The Annual Cycle of Sea Level in the Eastern Tropical Pacific

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(Manuscript received 29 July 1985, in final form 17 December 1985)

ABSTRACT

The long period response of sea level on the eastern boundary of the Pacific separates, for linear dynamics, into a remotely forced part mainly due to zonal winds along the equator to the west, and a locally driven part where sea level slopes to balance the alongshore wind. Examination of the annual component of sea level on the eastern boundary indicates that the locally forced part dominates, whereas the remotely forced part plays a major role at semiannual and interannual periods.

1. Introduction

Annual changes of sea level in the tropical Pacific are generally quite small, with amplitudes of the order of 5 cm in most places. At the eastern boundary however these amplitudes show a distinct increase away from the equator, reaching 15 cm at Balboa, Panama. Another notable feature of the eastern boundary variations is the 180° phase difference between the two hemispheres. This phase difference has been ascribed to heating effects (Wyrtki and Leslie, 1980), as it is prominent away from the equator everywhere in the Pacific. However, the phase change on the eastern boundary is particularly distinctive and occurs within a degree or so of the equator. Taken together with the relatively large amplitude on the boundary, other mechanisms for explaining the sea level cycle seem worth investigating.

The prominent annual signal on the eastern boundary has been related to a seasonal Kelvin wave generated further to the west (Kindle, 1979). Busalacchi and O'Brien (1980) used a reduced gravity model to study the seasonal cycle and found a strong semiannual oscillation at the eastern boundary which correlated well with observed changes in the depth of the 14°C isotherm. They did not, however, compare their results with sea level changes. Also, the remotely forced variations, being generated by Kelvin waves, would be expected to be symmetric about the equator, and so not exhibit the phase changes referred to above.

In this paper it will be shown that the long-period sea level response on the eastern boundary of the Pacific ocean can be divided into a remotely forced part and a locally driven part. The latter is due to the longshore component of the wind which is largely meridional. This meridional wind component is often ignored in models although it has been analysed in other contexts (see Enfield, 1981a, and his references). Northeastern Pacific longshore wind stress has been correlated with sea levels (Enfield and Allan, 1980) but the tropical area has not received such treatment. In the equatorial zone the locally driven part appears to be an important contributor to the annual period changes in the sea level on the eastern boundary and helps to explain the larger amplitude and the phase change described above.

A simple theory demonstrating the effect of purely meridional wind stress forcing on sea level will first be given and then this will be applied to the observed wind field. The fundamental structure of the observed annual cycle is reproduced well, both in phase and amplitude, including such peculiarities as the signal at Balboa. Differences between the theory and observations are related to the semiannual signal in the pycnocline, which responds to remote forcing (Busalacchi and O'Brien, 1980). It seems likely, therefore, that the meridional wind stress balance contributes the major part of the annual cycle. Remotely forced effects are, however, important for interannual changes. This can be seen from comparing the annual mean response with the response seen in particular years, including the strong 1982/83 event.

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2. Theory of annual fluctuations on the eastern boundary

The eastern boundary plays a unique role in the behavior of annual period waves because of the special nature of equatorially trapped long waves. For any given vertical mode, there is only one wave, the Kelvin wave, which can bring information to the boundary from the west. The remaining infinite set of planetary waves can only carry information away from the eastern boundary. Thus the eastern boundary response can be subdivided into two independent constituents: (i) the remotely forced part, i.e., that which is the result of Kelvin waves incident on the boundary, and (ii) the locally forced part.

The remotely forced component has received some attention in the past, but the locally forced response has not been the subject of comparable interest. This part is driven by the alongshore wind stress and satisfies a very simple equation. That is because coastal Kelvin waves carry information out of the equatorial zone in a very short time (a week or two) compared with the forcing period. Hence the alongshore momentum balance is effectively an equilibrium one. Treating the boundary as being purely meridional, with y as distance northwards of the equator and u as the normal component of velocity, the equilibrium balance is

$$\beta y u = -\frac{1}{\rho_0}\frac{\partial p}{\partial y} + \frac{1}{\rho_0}\frac{\partial Y}{\partial z} \tag{2.1}$$

where β is the gradient of the Coriolis parameter, p is the pressure perturbation, ρ_0 the density of the ocean, Y the horizontal longshore stress and z the vertical coordinate. If Y is assumed to vary linearly from its surface value Y_s to zero over an upper mixed layer of depth $H_{\rm mix}$ (e.g., see Gill, 1982), then the balance on the boundary (where $u = 0$) at the surface (where $p = \rho_0 g\eta$) is

$$g\frac{\partial \eta}{\partial y} = \frac{Y_s}{H_{\rm mix}} \tag{2.2}$$

which gives the surface elevation apart from an arbitrary constant. It is the determination of this constant that provides the special interest in this problem.

To see how the constant is determined requires a look at the theory of equatorially trapped modes which is given in chapter 11 of Gill (1982). First, the terms in

$$\beta y u = -g\frac{\partial \eta}{\partial y} + \frac{Y_s}{\rho_0 H_{\rm mix}} \tag{2.3a}$$

$$-\beta y v = -g\frac{\partial \eta}{\partial x} + \frac{X_s}{\rho_0 H_{\rm mix}} \tag{2.3b}$$

where X_s is the surface zonal wind stress, are expanded in terms of vertical modes (Gill, 1984, section 2). Second, the dependent variable projections are expanded in terms of parabolic cylinder functions, for instance

$$g\eta^{(m)},\ Y_s/\rho_0 H_{\rm mix} = \sum_{n=0}^{\infty} (\eta_n{}^{(m)},\ Y_n) D_n[(2\beta/c_m)^{1/2} y], \tag{2.4}$$

where $\eta^{(m)}$ is the projection of p/ρ_0 onto the mth vertical mode, c_m is the wave speed for this mode and D_n is a parabolic cylinder function of order n. Equation (2.3) then gives equations for the variables

$$q_n{}^{(m)} = \frac{g}{c_m}\eta_n{}^{(m)} + u_n{}^{(m)}$$

$$r_n{}^{(m)} = \frac{g}{c_m}\eta_n{}^{(m)} - u_n{}^{(m)}. \tag{2.5}$$

These variables represent the amplitudes of equatorially trapped waves (Gill, 1982). In particular, $q_0{}^{(m)}$ represents the Kelvin wave (Gill, 1982). The equation for $q_0{}^{(m)}$,

$$\frac{\partial q_0{}^{(m)}}{\partial t} + c_m\frac{\partial q_0{}^{(m)}}{\partial x} = \sigma_0 X_0 \tag{2.6}$$

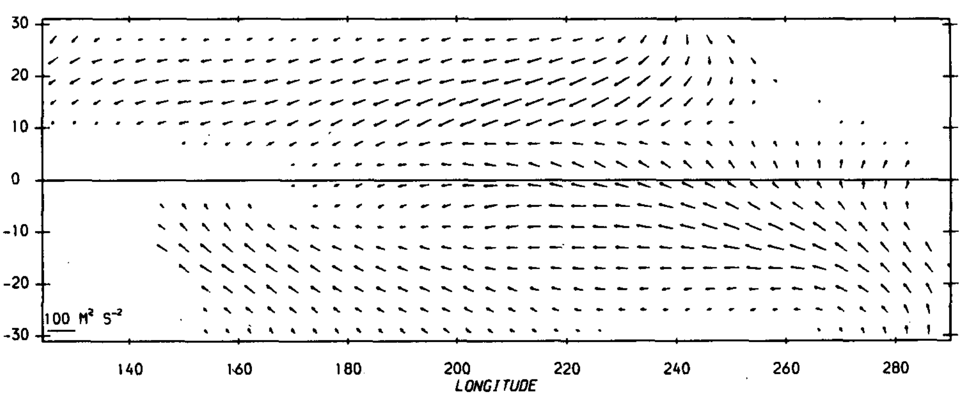

FIG. 1. Average Pacific pseudo wind stress for June 1979–May 1982. Very small stresses not shown.

where σ_m is the projection of a step function (unity in the mixed layer, zero below) onto mode m, is not forced by the meridional wind. Therefore it does not contribute at all to the solution of the mth vertical mode equation corresponding to (2.2)

$$\frac{g\partial\eta^{(m)}}{\partial y} = \sigma_m Y_s/\rho_0 H_{\text{mix}}. \tag{2.7}$$

At the eastern boundary, $\eta^{(m)}$ is directly proportional to $q^{(m)}$

$$q^{(m)} = \left(\frac{g}{H_m}\right)^{1/2} \eta^{(m)} \tag{2.8}$$

where $H_m = c_m^2/g$ is the equivalent depth for mode m. Therefore $\eta^{(m)}$ cannot contain a contribution from the zero order cylinder function. This is the condition which determines the arbitrary constant, namely that the projection of $\eta^{(m)}$ onto the zero order cylinder function belongs to the remotely forced contribution rather than the locally forced part.

Hence a precise procedure for calculating η is established. Equation (2.7) is integrated for each vertical mode subject to the constraint

$$\int_{-\infty}^{\infty} \eta^{(m)} D_0[(2\beta/c_m)^{1/2} y]dy = 0 \tag{2.9}$$

and η, the surface value, is obtained by adding modes. This determines the arbitrary constant involved in integrating (2.2).

An alternative way of seeing the arbitrariness which comes when trying to integrate (2.7) is to expand both $\eta^{(m)}$ and Y_s in cylinder functions, and substitute directly into (2.7). Equating coefficients gives

$$(2\beta/c_m)^{1/2}\eta_1{}^{(m)} = \sigma_m Y_0$$
$$(2\beta/c_m)^{1/2}(-\eta_0{}^{(m)} + \eta_2{}^{(m)}) = \sigma_m Y_1 \tag{2.10}$$

etc. The arbitrariness comes in the second equation. The appropriate condition is to choose $\eta_0{}^{(m)}$ as zero, so the second equation gives a value for $\eta_2{}^{(m)}$. Hence all the coefficients $\eta_n{}^{(m)}$ are assigned unique values. It can be shown that if $\eta_0{}^{(m)}$ were given a nonzero value, it would result in a constant being added to $\eta^{(m)}$.

Before examining the sea level structure of the east Pacific some immediate consequences of the above theory are worth stating. From the form of (2.7) it can be seen that an equatorially symmetric meridional wind stress gives rise to an antisymmetric sea level and vice versa. Also, because the equation is linear a wind stress can be examined in terms of its symmetric and antisymmetric parts. The form of the constraint determining the constant of integration, (2.10), dictates that the symmetric part of the stress has no such constant associated with it while the antisymmetric part does.

The condition (2.9) has a simple interpretation for it states that the weighted average of η in a small equatorial strip must be zero. For the first mode, the relative weights at 1, 3, 5 and 7 degrees of latitude are 1.00, 0.64, 0.26 and 0.08 respectively, which in practice means that the zero of η is within a degree or so of the equator. For higher modes the zero is even closer to the equator, so there is little error in neglecting the contribution of these modes.

The results of the following section were obtained by integrating (2.7) along 81°W, between 19°S and 9°N of latitude and using only the first vertical mode cylinder functions. Using tests on simple antisymmetric wind stresses as a guide the error so incurred was estimated to be less than 1 cm, and so not significant. The mixed layer depth chosen for the calculations was taken from Levitus (1982). As can be seen from (2.7) the effect of varying this parameter is inversely related to the sea level. The vertical mode speeds are also de-

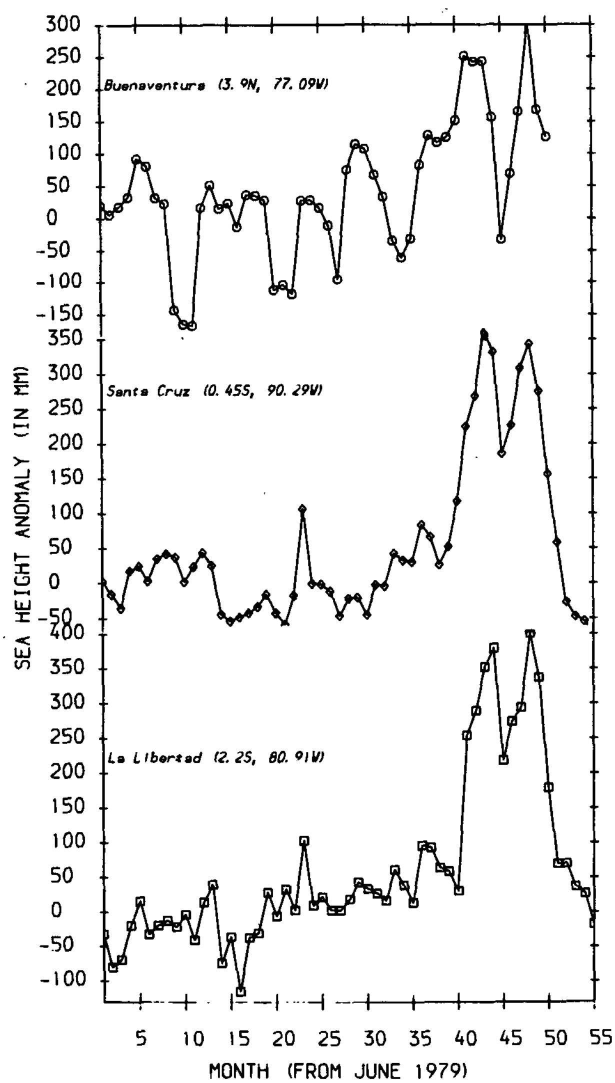

FIG. 2. Sea level anomalies in the eastern equatorial Pacific during the period June 1979–December 1983.

pendent on H_{mix} but by far the more important influence on the sea level is this inverse relationship.

3. The annual sea level cycle in the eastern Pacific

The average wind stress in the tropical Pacific for the period June 1979–May 1982 is shown in Fig. 1. The large longshore component on the eastern boundary is evident, as is the change in sign and the large gradient of this component associated with the Intertropical Convergence Zone. There is also considerable temporal variability in this field associated, among other things, with seasonal and interannual changes in the position of the ITCZ (Wyrtki and Meyers, 1975). Figure 2 shows sea level changes at three near equatorial stations for the period June 1979 to March 1983. The 1982/83 El Niño event distorts the last section of the record, but prior to that annual signals at each station can be discerned, and these vary considerably between stations. This is a signature of locally forced changes. Remotely forced changes would be expected to be similar at all three stations.

The contribution of meridional wind stress to the eastern equatorial Pacific annual sea level signal will now be examined in two ways. First, long term monthly means will be compared with the result of the simple theory of section 2 and, second, particular years will be scrutinized. In carrying out these two tests both the

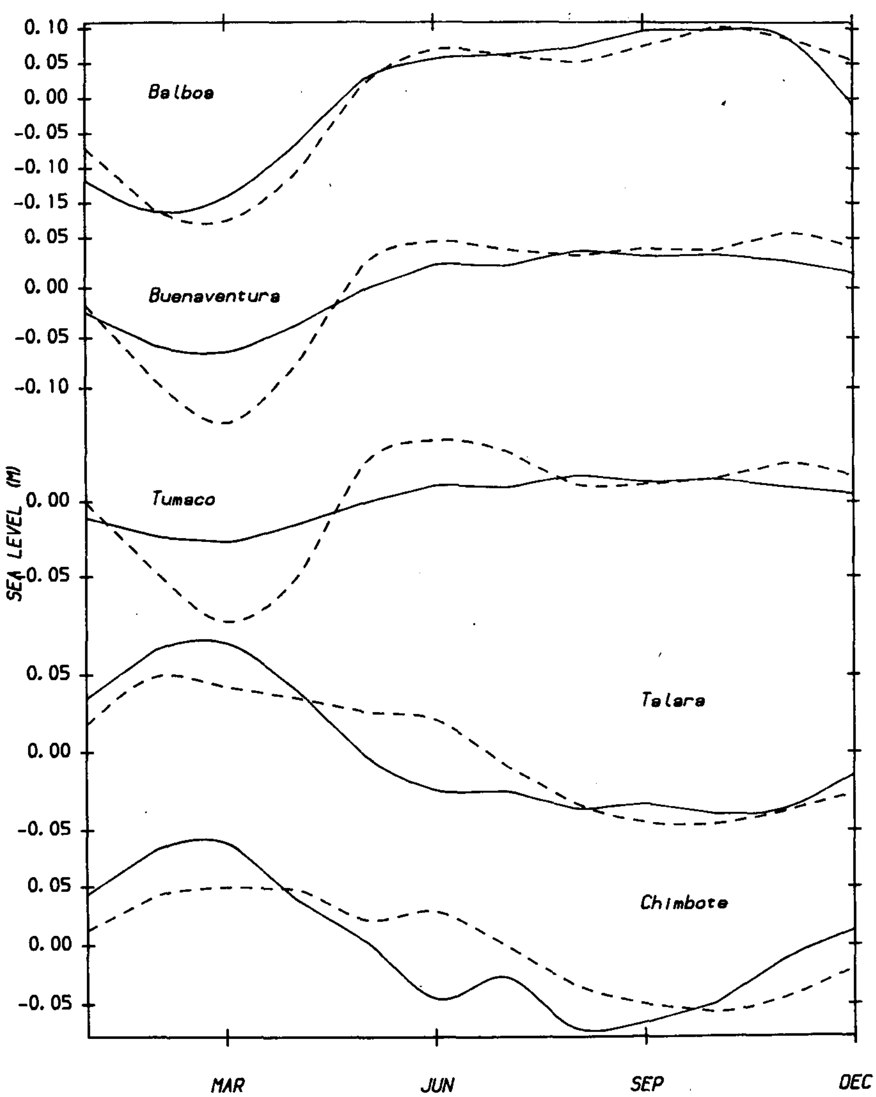

FIG. 3. Monthly average sea level in the eastern Pacific, both from theory (solid) and observations (dashed). H_{mix} = 7.5 m, c = 2.55 m s^{-1}. Stations along the South American coast at Balboa (8°58′N, 79°36′W). Buenaventura (3°54′N, 77°06′W), Tumaco (1°50′N, 78°44′W), Talara (4°37′S, 81°17′W) and Chimbote (9°05′S, 78°38′W).

climatological effect of the meridional wind stress in the annual cycle and a measure of the interannual variability will be studied. Also, some conclusions about remote forcing emerge from differences between our theory and observations. The eastern Pacific between the Galapagos Islands and Ecuador will be considered to have a mixed layer of 7–10 meters depth (Levitus, 1982) and a first-mode Kelvin wave speed of 2.55 m s^{-1}. The wind stress field used for the calculations is that of Florida State University. Details of its production are given in O'Brien and Goldenberg (1982) and Inoue and O'Brien (1984). To compute the actual stress values from the tabulated psuedostress a density of 1.2 kg m^{-3} and a drag coefficient of 1.7×10^{-3} (Inoue and O'Brien, 1984) were used. The necessary integrals were calculated between 19°S and 9°N by which stage the Gaussian function in the integrals is small. Sea level data were taken from Wyrtki and Leslie (1980), the Service for Mean Sea Level (1977) and data kindly supplied by K. Wyrtki.

In Fig. 3 the observed monthly average sea levels for five stations are compared with those given by theory. These stations, along the South American coast, show good agreement between theory and observations in both phase and amplitude. Even such an unusual signal as that at Balboa is well described by meridional wind stress arguments alone. From examining the monthly mean wind stress maps in Wyrtki and Meyers (1975) it is seen now that the large dip in the sea level north of the equator in the northern spring is due to the rapid southward retreat of the ITCZ. The accompanying relaxation of the winds south of the equator allows a higher sea level in this region, which is eroded as the wind front moves up the coast again in late spring, piling up water north of the equator. This can be illustrated by the theory's prediction of meridional sea level sections for March and September, shown in Fig. 4. It is difficult to compare these directly with data because of the uncertainty of sea level datum at different stations. However, examination of the Service for Mean Sea Level (1977) records reveals similar trends.

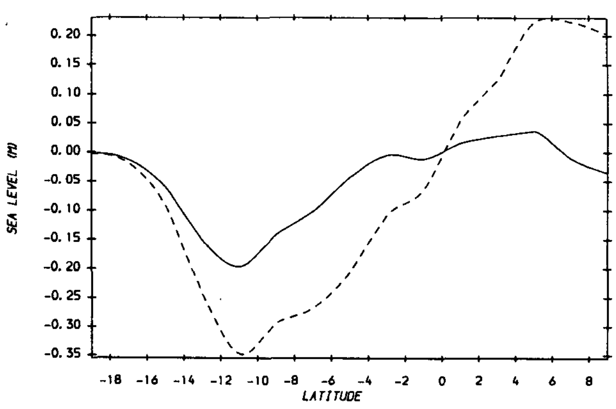

FIG. 4. Meridional sections of predicted sea level along the South American coast for March (solid) and September (dashed).

There is a tendency in Fig. 3 for the observed sea level to be lower in the spring than predicted, higher in the summer and lower once more in the fall. Comparison with the pycnocline anomaly diagram of Busalacchi and O'Brien (1980, Fig. 7), which shows a definite semiannual signal, suggests that the reason for this may be found in the influence on the sea level of disturbances propagating from the west. However, this remote forcing would appear to be only a secondary component of the annual cycle.

It should be noted that for this comparison the periods over which the sea level and wind fields were averaged are different. For the wind field the means were calculated over 1961–83, while the sea level averages were for various periods, dependent on the lifetime of the station, although in all cases the averaging periods included most of the 1960s. However, as the results of data studies by Barnett (1977), Goldenberg and O'Brien (1981), and Enfield (1981a,b), show relatively little interannual variability of winds in this region the different averaging periods should not cause much aliasing.

Having seen how well the average fields agree we now turn to an examination of particular years. In Fig. 5 the observed and predicted sea levels at the 5 stations considered in Fig. 4 are shown for the years 1963–64. It is immediately clear that the comparison is considerably worse. In any individual year other events are occurring to modify the average cycle and these may be of larger magnitude than the mean. However for these years the basic shape of the observed variation is reproduced in the theoretical predictions. Obviously there is a lot more going on but magnitudes and phases are roughly right. The divergence between model and observations follows a similar, but amplified, pattern to those in Fig. 3. From our earlier discussion this implies that in these years there were stronger than usual signals being fed into the area from the west. This particular period has been noted in the literature as one of considerable activity. In particular, 1963 saw unusually heavy rainfall in the central Pacific (Quinn, 1974) and is classified as a minor El Niño (Enfield and Allen, 1980). However, it has been seen that the meridional winds along the South American coast do not have a large interannual variability and indeed Wyrtki (1975) showed that even during El Niño disturbances the Peruvian coastal winds do not collapse, as was once thought.

To further emphasize the local forcing of the average annual sea level cycle, as opposed to the remote forcing of particular events, the contribution of the meridional wind balance to sea level during 1982 and the first quarter of 1983 is shown in Fig. 6. This period was chosen to correspond to that studied by Busalacchi and

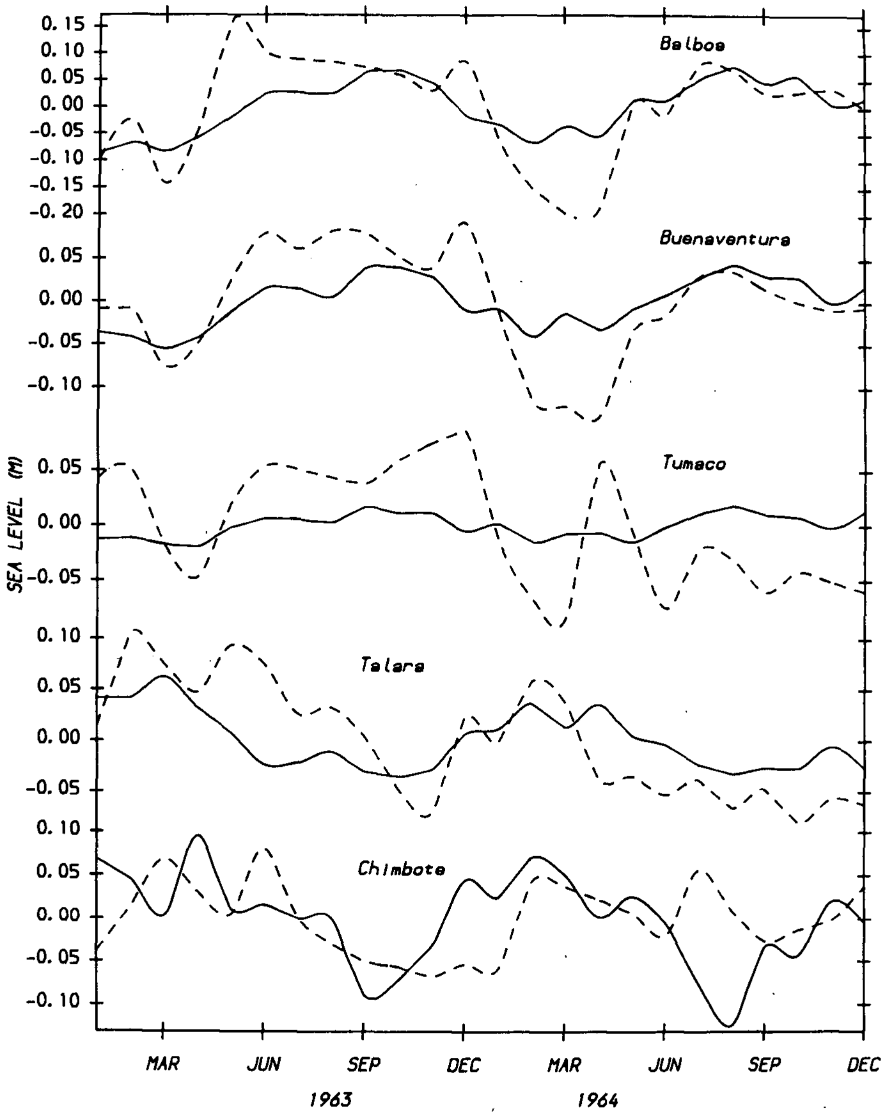

FIG. 5. Sea level anomalies for the years 1963–64 at the stations used in Fig. 3. For the legend see Fig. 3.

Cane (1985), using the model of Cane and Patton (1984). They give plots of sea level at a number of places in the equatorial Pacific, both observed and as calculated by their zonal wind-driven, quadruple vertical mode model. La Libertad, in mainland Ecuador, is in our area of interest. As can be seen the meridional wind contribution to the sea level is completely swamped by the effect of zonal remote forcing, well modeled by Busalacchi and Cane (1985).

4. Conclusions

Using only monthly mean meridional wind stress much of the average annual signal in the eastern tropical Pacific sea level can be explained. This involves a simple theory balancing the sea level and local wind stress forcing, without needing to consider the effect of remote forcing. The significant features, namely the phase and amplitude, both north and south of the equator, are well described. Differences between theory and observation correlate with the effects of remote forcing from the west.

For individual years, however, and especially those associated with El Niño disturbances, the predictions of the local theory can be quite poor, whereas calculations based on remote forcing are good. Thus interannual variability in sea level is dominated by remote rather than local forcing.

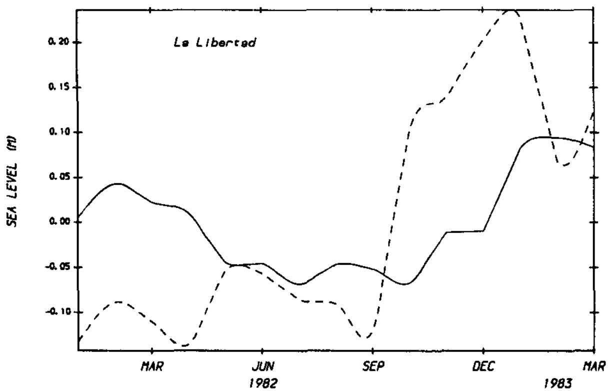

FIG. 6. Sea level anomalies at La Libertad (2°12′S, 80°55′W) for the period January 1982–March 1983. For the legend see Fig. 3. Cf. Busalacchi and Cane (1985, Fig. 4).

Acknowledgments. The authors would like to thank J. J. O'Brien for providing the wind stress data and K. Wyrtki for the sea level data from the Pacific sea level network.

This work is dedicated in memory of Adrian Gill, to whose careful thought the quality of the final result is due.

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