied by a compensatory shoaling in the west. (The subsurface warming in the east is evident in Figs. 6 and 8; data from the mooring at 0°, 165°E show the 20°C isotherm at essentially the same depth during the El Niño episodes of 1987 and 1991–92.) Thus, the warming seen in Fig. 8 can be viewed as an aspect of a decadal climate fluctuation that is likely governed by processes different from those that control interannual variations.

Associated with the changes in the subsurface seasonal cycle around 1989 is a change in the seasonal cycle of the zonal winds in the western and central equatorial Pacific. The mean annual cycle of sea surface temperatures and zonal wind speeds, before and after 1989, are shown in Fig. 9. [The results were obtained by means of a Bartlett bandpass filter applied to the COADS time series (Woodruff et al. 1987).] The annual cycle in zonal wind is seen to propagate westward across the basin between 1985 and 1988, but from 1989 to 1992 it propagates eastward between 160°E and 140°W, and the amplitude increases in the region from 160°E to the date line. The propagation of the annual cycle of SST is still westward after 1989 but becomes more confined to the east.

Weisberg and Tang (1983) and Tang and Weisberg (1984) explored the response of the ocean to eastward moving wind patches and conclude that the Kelvin waves excited by such moving winds are responsible for the observed thermocline depth changes during El Niño. McCreary and Lukas (1986) studied the response of equatorial oceans to a moving wind patch and found that the resonant and near-resonant waves can be excited over other equatorially trapped waves forced by wind. McCreary and Lukas (unpublished work) extended this study to include propagation of winds confined within a wind patch of finite zonal extent and found a significant amplification of Kelvin and Rossby wave radiation when the propagation speed of the wind was close to the phase speed of the wave. Such near-resonance of Rossby waves in response to the westward propagating annual cycle of the wind field was responsible for the observed variability in the thermocline (Kessler and McCreary 1993).

These theoretical studies suggest that the large vertical

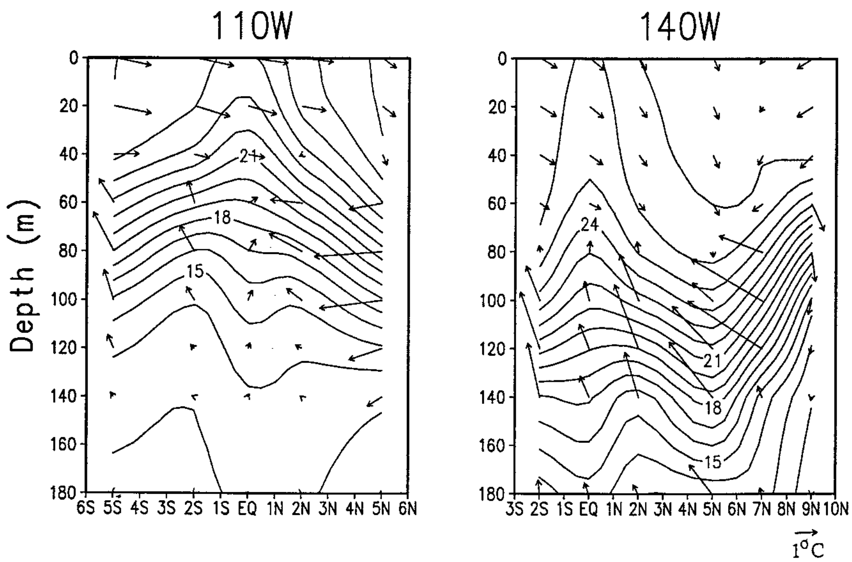

FIG. 4. The annual mean temperature along 110°W and 140°W. The contour interval is 1°C. The arrows indicate the amplitude and phase of the annual harmonics at locations where measurements are available. The length of an arrow corresponds to the amplitude. The length of the arrow in the right lower corner is 1°C. The direction of an arrow represents the phase. An upward pointing arrow indicates that the maximum of the annual cycle occurs in January, and a rightward pointing arrow indicates that the maximum of the annual cycle occurs in April.

displacements of the equatorial thermocline in the east after 1989 could correspond to oceanic Kelvin waves excited resonantly by the change in winds after 1989. To explore this possibility, we use a Kelvin wave model proposed by Kessler and McPhaden (1995). In this model, the thermocline excursion at 0°, 110°W is expressed as a

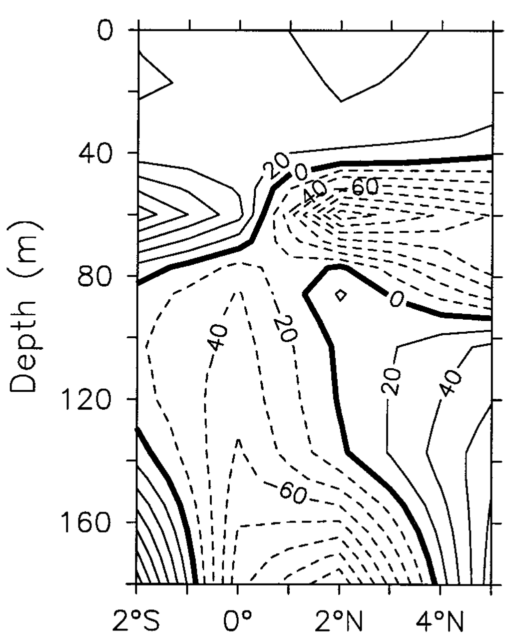

FIG. 5. The phase of the annual harmonic in temperature at 140°W minus that at 110°W. Negative values, which are dashed, indicate eastward propagation.

sum over vertical baroclinic modes, the response of each mode is an integration over the zonal winds along the Kelvin wave characteristic ($x - c_n t$), where c_n is the phase speed of the Kelvin wave associated with the nth baroclinic mode. The amplitude of the near-resonance response of the Kelvin waves is proportional to the length of the region where the zonal wind propagates eastward. The values of the baroclinic mode parameters for the first four modes are calculated from the mean equatorial density profile obtained during 46 equator crossings of the Hawaii–Tahiti Shuttle experiment (Kessler and McPhaden 1995). When the model is forced with the COADS zonal winds along the equator from 140°E to 110°W, for the period from 1983 to 1992, there is a distinct change in the amplitude of seasonal cycle in the thermocline displacement at 0°, 110°W around 1989 with larger amplitudes for the period from 1989 to 1992. The third mode and fourth mode, with phase speed 1.03 and 0.87 m s^{-1}, are in near resonance with the wind during the period from 1989 to 1992. The second mode with phase speed of 1.32 m s^{-1} is farther away from the resonance period but also shows an amplitude increase during this period. However, the magnitude of the increase in amplitude in this model is about 50%, smaller than the observed increase. This model appears to be able to account for part of the observed phenomenon but needs to be refined by including, for example, Rossby waves and other neglected processes.

5. Discussion

This paper has documented the seasonal cycles of winds, currents, and temperature based on TAO data at

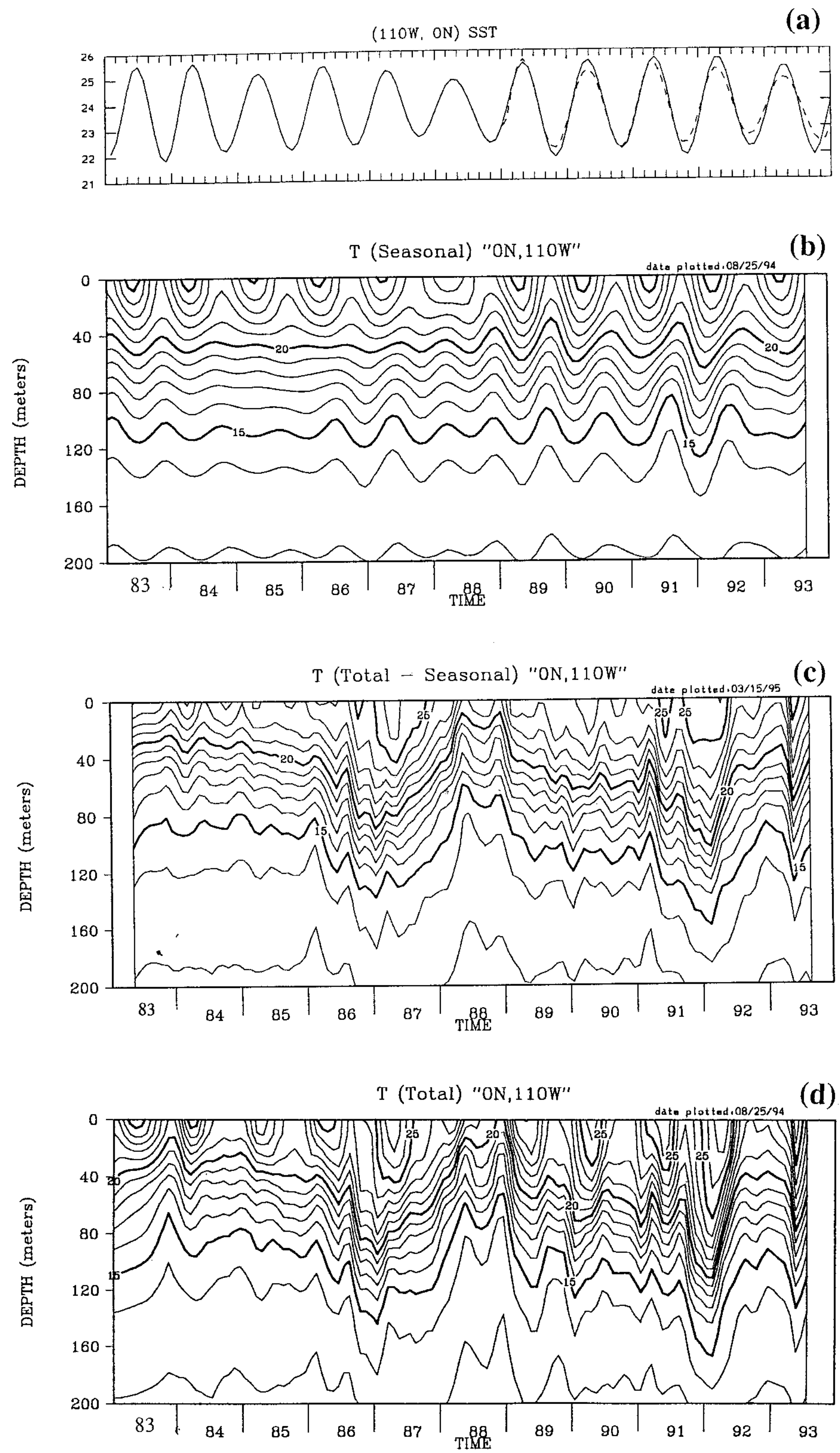

FIG. 6. (a) The seasonal variation of SST at 0°, 110°W. The dashed line superimposed on 1989–93 corresponds to the seasonal cycle for the period 1984–88. (b) The seasonal cycle of temperature in the upper 200 m at 0°, 110°W. (c) Interannual variations of temperature in the upper 200 m at 0°, 110°W. (d) The monthly mean temperature in the upper 200 m at 0°, 110°W. The sum of (b) and (c) is equal to (d). The contour interval is 1°C in all panels.

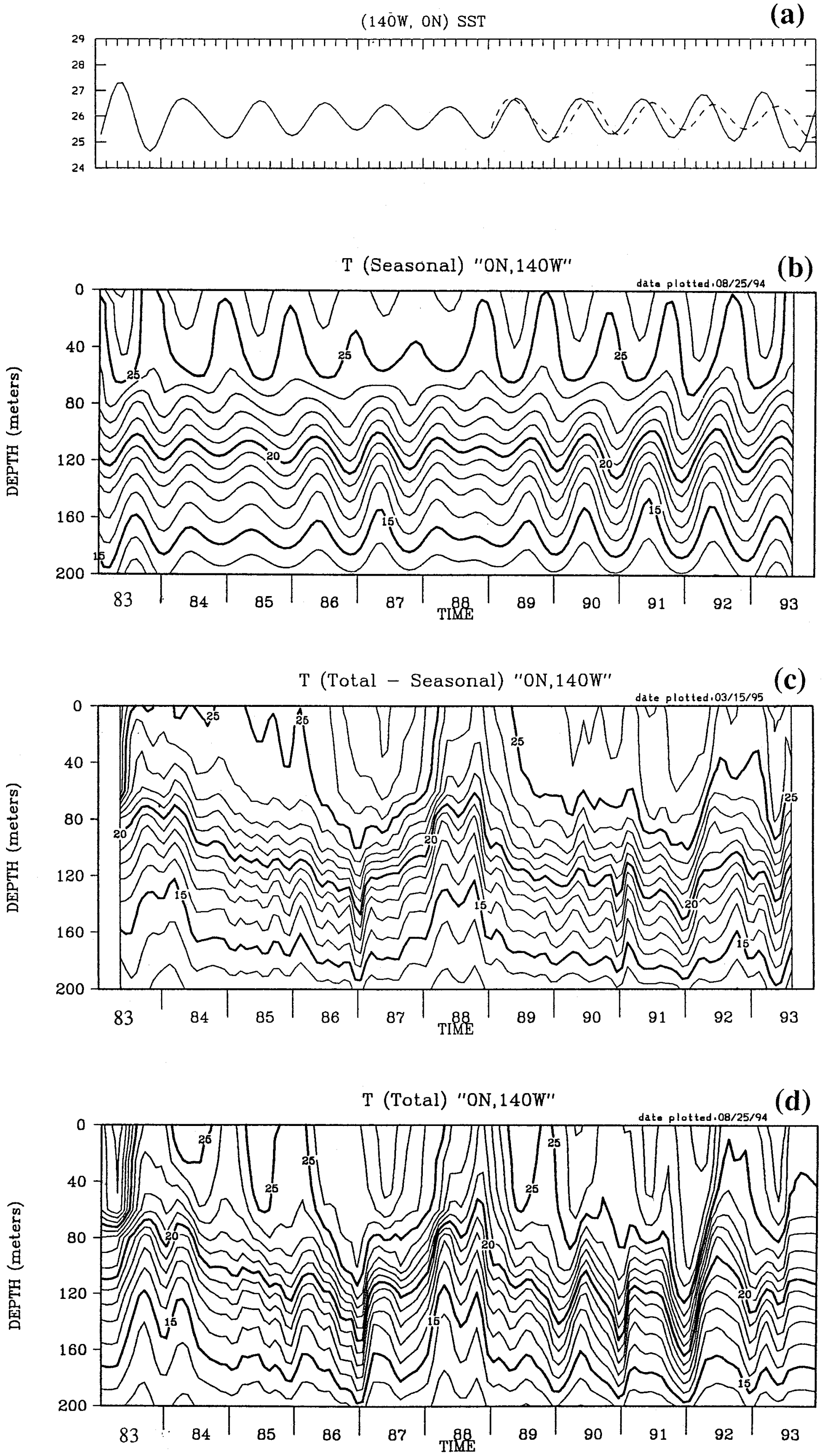

FIG. 7. Same as Fig. 7 but for 0°, 140°W.

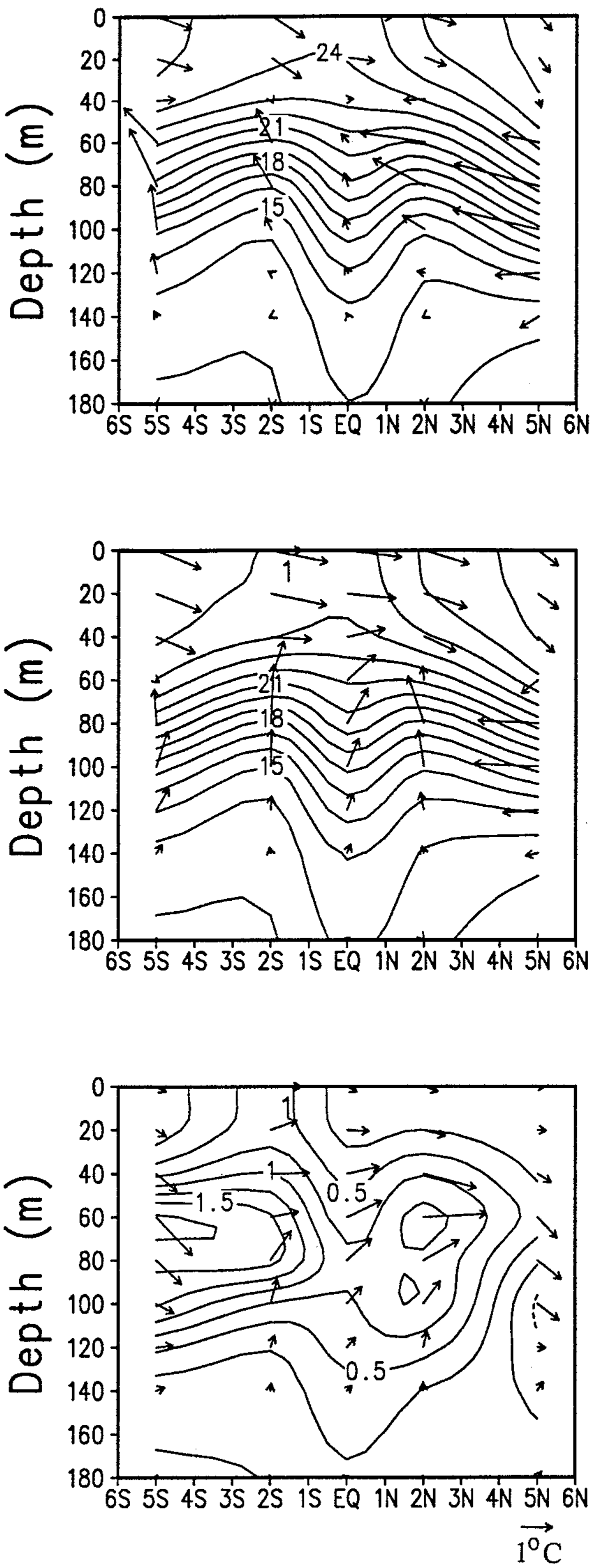

FIG. 8. Upper panel: time average (contours) and the annual harmonics of temperature (arrows) for the years 1985–87. Middle panel: same as upper panel but for the years 1990–92. Lower panel: the difference in the time average and annual harmonics of temperature between the two periods (1990–92 minus 1985–87).

110°W and 140°W over a 10-yr interval. Such documentation will be beneficial to developers of coupled ocean–atmosphere GCMs that attempt to simulate the tropical climate (e.g., Mechoso et al. 1995). One important finding of this paper is that the seasonal excursion of the thermocline has little effect on the seasonal cycles of SST. The most remarkable result, however, is the large increase in the amplitude of the seasonal cycle at the depth of the thermocline in the eastern equatorial Pacific after 1989. This subsurface change in the east was associated with changes in the surface winds: the seasonal cycle of the zonal wind in the central and western tropical Pacific propagated westward from 1985 to 1988, but eastward from 1989 to 1992.

The change in the winds over the central and western tropical Pacific in 1989 can, according to the model of Kessler and McPhaden (1995), force near-resonant oceanic Kelvin waves that are partially responsible for the observed amplitude change of the seasonal excursion of the thermocline. Another factor that could be important to this phenomenon is the gradual deepening of the thermocline during the 1980s (see Figs. 6 and 8), which coincided with a decrease in the westward extension of the seasonal cycle of SST.

If the change in the zonal winds over the central and western Pacific caused the modulation of the seasonal cycle in the thermocline, then what caused the change in the winds? Along the equator in the Pacific two sets of processes control seasonal variations at the ocean surface. In the east, a westward propagating ocean–atmosphere mode is dominant (Gu and Philander 1995; Xie 1994) but in the far west monsoons, possibly under the influence of the Indian Ocean, come into play. When the equatorial thermocline in the eastern Pacific is relatively shallow, as was the case before 1989, the annual ocean–atmosphere mode extends farther to the west. A gradual deepening of the thermocline in the eastern Pacific, as occurred during the 1980s, may cause an eastward contraction of this mode and an eastward expansion of monsoonal effects. This expansion may then modify the phasing of the forcing so that oceanic Kelvin waves cause a larger seasonal excursion of thermocline displacement in the east. This explanation—that is, that the eastern tropical Pacific thermocline depth anomalies are a key factor—is qualitatively consistent with the limited data described here, but the idea needs to be checked by analyses of more comprehensive datasets and models.

The changes in atmospheric and oceanic conditions described above could be related to another phenomenon that recently has attracted attention: the unexpected persistence of warm surface waters over the eastern equatorial Pacific during the early 1990s. This warming can be viewed as part of a decadal variability that involves changes in the depth of the tropical thermocline because of exchanges between the Tropics and extratropics (Gu and Philander 1997). Indeed, the associated changes in the surface winds were similar to those de-

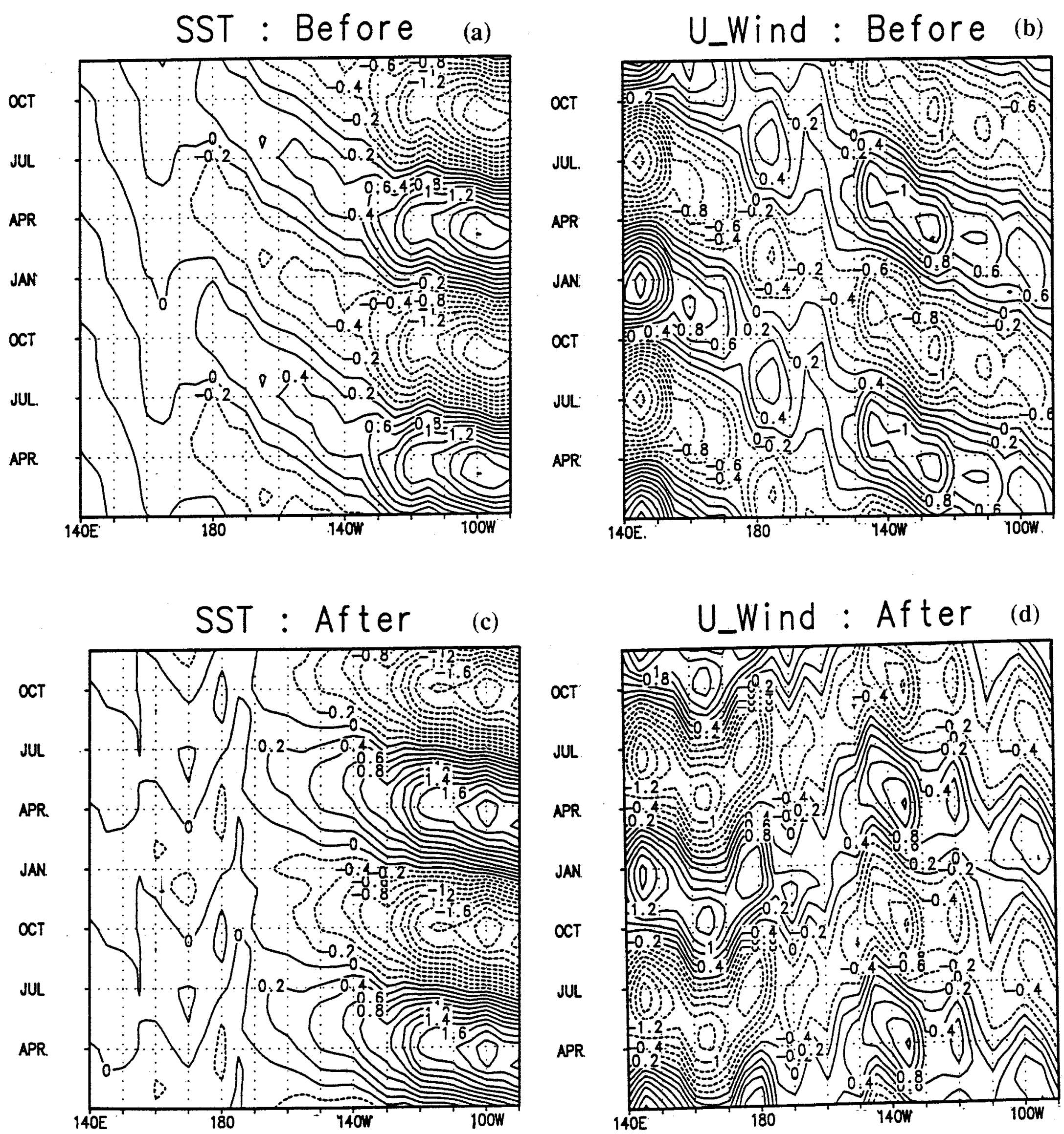

FIG. 9. Two-year repeated plots of mean seasonal cycle of SST and zonal wind along the equator from COADS. (a) and (b) 1985–88 (labeled as *Before*). (c) and (d) 1989–92 (labeled as *After*). The time direction is from bottom to top.

scribed in the previous paragraph, including the appearance of westerly wind anomalies over the western and central tropical Pacific (Latif et al. 1997). The connections, and possible feedbacks between the various facets of the decadal fluctuations mentioned here, including changes in the seasonal cycle of the thermocline depth, are important topics that remain to be explored.

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