Further Evidence for Coastal Trapped Waves along the Peru Coast

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ABSTRACT

Time series of coastal sea level during 1976–77, from 2°12’S to 17°S along the west coast of South America, show that low-frequency, \( \omega < 0.25 \) cycles per day (cpd), fluctuations propagate poleward with the phase speed of baroclinic Kelvin waves (2–3 m s\(^{-1}\)). The alongshore coherence is highest in the frequency band 0.1–0.2 cpd. Computing the frequency-domain empirical orthogonal functions (EOF) for alongshore current, from an array of current meters extending from 5°S to 15°S during March–May 1977, gives 70% of the variance in the 0.1–0.2 cpd frequency band to an EOF mode with poleward phase propagation at 2.75 m s\(^{-1}\). The vertical structure of the alongshore current fluctuations (0.1–0.2 cpd) over the continental slope at 5°S and 15°S is consistent with a first-mode baroclinic Kelvin wave. The current and sea-level fluctuations are coherent and propagate poleward through latitudes where their frequency equals the local inertial frequency. The fluctuations are not significantly coherent with coastal winds from 4°S to 15°S and are therefore presumed to have an equatorial origin. Intermittent sea-level data at the Galapagos Islands during the period provide tenuous evidence that these fluctuations, propagating poleward as coastally trapped waves, previously traveled in the equatorial wave guide.

1. Introduction

Evidence for the existence of coastally trapped waves along the central Peru coast has been previously presented by Smith (1978). He showed that measurements made during the CUEA (Coastal Upwelling Ecosystems Analysis) JOINT-2 experiment in May–July 1976 and March–May 1977 indicated poleward propagation of fluctuations in sea level and currents between 10°S and 15°S with phase speeds of about 200 km day\(^{-1}\). The structure of the current fluctuations and the dynamical balances suggested that the waves were baroclinic Kelvin waves (Smith, 1978; Brink, et al., 1978; Brink et al., 1980). Although the frequency range of the propagating fluctuations [0.1–0.2 cycles per day (cpd)] was similar to that of weather events, the fluctuations were not correlated with winds measured at 12°S or 15°30’S and an equatorial origin was hypothesized.

Additional, and in some cases considerably longer, time-series measurements have become available for the same time period as the CUEA experiment: meteorological and tide-gage data from the Ecuador and Peru coasts equatorward of 10°S, and current-meter data from the ESACAN (Estudio del Sistema de Afloreamiento Costero en el Area Norte) experiment conducted by the Institut für Meereskunde an der Universität Kiel and Instituto del Mar del Peru at 5°S off the Peru coast during March–May 1977. The ESACAN data set is of particular interest because it provides an opportunity to study the wave structure at lower latitudes and the possibility of linking equatorially generated fluctuations to those observed between 10° and 15°S during the CUEA experiment.

The combined data set includes current and temperature measurements near 5°, 10°, 12°, and 15°S off Peru, hydrographic sections normal to the coast at those latitudes, coastal winds near 4°, 7°, 12°, and 15°S, and sea level from the tide gages on the Galapagos Islands (0°27’S) and on the South American continent between 2° and 17°S. Atmospheric pressure data were obtained near many of the tide gages. An analysis of the data near 15°S is presented by Brink et al. (1980), and a discussion of the ESACAN data near 5°S is given by Fahrbach et al. (1981). The statistics of the CUEA and ESACAN current meter data are given by Brockmann, et al. (1980) in a study on the Peru upwelling.

In this study we utilize the CUEA and ESACAN current meter and coastal wind data, and the longer tide gage records, to reexamine the conclusions presented by Smith (1978) and to extend his analyses equatorward. The locations of the observations used are shown in Fig. 1. Unless explicitly stated otherwise, the time series presented and analyzed in this paper have been initially subjected to a low-pass filter (half-power point = 1.96 days) to eliminate diurnal and higher frequency phenomena. The term sea level in this paper refers to “adjusted” sea level: The atmo-
spheric pressure records were used to adjust sea level to provide an equivalent to a subsurface pressure, i.e., the "inverted barometer" effect was removed by adding the atmospheric pressure (in mb) to sea level (in cm). A more detailed discussion of the data set is given in the Appendix.

2. The longer time series

The longest reliable data records are those of sea level and atmospheric pressure at La Libertad (2°12'S), Callao (12°03'S) and San Juan (15°21'S), and those of current and wind near San Juan. The common period for continuous data extends for nearly 14 months from late March 1976. The adjusted sea levels, the current measured at 55 m over the shelf (120 m) at 15°S and the coastal wind at San Juan are shown in Fig. 2. The dominant fluctuations in sea level are clearly similar at the three tide gage stations, and the tendency for the extrema at a southern station to lag the equivalent extrema at a northern station may be seen in Fig. 2 (cf. Fig. 2 of Smith, 1978). The current fluctuations are obviously related to those of sea level, e.g., equatorward (northwestward along the coast) flow occurs in conjunction with sea level minima. The coastal wind is remarkably uniform in direction and is apparently not the cause of the fluctuations in the coastal current.

Figure 3 shows the autospectra of Callao and San Juan sea levels and the coherence and phase between them for the long records. The high coherence and the linear dependence of phase on frequency, for frequencies less than 0.2 cpd, indicates non-dispersive poleward propagation at about 240 km day$^{-1}$ (2.75 m s$^{-1}$), a speed consistent with a first baroclinic mode Kelvin wave (Moore and Philander, 1977, p. 326). The phase speed of 240 km day$^{-1}$ was estimated from the phase difference at the frequency of maximum coherence ($\omega = 0.165$ cpd), and appears to be a good estimate for $\omega < 0.25$ cpd. The "drop out" in coherence near 0.14 cpd is puzzling but seems real because
Auto- and cross-spectra are shown in Fig. 4 for La Libertad and San Juan sea level. The distance between these stations (1650 km) is much greater than between Callao and San Juan (450 km), and the coherence is generally lower. However, there is high coherence in the 0.10–0.19 cpd band. A linear dependence of phase on frequency for frequencies less than 0.3 cpd is suggested, with phase consistent with poleward propagation at a speed slightly greater than 240 km day$^{-1}$.

Although the cross-spectra suggest that the non-dispersive wave-like propagation extends to lowest frequency, there is a theoretical lower bound on fre-

**FIG. 3.** Autospectra (spectral density: cm$^2$/cpd) of sea level from the Callao (solid line) and the San Juan (dashed line) tide gages, based on 405 days of data beginning 27 March 1976, and the coherence and phase (positive for Callao leading) between Callao and San Juan sea level. Smoothing was done with a moving spectral window. The smoothed estimators have 20 degrees of freedom; the band-width is 0.025 cpd. Phases associated with coh$^2 < 0.04$ are not shown for $\omega < 0.3$ cpd in order to reduce visual clutter. The 99% significance level is indicated on the coherence plot; the 95% confidence interval for phase is $\pm 30^\circ$ for squared coherence $> 0.3$. The line on the phase-versus-frequency diagram represents 240 km day$^{-1}$ phase speed for poleward propagating nondispersive waves.

it appears in the cross-spectrum involving any portion of the Callao record sufficiently long to resolve $\pm 0.02$ cpd. A hint of the drop out is seen in the cross-spectrum of San Juan sea level and current (Fig. 6 herein) and in the cross-spectrum of 72-day records of alongshore current at 12$^\circ$ and 15$^\circ$S from 1976 (Fig. 6 of Smith, 1978).

**FIG. 4.** Autospectra of sea level from the La Libertad (solid line) and the San Juan (dashed line) tide gages, and the coherence and phase (positive for La Libertad leading) between La Libertad and San Juan sea level, as in Fig. 3 except the 95% significance level is indicated on the coherence plot.
quency for coastal trapping (Allen and Romea, 1980). For co
tally trapped baroclinic waves, $\omega > \beta \delta R/2$, 
where $\beta$ is the variation of the Coriolis parameter 
($f$) with latitude and $\delta R$ is the internal Rossby radius 
of deformation. For example, this condition is satis
fied at $12^\circ S$ for $\omega > 8 \times 10^{-7}$ s$^{-1}$ ($1.1 \times 10^{-2}$ cpd) 
with alongshore phase speed $C = \delta Rf = 240$ km day$^{-1}$. 
However, the theoretical lower bound on frequency 
for coastal trapping at the latitude of La Libertad 
($2^\circ 12^\prime S$) is $8 \times 10^{-2}$ cpd. Energy at frequencies less 
than that at La Libertad would not be trapped in the 
coastal waveguide, and thus would not be efficiently 
propagated. This may explain why the coherence at 
$\omega < 0.1$ cpd is so much lower between La Libertad and 
San Juan ($15^\circ S$) than between Callao ($12^\circ S$) and 
San Juan, especially in comparison with the coherence 
in the band $0.1$ cpd $< \omega < 0.2$ cpd.

In contrast to the sea level, the atmospheric pres
sures at La Libertad, Callao, and San Juan are in 
phase and highly coherent for frequencies $\omega < 0.2$ 
cpd. Figure 5 shows the spectral computations for 
atmospheric pressures at La Libertad and San Juan. 
The figure indicates that the wave-like propagation 
observed in sea level is not a forced response to large 
scale atmospheric pressure systems.

If the fluctuations represent baroclinic co
tally trapped waves with the alongshore velocity in geo
strophic balance, sea level and currents should be 
highly coherent. In the upper layer the phase relation 
should be $180^\circ$ for inviscid motion, i.e., southward 
flow near the coast should be associated with higher 
sea level near the coast. Figure 6 shows the cross-
spectra between sea level at San Juan and alongshore 
current at M ($55$ m). The alongshore direction has 
been defined as the principal axis direction from the 
current record. This is close to the orientation of the 
local isobaths (Smith, 1981). Alongshore current and 
sea level are highly coherent for $\omega < 0.2$ and nearly 
$180^\circ$ out of phase, consistent with a geostrophic balance. 
Friction may account for the lack of an exact 
$180^\circ$ phase relationship (Brink, 1982a), or it may simply 
result from the fact that the current meter location 
is $50$ km equatorward of the tide gage. This latter fact 
would account for the observed phase difference 
near $0.17$ cpd, assuming a propagation speed of $240$ 
km day$^{-1}$.

Finally, we examine the relationship between local 
winds at San Juan and the alongshore current at M 
($55$ m). The cross-spectrum (Fig. 7) shows relatively 
high coherence for $0.06$--$0.09$ cpd but low coherence 
for $0.1$--$0.2$ cpd. Although this suggests that some low-
frequency motions may be locally wind driven, it sup-
ports the hypothesis that the current fluctuations in 
the $0.1$--$0.2$ cpd band, which are observed as propaga-
ting disturbances in the long sea level records, are 
not locally wind driven.

The analysis of the long records is consistent with 
the findings of Smith (1978), based on considerably 
shorter records. The long records have provided 
greater confidence and resolution in the cross-spectral 
computations. The frequency band around $0.17$ cpd 
is especially suggestive of free propagating coastally 
trapped waves since all cross-spectra of sea level or 
alongshore current (cf., Smith, 1978) have coherence 
maxima near that frequency with a phase relationship 
consistent with poleward propagation at $240 \pm 40$ km 
day$^{-1}$. In the analysis of the shorter record of current-
meter data from March--May 1977, we shall focus on 
the frequency band centered near $0.17$ cpd. The 
shortness of the common period for the data from 
the alongshore array of current meters precludes 
studying the very low-frequency variability.

**Fig. 5.** Autospectra (mb$^2$/cpd) of La Libertad (solid line) and 
San Juan (dashed line) atmospheric pressure, and the coherence 
and phase between them, as in Fig. 3. The 99% significance level 
is indicated on the coherence plot; phase positive for La Libertad 
leading.
Are the empirical modes at 5°S and 15°S (the latitudinal extrema of the current measurements) similar? The latter is of interest because of the theoretical finding of Allen and Romea (1980) that coastally trapped baroclinic fluctuations, propagating poleward from the equatorial region, may take the form of barotropic shelf waves at mid-latitudes.

a. Sea level

The additional sea-level data series available during March–May 1977 are not especially useful for spectral analyses because the various tide gages malfunctioned at different times during the period, leaving only a short common period with reliable data. How-

3. Observations during March–May 1977

The most extensive and complete data set exists for the period of March–May 1977. Data from all sites shown in Fig. 1 were obtained during that period. Brink et al. (1980), using the data near 15°S, showed that the local subtidal momentum balances \( \langle u \rangle, p_x; v, T_x; v, p_y \) and the empirical modes were consistent with those expected of free internal Kelvin waves. Using the spatially extensive data set from the array of measurements indicated in Fig. 1, can we demonstrate that consistent poleward propagation of the fluctuations in current and sea level extends from the equatorial zone? Can we eliminate the coastal winds as the dominant driving force for the fluctuations?

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**Fig. 6.** Autospectra of alongshore current [solid line: \((cm^2 \cdot s^{-2})/(cpd)] at M (55 m) and sea level at San Juan (dashed line: \(10^{-1} \ cm^2/(cpd)\) and the coherence and phase between them, as in Fig. 3. The 99% significance level is indicated on the coherence plot; phase positive for current leading.

**Fig. 7.** Autospectra of alongshore wind [solid line scale on left: \((m^2 \cdot s^{-2})/(cpd)] at San Juan and alongshore current [dashed line scale on right: \((cm^2 \cdot s^{-2})/(cpd)] at M (55 m), and the coherence and phase between them, as in Fig. 3. The 95% significance level is indicated on the coherence plot; phase positive for wind leading.
ever, evidence for consistent poleward propagation in sea level perturbations can be summarized by the lagged cross-correlation of the data from pairs of stations. Since the cross-spectral analysis of the long sea level records suggests nondispersive propagation, the time lag at which the maximum correlation occurs between sea level stations should depend linearly on the alongshore separation of the stations.

For each pair of sea level records, we have used the longest common record length for the pair that extends into the March–May 1977 period to compute lagged cross-correlations. In Fig. 8, the time lag at which maximum correlation is obtained is plotted as a function of alongshore separation for each adjacent pair of sea level records and each record versus Callao sea level, except Talaro which had occasional monthly datum shifts of uncertain magnitude. We include San Martin versus Matarani because that pair has the same spatial separation, but displaced several degrees longitudinally, as the subsurface pressure gages at P and near M. The correlations at the lag indicated in Fig. 8 are the highest correlations obtained for the pair and are significant at the 99.5% level (see Appendix), i.e., the probability that the correlation coefficient would result from uncorrelated data is less than 0.005. At the equivalent leads (in contrast to the lags in Fig. 8) the correlations are not significant at even the 90% level for spatial separations greater than 500 km. The conclusion from the lagged correlations of this suite of data, which extends from nearly 2°S to 17°S, is the same as that from the cross-spectral analysis of the three longer sea level records: perturbations in sea level propagate poleward at about 240 km day⁻¹.

The time lag for maximum correlation between the atmospheric pressure at Callao and that at La Libertad, Talaro, San Juan, and Matarani is also shown in Fig. 8. A well defined maximum correlation (correlation coefficient greater than 0.9 with significance level greater than 99.9%) at lags less than or equal to 0.5 days exists. Moreover, the pressure data are not significantly correlated (95% confidence level) at lags commensurate with the observed propagation of fluctuations in sea level. This corroborates the findings from the cross-spectral analysis of atmospheric pressure that the pressure systems do not move in such a way as to force the observed propagation of sea level.

b. Current velocity

The combined ESCACN–CUEA current meter data set from March–May 1977 provides a useful alongshore array of current observations that elucidates the structure and behavior of the propagating fluctuations better than sea level. The current meter data used here have been rotated into a coordinate system in which the alongshore direction is defined by the principal axis of the vector time series.

The currents from 10°S (P) and 15°S (M and L) have been analyzed by Smith (1978) and have been shown to be coherent with those at 12°S (Y), with phase differences consistent with free propagating coastal trapped waves. From the additional records at 5°S, we use the C2 mooring (with current measurements from 86, 126, 197, 560, and 860 m) which is located over the slope about 50 km from the coast in 1360 m water depth. Two other moorings were deployed at 5°S, one on the shelf nearly within the sight of the Bay of Paiata and the other 25 km farther offshore from C2. The current meters from the shelf mooring are “contaminated” by motions in the Bay of Paiata (Fahrbach et al., 1981) and the current measurements at the mooring farthest offshore were too deep (the shallowest current meter was at 195 m) to reveal clearly the baroclinic waves.

In Figs. 9 and 10 we show the spectra and cross-spectral computations based on the hourly alongshore velocity data from the 86 m current meter at C2 and the 97 m current meter at Y. In the subtidal frequencies high coherence is found in the 0.00625–0.00804 cph (0.15–0.19 cpd) band with a maximum at 0.00714 cph (0.17 cpd). There is a shoulder or relative maximum in both the C2 and Y variance spectra in the frequency band with high coherence. This is also found in the Peyote (10°S) current and Callao sea level spectra for this period (cf. Fig. 8 of Smith, 1978). Figure 10 indicates the phase by which Y leads C2 or, equivalently, C2 lags Y. A phase plot is ambiguous in showing whether a signal actually leads or lags: Figure 10 may be interpreted as C2 lagging Y by 115° or C2 leading by 360° – 115° = 245°. The ambiguity is resolved by considering the phase between other pairs: C2–P, P–Y, P–M, Y–M and by considering the lagged correlation; C2 leading Y by 245° is the quantitatively consistent interpre-

![Fig. 8. The lag (equatorward series leading) at maximum correlation between: 1) pairs of sea-level records (circles) from coastal tide gages in Fig. 1; 2) sea bottom pressure gages (cross) near M (15°S) and P (10°S); 3) each atmospheric-pressure record versus Callao atmospheric pressure (triangles). See text.](image-url)
Fig. 9. Autospectra of alongshore currents at C2 (86 m) (solid line) and Y (97 m) (dotted line) based on 1120 hourly values beginning 2100 GMT 31 March 1977. The 95% confidence interval is shown.

The C2 and Y moorings are 900 km apart along the coast. The phase difference (at 0.17 cpd) gives an estimated alongshore propagation speed of 225 km day^{-1} (2.6 m s^{-1}) poleward.

A geostrophic balance for the alongshore velocity in the frequency band 0.1–0.2 cpd is consistent with the current measurements at 5°S and 12°S obtained during March–May 1977. The cross-spectrum between alongshore current at C2 (86 m) and sea level at Talara shows high coherence (greater than the 95% significance level) in the 0.15–0.19 cpd frequency band with a maximum (coh^2 = 0.72) at 0.17 cpd. Talara sea level lags by 154°, which differs from 180° in the sense expected since Talara is equatorward of C2. At 12°S, Callao sea level and the alongshore current at Y (97 m) are coherent above the 95% significance level in the band 0.10–0.19 cpd with a maximum coherence at 0.16 cpd (coh^2 = 0.79) and a phase difference of 180°. These spectral calculations, along with those at 15°S (Fig. 6), support our hypothesis that the sea level and current fluctuations are manifestations of the same phenomena, and that the fluctuations represent coastally trapped waves.

The most efficient way to examine the array of alongshore current meter data is to compute empirical orthogonal functions (EOFs), utilizing the method of EOF analysis in the frequency domain to obtain the alongshore amplitude and phase distribution of the variability (see Wallace and Dickinson, 1972a, for details of the method). This technique has been successfully used by Wang and Mooers (1977) for analyzing low-frequency sea level and atmospheric pressure fluctuations off the west coast of the United States. The analysis expresses time series in terms of a linear combination of the eigenvectors of the cross-spectrum matrix for the frequency interval of interest and has the advantage over individual cross-spectra between record pairs since it exploits the interrelationships between all records. The method is closely related to the use of EOFs in the time domain applied to band-pass filtered data (cf. Kundu et al., 1975). However by using the cross-spectral matrix we obtain the phase as well as the magnitude of the correlation between the series, i.e., we obtain complex eigenvectors.

For the EOF calculations we utilize the longest time series possible as defined by the common record length of the measurements involved. In order to isolate the wave band and guided by the result of the individual cross-spectral computations, we choose the frequency interval centered near 0.17 cpd and band-average 5 spectral estimates to obtain 10 degrees of freedom.

To give approximately equal weight to each latitude we have chosen two current meter records at each of the four latitudes (C2, P, Y, and M); to include the baroclinicity we choose the uppermost current meter and the current meter nearest 100 m. At C2,

Fig. 10. Coherence and phase between alongshore currents at C2 (86 m) and Y (97 m), for the same time period as in Fig. 9. Phase is positive for Y leads C2. The 95% significance level is shown on the coherence plot.
the 86 m and 126 m records had to suffice but at P, Y, and M the current meter records were from 37 and 96 m, 37 and 97 m, and 39 and 100 m, respectively. The common record length for the four locations is 45.25 days; we choose a frequency interval of 0.12–0.21 cpd for the EOF analysis. Figure 11 shows amplitude and phase relative to Y as a function of alongshore separation for the first EOF. The first EOF accounts for 72% of the total variance in the frequency interval. Also shown on Fig. 11 is the fraction of the variance in each record contributed to the first EOF. This measure corresponds to the coherence squared between each record and the first EOF; it is useful in order to decide whether the EOFs have a physical significance or are simply a fabrication of the statistical computations. With the exception of the deeper record at C2, each velocity record is highly coherent with the first EOF.

Figure 11 also shows that at each alongshore location the shallow and deep pair of records are nearly in phase, but that there is a poleward phase propagation alongshore. The phase is plotted relative to the Y-97 m record, and the line indicates the alongshore phase distribution expected for a 0.17 cpd wave with phase speed of 240 km day\(^{-1}\) poleward. We conclude that fluctuations in alongshore velocity propagate coherently between 5°S and 15°S, consistent with the results from individual cross-spectra, e.g., Fig. 10.

The amplitude of the first EOF is in the range 1–3 cm s\(^{-1}\) for C2 and P and 3–6 cm s\(^{-1}\) for Y and M, suggesting a doubling of the amplitude between 5 and 15°S. This may be the consequence of the conservation of wave energy flux. Assuming the alongshore group speed is constant, which is supported by the non-dispersive propagation of sea level fluctuations, the total energy density of the waves remains the same as it travels along the coast (neglecting frictional and other dissipative effects). Since the offshore scale of the wave decreases as it travels poleward, its amplitude must increase. An estimate of this increase is given by (see, e.g., Miles, 1972; Allen and Romea, 1980):

\[
\left(\frac{A_{15}}{A_5}\right)^{1/2} = 1.7,
\]

where \(A\) and \(f\) are local values of amplitude and Coriolis parameter, and the subscripts refer to latitude. The predicted amplification with latitude relative to Y-97 m is shown in Fig. 11 for the velocities near 100 m. This agreement between observations and theory is consistent with Brink's (1982a) conclusion that frictional decay scales for the Peru coast are very long and also helps explain why the sea level and current fluctuations propagate coherently over long distances.

4. Wind forcing

We can use the same EOF technique to analyze the coastal wind data. Figure 12 shows the statistics for the first two EOFs formed from the alongshore components (defined as the local principal axis) of the coastal winds at Talara, Chiclayo, Callao, and San Juan. The record length and frequency interval are identical to those for the alongshore current EOFs of Fig. 11. The first EOF contains 63% of the total variance while the second EOF contains 28%. We show both EOFs because the winds at Talara and Chiclayo fall primarily into the second EOF while the winds at Callao and San Juan fall primarily into the first EOF. This may be seen from the coherence plot in Fig. 12: the modal amplitude of the first EOF is largest for Callao and San Juan while the amplitude of the second EOF is largest for Talara. The phase of the first EOF suggests that the Callao wind leads the wind at San Juan with a phase consistent with poleward propagation at 200 km day\(^{-1}\). This phase behavior in the first EOF extends equatorward to Chiclayo but the coherence of the Chiclayo wind with the first EOF is very low. The phase structure of the second EOF indicates that the winds at Talara and Chiclayo are approximately in phase.
Using the long records, we have shown (Fig. 7) that the local winds at San Juan and the alongshore current at M are not coherent in the 0.1–0.2 cpd frequency band. Table 1 shows the coherence squared and phase from cross-spectral calculations using coastal winds at Talara, Chiclayo, Callao, and San Juan, and alongshore currents at C2, P, Y, and M. The record length and frequency interval are identical to those used in the EOF computations. The coherence is low (less than 95% significance level) between local winds and currents. However, there is a suggestion that the coherence is higher with the wind measured equatorward of the current (e.g., Chiclayo wind and Y-97 m).

We can use the EOF analysis on the mixed set of variables: wind and currents. The current velocities are normalized to unity total variance, as are the winds, to give the two a priori equal weights. This procedure was applied by Wang and Walsh (1976) and Brink et al. (1978) in problems involving modes of mixed quantities. Figure 13 shows the statistics for the first EOF for the winds and the alongshore currents nearest 100 m, computed for the same record length and frequency interval as in Fig. 11 and 12. The phase behavior of both current fluctuations and the wind is consistent with poleward propagation at about 200 km day⁻¹, suggesting the possibility of resonant forcing. However, the phase relation between currents and wind shows the currents leading by approximately 90°. In addition, all of the currents are highly coherent with the first EOF while none of the winds are. These calculations are consistent with those of Brink (1982b), who used observed winds and currents along the Peru coast as input to hindcast alongshore currents and sea level in the 0.1–0.2 cpd frequency band. The results from his forced-wave hindcast model suggest that most of the observed sea level and alongshore current fluctuations in the band are due to free waves originating equatorward of 5°S.
while the winds between 5 and 15°S contribute little to the observed variance.

We should point out that, in contrast to the alongshore current, fluctuations in the cross-shelf component are correlated with the local wind at time lags of less than a day (Brink et al., 1980; Smith, 1981). Coastal upwelling is occurring in response to the local wind, but less than 24% of the subtidal velocity variance in the water column is correlated with the wind off Peru, whereas more than 80% is off Oregon and Northwest Africa (Smith, 1981).

5. Vertical modal structure

a. Empirical modes

In order to examine the vertical structure of the fluctuations in the currents we compute vertical EOFs, using the alongshore velocities from C2 at 5°S and L at 15°S. These two locations provide the greatest depth range and the highest vertical resolution, and they are both located over the continental slope at about an equivalent distance offshore (scaled by the baroclinic radius of deformation). The estimated Rossby radius scale \( \delta_R \) is larger at 5°S than at 15°S on account of a factor of \( f \), \( \delta_R = c/f \) where \( c \) is alongshore phase speed. In Fig. 14, where we have scaled the offshore coordinate of each profile with the local \( f \), \( \delta_R \) spans the entire shelf–slope region at 5°S while at 15°S \( \delta_R \) extends out from the coast to about 200 m depth over the slope.

The common record length for C2 and L is only 40 days. Instead, we used the longest individual records (47.5 days for C2 and 53.75 days for L) in computing the vertical EOFs at each location. The frequency intervals were centered near 0.17 cpd and 10 degrees of freedom were maintained.

Figures 15a, b show amplitude, phase and coherence as a function of depth for the first EOF at C2 and L, respectively. In each case the first EOF contains most of the total variance in the frequency interval (82% at C2 and 87% at L). Time series of the EOFs were obtained following Wallace and Dickinson (1972a; see 1972b for details and an application of the method). The first EOF time series at C2 (5°S) and L (15°S), which were independently computed, are coherent (>95% significance level) with a phase difference indicating C2 leads L by 353° (or lags by 7°; but a lead of about 360° is the interpretation con-
sistent with Fig. 11). Moorings C2 and L are separated by 1300 km, and thus a phase difference of $353^\circ$ gives a phase speed of 225 km day$^{-1}$ (2.6 m s$^{-1}$) at the center frequency of the band (0.17 cpd). The first EOFs at C2 and L are also coherent (>95% significance level) in the frequency band with the nearby coastal sea level, Talara and San Juan, respectively.

Both EOFs are concentrated near the surface; the EOF at C2 has a zero-crossing at about 700 m depth, while the EOF at L presumably has a shallower zero-crossing which is below the deepest current meter (512 m). The coherence of each record with the respective EOF is high near the surface and very low near the zero-crossing of the EOFs, as would be expected. With the exception of the points near the zero-crossing, where the phase is poorly determined, both EOFs are within ±20° of being in phase over their whole depth.

Allen and Romea (1980) have shown theoretically that coastally trapped baroclinic fluctuations propagating poleward from the equatorial region may take the form of barotropic continental shelf waves at mid-latitudes. The EOFs calculated for vertical structure suggest that the transformation of modal structure has not taken place between 5 and 15°S.

### b. Dynamical Modes

In order to establish the physical significance of the EOF structure, it is necessary to show consistency with dynamical constraints. We have used the smoothed $N^2(z)$ profiles, shown in Fig. 16, to calculate the “flat bottom” dynamical vertical modes at 5 and 15°S, obtaining estimates of vertical structure, Rossby radius and alongshore phase speed for each mode. These $N^2$ profiles were obtained from CTD observations made during the CUEA experiment in 1977 and are typical. (Figure 12 of Smith, 1978, also shows the secondary maximum in $N^2$ between 300 and 400 m depth for a CTD station near 16°S in August 1976, suggesting that this structure is persistent in both space and time.)

The dynamical modes were computed by integrating the governing equations by means of a fourth-order Runge-Kutta scheme, with a trial-and-error procedure for determining the proper eigenvalue so that the boundary conditions are satisfied (see Kundu et al., 1975, Section 5 for a discussion of the equations and methods). Since the density profiles extend only to 780 m depth at 5°S and 980 m depth at 15°S, the $N^2$ profile has been extrapolated in both cases to $0.1 \times 10^{-2}$ s$^{-2}$ at 4000 m depth by an exponential profile below 600 m. (The first modes are insensitive to the details of the exponential extrapolation, e.g., doubling the value of $N^2$ at the bottom increases the calculated phase speed of the deep ocean mode by less than 10%.) The dynamical modes were computed assuming a flat bottom; the alongshore phase speeds, zero-crossing depths, and Rossby radii of the first dynamical modes corresponding to various flat bottom depths are listed in Table 2.

The dynamical modes computed using the actual depth at the moorings (1360 m at 5°S, 650 m at 15°S) give slow phase speeds and shallower zero-crossings

<table>
<thead>
<tr>
<th>Total depth (m)</th>
<th>Zero-crossing (m)</th>
<th>$c_1$ (m s$^{-1}$)</th>
<th>$\delta_R$ (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>(a) 5°S (1977)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1360</td>
<td>400</td>
<td>1.6</td>
<td>120</td>
</tr>
<tr>
<td>2000</td>
<td>760</td>
<td>1.8</td>
<td>140</td>
</tr>
<tr>
<td>4000</td>
<td>1680</td>
<td>2.4</td>
<td>190</td>
</tr>
<tr>
<td>(b) 15°S (1977)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>650</td>
<td>320</td>
<td>1.0</td>
<td>26</td>
</tr>
<tr>
<td>1400</td>
<td>510</td>
<td>1.5</td>
<td>39</td>
</tr>
<tr>
<td>4000</td>
<td>1610</td>
<td>2.5</td>
<td>66</td>
</tr>
<tr>
<td>(c) 15°S (1976)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>650</td>
<td>320</td>
<td>1.1</td>
<td>28</td>
</tr>
<tr>
<td>1400</td>
<td>500</td>
<td>1.6</td>
<td>42</td>
</tr>
<tr>
<td>4000</td>
<td>1550</td>
<td>2.6</td>
<td>69</td>
</tr>
</tbody>
</table>

**FIG. 16.** The $N^2(z)$ profiles based on CTD measurements near the Peru continental slope: at 4°58.9'S, 81°33'W, 1922 m water depth, 22 May 1977 (dashed line); 15°15.5'S, 75°40.0'W, 1353 m water depth 9 May 1977 (solid line).
c. Inertial motions

The local inertial frequency at the C2 mooring (5°S) is 0.175 cpd, which is exactly (to within the resolution of the spectral computations) the frequency at which the maximum alongshore coherence between the currents from 5°S to 15°S and between the sea level from 2°S to 15°S is observed. In other words, the 0.16–0.19 cpd frequency band is dominated by free baroclinic coastally trapped waves which propagate poleward through latitudes where their frequency is equal to the local inertial frequency. This is not surprising since Kelvin waves are well-behaved when their frequency is equal to $f$ (Pedlosky, 1979, p. 81), but it is of interest to examine the relative strength of the inertial motion and the coastally trapped waves at 5°S since the former would have been unavoidably included in the analyses presented in this paper. Evidence from the other latitudes suggest that the inertial signal is relatively weak (see Fig. 2 of Brink et al., 1980, and Fig. 9 herein).

We first use the rotary spectrum (Mooers, 1973) to examine the velocity records with the expectation that the inertial signal will be evident because of its preferred rotary sense. The method decomposes a two-dimensional vector time series into clockwise and counterclockwise rotating parts. The sum of the two components of the rotary spectrum yields the conventional kinetic energy spectrum at a given frequency.

Figure 18 shows $R$, the ratio of the variance of the counterclockwise to clockwise rotating parts for the velocity time series at C2 and L, calculated from the rotary spectrum, for the frequencies 0.094 cpd, 0.187 cpd (the nearest spectral estimate to the local inertial frequency at 5°S) and 0.516 cpd (the nearest spectral

![Figure 18](image_url)
estimate to the local inertial frequency at 15°S). Hourly data with a record length of 1024 hours were used for both C2 and L; the bandwidth for the spectral estimates was ±0.062 cpd. At 0.094 cpd, which is lower than the local inertial frequency at either 5° or 15°S, R is near unity for both C2 and L for all depths; there is no preferred rotary direction for frequencies less than f. To be significantly different from unity at the 95% confidence level requires R < 0.5 or R > 3. For 0.187 cpd, there is more energy in the counterclockwise direction at 5°S, but significantly so only at 197 m, while R remains near unity at 15°S, i.e., in the “wave band” the signal shows a rotary tendency at 5°S but is nearly rectilinear at 15°S. At 0.516 cpd, i.e., near the local inertial frequency at 15°S, the signal at L shows significantly more energy in the counterclockwise direction at most depths. At C2 (5°S), where the frequency 0.516 cpd is superinertial, R is not significantly different from unity. One concludes that inertial motion at 5°S is weak and is generally weak off Peru compared with other locations where $R \sim 100$ (Fu, 1981).

There is another way to attempt to separate the inertial signal at 5°S: We can use the same EOF technique and current velocity records used above to compute the vertical modes at C2, but instead use both velocity components (alongshore and onshore). The inertial signal should have about equal variance in both components and $u$ (onshore) should lead $v$ (alongshore) by 90°. Since inertial oscillations tend to have short vertical coherence scales (Fu, 1981), the inertial motion might show up as several different modes with a single depth dominating each mode. The first four EOFs, computed using both horizontal velocity components, account for 97% of the variance at C2 in the band 0.12–0.21 cpd. These modes are shown in Table 3.

The first EOF contains 56% of the total variance, and $v \gg u$ (the ratio $v/u$ is 3.6, 4.3, and 8 at 86, 126, and 197 m, respectively). Ratios $v/u \gg 1$ are expected for Kelvin waves. The alongshore component ($v$) shows a structure and phase very similar to that of the first EOF computed from the alongshore velocity only (cf. Fig. 15). The signals at 560 and 860 m both have a low coherence with the first EOF, and the $v/u$ ratios are low, reflecting the low amplitude of the first-mode internal Kelvin wave near its zero-crossing. The phase relations for the first EOF are such that the alongshore currents at all depths are in phase to within ±30° and the top three $u-v$ pairs rotate counterclockwise ($u$ leading implies counterclockwise rotation) while the bottom two $u-v$ pairs rotate clockwise.

This agrees qualitatively with predictions that the perturbation effect on free internal Kelvin waves due to topography gives $u-v$ pairs which are 90° out of phase with counterclockwise rotation in the upper layer and clockwise rotation in the lower layer (Allen and Romea, 1980). Presumably, the agreement with the theory is not more precise because the uncertainty in the observed phase relation for the $u-v$ pairs is large on account of the low coherence of $u$ with the first EOF. The first EOF shows no evidence of inertial oscillations; its characteristics are consistent with a first-mode baroclinic Kelvin wave.

The remaining three EOFs that contain non-negligible amounts of the total variance are less amenable to interpretation. Only at 197 m in EOF 4 is there reasonably clear evidence of an internal oscillation: $u$ leads $v$ by 84° and the amplitude of $u$ and $v$ are nearly equal. This velocity record was also the only 5°S record to show significantly greater counterclockwise energy than clockwise in the rotary-spectra calculations (Fig. 18). We conclude that the frequency band near 0.17 cpd, which at 5°S is shared by both

| Table 3. Amplitude (A), phase (θ) and coherence squared (γ²) for the first 4 EOFs of u-v pairs at C2 for the frequency interval 0.13–0.21 cycles per day, calculated from same record as in Fig. 15a. The amplitudes are normalized such that the sum of the squares of each EOF equals unity. EOFs 1, 2, 3 and 4 contain 56%, 20%, 14% and 7% of the total variance in the band, respectively. The phase is such that θ₁ > θ₂ implies 2 leads 1. The γ² value represents the fraction of the variance in the record occurring in the EOF. |
|---|---|---|---|---|---|---|---|---|---|---|---|
| C2 | A | θ | γ² | A | θ | γ² | A | θ | γ² | A | θ | γ² |
| 86 m | u | 0.14 | -90 | 0.16 | 0.16 | 1 | 0.08 | 0.37 | 37 | 0.31 | 0.61 | 66 | 0.41 |
| v | 0.50 | 12 | 0.83 | 0.05 | 6 | 0.01 | 0.24 | 86 | 0.05 | 0.36 | -114 | 0.06 |
| 126 m | u | 0.14 | -150 | 0.39 | 0.15 | -73 | 0.18 | 0.15 | -79 | 0.13 | 0.21 | 114 | 0.11 |
| v | 0.60 | -3 | 0.84 | 0.36 | -82 | 0.11 | 0.12 | -60 | 0.01 | 0.25 | -20 | 0.02 |
| 197 m | u | 0.06 | -57 | 0.09 | 0.05 | 93 | 0.03 | 0.14 | -88 | 0.14 | 0.46 | 9 | 0.73 |
| v | 0.48 | 7 | 0.82 | 0.25 | -113 | 0.08 | 0.22 | -152 | 0.04 | 0.32 | 93 | 0.05 |
| 560 m | u | 0.26 | 20 | 0.21 | 0.59 | 111 | 0.41 | 0.66 | -6 | 0.36 | 0.10 | -90 | 0.01 |
| v | 0.21 | -23 | 0.17 | 0.63 | 71 | 0.55 | 0.52 | 156 | 0.26 | 0.22 | 22 | 0.02 |
| 860 m | u | 0.03 | 59 | 0.20 | 0.06 | -80 | 0.37 | 0.02 | 146 | 0.07 | 0.07 | 113 | 0.21 |
| v | -0.05 | -3 | 0.34 | 0.06 | 148 | 0.19 | 0.08 | 89 | 0.28 | 0.07 | -2 | 0.09 |
baroclinic Kelvin waves and internal oscillations, is dominated by the former.

6. Summary and discussion

The analyses of the sea level and current velocity data from 2°12'S to 17°S along the west coast of South America show that the low-frequency (ω < 0.25 cpd) fluctuations propagate poleward along the coast with phase speed (2.5–3.0 m s⁻¹) consistent with a first-mode baroclinic Kelvin wave. The alongshore coherence is highest in the frequency band 0.1–0.2 cpd. The analyses done with long sea level records suggest the phenomenon extends to lowest frequency but the shortness of the current meter records, O(50 days), limits the more definitive analyses to ω > 0.1 cpd.

Frequency-domain empirical-orthogonal-function analysis of the combined current meter data from 5, 10, 12 and 15°S places more than 70% of the variance, in the 0.1–0.2 cpd band, into a mode showing poleward phase propagation (~2.75 m s⁻¹). Similar analyses of the current meter data from the vertical arrays at 5°S and 15°S again show that more than 80% of the variance in the 0.1–0.2 cpd frequency band is due to a mode whose vertical structure and alongshore phase speed are consistent with a first-mode internal Kelvin wave. The vertical structure of the EOF mode at both 5°S and 15°S is closely approximated by the structure of the first vertical dynamical mode, calculated using realistic stratification and a flat-bottom depth equal to the shelf-slope depth averaged over the local Rossby radius scale. However the alongshore phase speeds obtained with a deep ocean bottom depth (~2.5 m s⁻¹) are in better agreement with observations than speeds obtained with the averaged slope topography (~1.7 m s⁻¹).

The fluctuations in the 0.1–0.2 cpd frequency band do not seem to be the result of forcing by coastal winds, either near the equator (4°S) or locally. Although the winds along the coast show some alongshore coherence, albeit less than shown by sea level pressure or currents, they are not significantly coherent with the currents. The propagating fluctuations in the currents are of sufficient magnitude to mask the effects of local winds on the alongshore current, at least below the shallow surface Ekman layer (cf. Smith, 1981). This conclusion is corroborated by Brink’s (1982b) forced-wave hindcast model that uses observed winds and currents as input. The wind driving accounts for no more than 25% of the amplitude of the predicted current time series.

The interesting question remains: where and how are the fluctuations first energized? Luther (1980) suggests that a basin-wide barotropic oceanic response to large-scale atmospheric forcing exists in the 4–6 day period band at all longitudes from 60°N to 60°S latitude. We have shown that the fluctuations along the Peru coast are not a response to the atmospheric pressure fluctuations, which we assume are manifestations of the large-scale atmospheric systems. A more promising source is the equatorial waveguide.

Moore and Philander (1977) show that equatorially trapped baroclinic waves incident on an eastern boundary may be partially transmitted north and south along the coast as boundary-trapped internal Kelvin waves (see also Anderson and Rowlands, 1976; Cane and Sarachik, 1977). In general, the response at the eastern boundary to incident disturbances will consist of reflected waves for ω < ωc, and coastally trapped Kelvin waves for ω > ωc, where ωc = βc/2f, the coastal trapping condition for baroclinic disturbances (Section 2). La Libertad (2°12'S) is near the edge of the equatorial waveguide (δE ~ 260 km, Allen and Romea, 1980) and is presumably exposed to equatorially trapped disturbances. Waves (disturbances) with ω > 0.08 cpd (period of 12 days) would be transmitted poleward as internal Kelvin waves. According to the theoretical studies of Cane and Sarachik (1977) the response for lower frequencies would be primarily reflected waves but would exhibit “some of the characteristics of a Kelvin wave”. This may well explain both the weak evidence for poleward propagating disturbances at ω < 0.1 cpd and the strong evidence at frequencies ω ~ 0.1–0.2 cpd.

There is apparently no lack of equatorially trapped disturbances in the frequency band ω ~ 0.1–0.2 cpd. Wunsch and Gill (1976) show evidence that peaks in equatorial sea level are manifestations of first-baroclinic mode inertial-gravity waves. Luther (1980) reports peaks in spectra of observed sea level in the equatorial Pacific which correspond to the first-baroclinic first-meridional inertial-gravity wave with frequency 0.17–0.20 cpd and the second-baroclinic first-meridional mode with frequency 0.13–0.15 cpd. These modes are equatorially trapped with most of their energy equatorward of 4°. Ripa and Hayes (1981) have presented spectra of bottom pressure and temperature in shallow water from the western side of the Galapagos Islands at latitudes between 1°24'N and 0°59'S, which show relatively energetic baroclinic motions in the 0.18–0.2 cpd frequency band.

Unfortunately, the sea level record from Baltra (0°27'S, 90°17'W) at the Galapagos Islands is the only data available for the period of our other observations that enables us to extend our analysis, and attempt to trace the fluctuations in sea level, to the equator. The Baltra record has several gaps which prevent us from using a record of comparable length to those from La Libertad or Callao. However, three periods longer than 100 days each were available. An ensemble-averaged cross-spectrum between sea level at Baltra and La Libertad (and between La Libertad and Callao) was computed. The cross-spectrum of sea level at Baltra and La Libertad shows coherence
above the 90% significance level in the 0.187-0.214 cpd band. The peak coherence, which is significant at 95%, is at 0.205 cpd with Baltra leading by 398°. Baltra and La Libertad are separated by about 1050 km, which gives a phase speed of about 2.25 m s⁻¹. The cross-spectrum between sea level at La Libertad and Callao computed for the same record periods shows the 0.161-0.187 cpd band is coherent above the 90% significance level with a maximum coherence (99% significance) at 0.170 cpd. La Libertad leads Callao by 300° at maximum coherence, implying a phase speed of 2.75 m s⁻¹.

It should be noted that the phase relation between Baltra and La Libertad is ambiguous since we have no intermediate points to add confidence to our interpretation of the phase difference. Furthermore, the coherence between Baltra and La Libertad is not as good for the longer time series of sea level along the Peru coast. Nevertheless, this tenuous evidence suggests an equatorial origin for the several-day fluctuations in current and sea level along the Peru coast, and encourages us to continue the search as better tide gage data and longer current records become available in the eastern equatorial Pacific.

Acknowledgments. We are indebted to our colleagues and friends in Ecuador, Germany and Peru for data. The CUEA experiment, and much of the subsequent analyses, was supported by the National Science Foundation. A visit, sponsored by the Deutsche Forschungsgemeinschaft, to the Institut für Meereskunde (Kiel) enabled one of us (RLS) to begin the study. This paper has been completed with the support of National Science Foundation Grants OCE-8024116 and OCE-8017929. Helpful discussions with John Allen, Ken Brink and George Halliwell are acknowledged.

APPENDIX

Data Sources

A summary of the observations obtained during the ESACAN experiment is given by Brockmann et al. (1978). The complete set of data from moored instruments and hydrographic stations from the CUEA program has appeared in a series of data reports published by Oregon State University (e.g., Enfield et al., 1978; and Huyer et al., 1978). In both the ESACAN and CUEA experiments the currents were measured by Aanderaa current meters on taut subsurface moorings. The sampling interval was 10 min for the ESACAN array and 15 or 20 min for the CUEA arrays. These data were averaged to provide hourly data sets. The data from the CUEA anemometer at Callao and the subsurface pressure gages at P and near M were processed in the same manner as the current meter data. Except for the spectral computations (shown in Fig. 9, 10 and 18), the hourly data were filtered with a low-pass filter with a one-half-power point of 1.96 days (<1% amplitude at 1 day, 90% amplitude at 2.5 days), to remove diurnal and shorter-period variations, and decimated to six-hourly values. Currents at each mooring were rotated into a coordinate system in which the alongshore direction is defined by the averaged principal-axis direction of the vector time series at the mooring: C2 (15°) (counterclockwise from north), P (10°), Y (25°), M and L (45°). The winds also were rotated: Talara (15°), Chiclayo (0°), Callao (35°) and San Juan (45°).

Hourly tide gage and three-hourly atmospheric pressure data at Baltra and La Libertad were obtained from Instituto Oceanografico de la Armada (Ecuador). Hourly tide gage data at Talara, Callao, San Juan, and Matarani were obtained from Direcccion de Hidrografía y Navigacion de la Marina (Peru). Hourly atmospheric pressure data and wind data at Talara, Chiclayo, Callao, San Juan, and Matarani were obtained from Corporacion Peruana de Aerolineas Comerciales. The coastal tide gage at San Martin and additional coastal anemometers at Callao and San Juan, maintained by the CUEA program, provided hourly data.

The significance levels for the statistical calculations were obtained in the usual manner. In computing cross correlations, the degrees of freedom were found using the method suggested by Davis (1976) and the significance level obtained from Pearson and Hartley (1970, p. 146). In computing spectra and cross-spectral we relied on Jenkins and Watts (1968), in general, and Thompson (1979) for coherence significance levels.

REFERENCES

Davis, R. E., 1976: Predictability of sea surface temperature and


