Analysis and Interpretation of Deep Equatorial Currents in the Central Pacific*

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ABSTRACT

Analysis of vertical profiles of absolute horizontal velocity collected in January 1981, February and April 1982 in the equatorial central Pacific revealed two significant narrowband spectral peaks in zonal velocity, centered approximately at vertical wavelengths of 560 and 350 stretched meters (sm). Energy in the 560 sm band roughly doubled between the first and last cruises. Time-lagged coherence results suggested upward phase propagation at periods of about 4 years. East—west phase lines computed from coherence over zonal separations tilted downward towards the west, implying westward phase propagation and zonal wavelengths on the order of 10 000 km. The peak that was centered at 350 sm occurred at the vertical scales of the conspicuous alternating flows in the records, generically called the equatorial deep jets in the past (the same terminology is used here). It showed a more steady character in amplitude and a higher signal-to-noise ratio in comparison with the 560 sm peak. The deep jets were best defined as a finite narrowband process in vertical wavenumber (311—400 sm), accounting for 20% of the total variance present in the broad-band energetic background. At the jets' wavenumber band, latitudinal energy scaling compared reasonably well with Kelvin wave theoretical values and a general tilt of phase lines downward towards the east yielded estimates of 10 000—16 000 km for the zonal wavelengths. Time-lagged coherence calculations revealed evidence for vertical shifting of the jets on interannual time scales. Interpretation of both signals in terms of equatorial waves was ambiguous, because of their relatively long spatial and temporal scales compared to the records. The simplest hypothesis of linear waves in a resting basic state ocean could not be rejected, but more complicated physics cannot be ruled out.

At most wavenumber bands, power levels decayed away from the equator over scales broader than the Kelvin wave scale. Within ½° of the equator, zonal current led (lagged) vertical displacement by π/2 with depth for the 933 sm (140—400 sm) band. The result at the 140—400 sm band agree with the findings of Eriksen (1981) in the western Pacific, and thus seems to be a general feature of the deep equatorial Pacific fields.

1. Introduction

An interesting aspect of the deep equatorial circulation is the presence of strong, alternating zonal currents trapped to within 1° of the equator, and with vertical scales on the order of 100 m. After their discovery by Luyten and Swallow (1976) in the western Indian ocean, these energetic flows called the equatorial deep jets (EDJ) were also observed in the Pacific (Hayes and Milburn 1980; Leetmaa and Spain 1981; Eriksen 1981). A simple explanation was given in terms of surface forced vertically propagating linear waves (Wunsch 1977; McCreary 1984), but there appear to be difficulties in getting energy at low frequencies and high vertical wavenumbers to penetrate to great depth (McCreary 1984; McPhaden et al. 1986). More fundamentally, the existing records lacked the temporal and spatial coverage needed to test the validity of the general wave hypothesis. For example, Eriksen (1981, 1982) was able to infer the presence of Kelvin waves in his deep velocity records (collected in the western Pacific, near the Gilbert islands), based on coherence calculations showing zonal current lagging vertical displacement by π/2 with depth and the reasonable agreement found between direct current measurements and geostrophic velocities. However, to obtain statistical reliability in his cross-spectral analysis, Eriksen had to average over several vertical wavenumber bands, with inherent loss of resolution. Therefore, it was not clear whether his results represented a description of the apparent narrowband deep jets, or instead reflected the characteristics of the broadband background.

These earlier studies on the jets stimulated interest in the deep equatorial circulation and demonstrated the need for better data sets. In particular, evidence for vertical propagation of the jets was extremely difficult to document owing to, among other factors, the noisy character of the observations, the difficulty of distinguishing between phase and energy propagation effects and, perhaps most important, the lack of long enough records to resolve the apparent long time scales of the jets (Eriksen 1981; O'Neill and Luyten 1984). More generally, questions about the energy source of these
flows or what set their conspicuous vertical scales remained intriguing. The Pacific Equatorial Ocean Dynamics (PEQUOD) experiment which took place during the years 1981 through 1983, was designed to clarify some of these issues. The program included a moored current meter array (Eriksen 1985), a 16-month time series of vertical profiles of horizontal velocity along 159°W (Firing 1987), and a series of 76 vertical profiles of horizontal velocity taken in the central Pacific during 1981 and 1982. It is this last data set that will be considered here.

One of the main purposes of this study is to investigate whether the spatial and temporal scales of the jets (or any other energetic signals present) as observed during PEQUOD correspond to a point (or a region) in the dispersion curves for equatorially trapped waves, thereby testing the linear wave hypothesis advanced in earlier studies. The rest of this paper is organized as follows. A complete description of the data and its general treatment is given in section 2. Analysis of the strong zonal velocity signals (including the ubiquitous jets) and their interpretation in terms of linear equatorial waves is provided in sections 3 and 4. In section 5, we examine the evidence for equatorial waves at the other vertical scales, in both the velocity and vertical displacement records. The final section includes a summary of the main results and a general discussion.

2. The data and its general treatment

The PEQUOD data analyzed here consists of 76 vertical profiles of horizontal velocity, pressure, temperature, and conductivity (salinity), collected during three cruises to the central Pacific which took place in January 1981, February and April 1982 (Voorhis et al. 1984). The profiles were obtained at 17 different sites along and across the equator (Fig. 1), approximately spanning 15° of longitude and 6° of latitude. Casts were repeated in all three cruises. The instrument used in collecting the data was the White Horse profiler, a freefall acoustic dropsonde also employed by Luyten and Swallow (1976) and Eriksen (1981) in earlier observations of the deep jets. Its characteristics and performance are discussed in detail by Luyten et al. (1982). The instrument is tracked acoustically by means of three transponders placed on the ocean floor, and its horizontal deviations as it travels vertically in the water column essentially give the horizontal transport averaged vertically over roughly 15 meters (Voorhis et al. 1984). The two profiles (up and down trace) obtained for each cast were averaged together and interpolated at 25 m depth intervals.

Zonal velocity records (Figs. 2 and 3) revealed a strong Equatorial Undercurrent (EUC) flowing eastward in the upper ocean and a series of alternating

**FIG. 1.** Location of the sites occupied in the central Pacific during PEQUOD.
eastward and westward flows below it, generally referred to as the EDJ. These short vertical scale currents were correlated over at least $10^\circ$ of longitude and strongly trapped to the equator. Some of the jets remained at nearly the same depth from cruise to cruise, while others migrated vertically by several tens of meters. The records suggested long time and space scales associated with zonal velocity flows, contrasting with the high variability present in the meridional velocity profiles in general (not shown).

Vertical displacement is not directly measured by the White Horse instrument, but can in general be inferred by estimating the deviations of the density at each depth from its spatially and temporally averaged value, divided by the mean vertical density gradient (e.g., Eriksen 1981). This method is frequently used in internal wave studies. The temperature and salinity data, previously interpolated to 2 db pressure intervals, were used to calculate density profiles for each cast, from a linear least squares fit of specific volume over

**Fig. 2.** Zonal velocity along a longitudinal section at the equator, taken during the April 1982 cruise. All casts collected at each site were averaged together.

**Fig. 3.** As in Fig. 2 but along a latitudinal section at $138^\circ$W; there was no velocity data from site A at $4^\circ$N.
10 db at the top, and 50 db below 1000 db. We then averaged horizontally all the 76 density profiles and their corresponding vertical gradient profiles (the slopes from the linear fit over each pressure bin), to obtain the mean density and mean density gradient, respectively. Since we use data from only two different years in estimating the mean quantities, we do not expect to resolve the interannual variability (if any) in the vertical displacement records. Vertical displacement records were interpolated to 25 m depth intervals using cubic splines, and cut to the length of the shortest cast available. Profiles of vertical displacement (Fig. 4) show in general more variance than the velocity records, perhaps due to the inherently noisy character of the procedure outlined above. Near-surface values are suspicious, because of the extraneous effects that mixed layers and surface heat fluxes can have on the calculations. They will not be used in the analysis.

To apply conventional spectral techniques to the records, they were made vertically homogeneous by a WKBJ stretching and scaling procedure (O'Neill and Luyten 1984). The method has worked reasonably well in previous studies (e.g., Eriksen 1985). It corrects for changes in vertical scale and amplitude which occur with depth in a vertically stratified ocean. The new stretched vertical coordinate and scaled variables (denoted by an asterisk) are given by

\[ dz^* = \frac{N(z)}{N_0} dz \]  
\[ (u^*, v^*) = \left[ \frac{N_0}{N(z)} \right]^{1/2} (u, v) \]  

where \( N(z) \) is the buoyancy frequency, \( N_0 \) is an arbitrary reference value, and \( u, v \) and \( z \) stand for zonal and meridional velocity, and vertical displacement, respectively.

The buoyancy frequency profiles needed to carry out the stretching and scaling procedure were calculated using a linear least squares fit of specific volume over 10 db at the top, and 50 db below 1000 db as before. No significant systematic spatial variations were found in the buoyancy frequency profiles below the thermocline, so that we averaged horizontally over all profiles obtained during each particular cruise. Plots of the averaged \( N(z) \) for the January 1981, February and April 1982 cruises (Fig. 5) show no substantial differences, especially below 500 m. Also shown in Fig. 5 is the profile obtained by averaging in time over all three cruises. We chose \( N_0 = 0.189 \times 10^{-2} \text{ s}^{-1} \), to have a rough correspondence between the total depth of the records in the stretched and unstretched vertical coordinates (for this value, \( \Delta z \sim \Delta z^* \) around 1400 m).

Vertical displacement and velocity profiles were stretched and scaled according to (1–3), using the respective \( N(z) \) for each cruise. For convenience, records used in the time-lagged coherence analysis of section 3c were stretched and scaled with the same \( N(z) \) (shown in Fig. 5d). The use of the stretched coordinate gave records which were not equispaced. Spline interpolation to 25 cm grid was performed on the data. Features with wavelengths shorter than twice the largest stretched interval became aliased in the new coordinate.

![Figure 4](image.png)

**Fig. 4.** Vertical displacement along a longitudinal section at the equator during April 1982. All profiles calculated at each site were averaged together.
system. Results for wavelengths shorter than 80 sm should be regarded with suspicion. Scaled zonal velocity profiles in Fig. 6 appear more homogeneous in amplitude and vertical scale with depth than the raw profiles of Fig. 2, especially below the EUC. This was in general true for the other profiles.

The presumed homogeneous records were all cut to a standard length of 2800 sm, where approximately the upper 1500 sm of data were dropped, roughly corresponding to the thermocline region where the WKBJ stretching and scaling procedure is not valid (Eriksen 1981; O’Neill and Luyten 1984). Fourier decomposition was performed in each profile, after subtracting the mean and tapering 10% of the data at both end points with a ½ cosine bell. The Fourier coefficients were used in power spectra and coherence calculations. In all the spectral computations, we relied heavily on piece-averaging to gain statistical reliability. Our intention was to have the best possible resolution of the vertical scales present in the data. Each cruise data set was first treated separately, to reveal any variability over time, but in general we grouped all the data together to increase the number of degrees of freedom. Given the apparent long temporal and spatial scales associated with the short vertical scale features in Figs. 2 and 3, only one independent realization of deep jets structure was probably sampled. Thus, statistical inference may be interpreted as representing an estimate of errors introduced by the presence in the records of “noise” fields with decorrelation scales comparable to the time and space separation between observational sites.

Preliminary spectral calculations using all 32 equatorial casts available revealed two strong narrowband peaks appearing in zonal current, roughly centered at

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Fig. 5. Vertical profiles of buoyancy frequency $N$ in cycles per hour (cph), computed from CTD data collected during (a) January 1981, (b) February 1982, (c) April 1982. Time average of other three profiles is shown in (d).

Fig. 6. Same velocity profiles of Fig. 2 after being stretched and scaled as described in the text.
wavelengths of 560 and 350 sm (Fig. 7), with no counterparts in the $v$ and $\varphi$ spectra. The 95% confidence error bars were computed according to Koppmans (1974), with 64 degrees of freedom (i.e., two for each profile). The peak at 560 sm is statistically significant although it was not apparent from a visual inspection of the records. The peak at 350 sm rises higher above the background as compared with the other feature. It accounts for roughly 20% of the total variance in the records, and occurs at vertical scales corresponding to the scales of the conspicuous wiggles present in the $u$ velocity profiles (Figs. 2 and 6), and generically called the equatorial deep jets in the past. Furthermore, its amplitude was more steady in time than the feature at 560 sm (Ponte 1988a). For these reasons, we will refer to the peak centered at 350 sm as being the EDJ peak. Previous studies (Eriksen 1981; O’Neill and Luyten 1984) were unable to portray the EDJ as a well-defined power peak in the zonal velocity spectrum, perhaps because they had to resort to wavenumber band averaging to gain statistical reliability. The zonal velocity signals are treated separately from what we call the background spectra. Their analysis and interpretation constitutes a major part of this work.

3. Scales of the zonal velocity signals

With the spectrum of Fig. 7 quantitatively defining the vertical wavelength of the zonal velocity signals in the records, we focus our attention here on the determination of their temporal and spatial (zonal and meridional) scales.

a. Latitudinal energy scaling

Figure 3 qualitatively revealed the equatorially trapped nature of the high baroclinic mode flows. Coherence between equatorial and off-equatorial $u$ profiles showed significant amplitudes at zero phase for the $u(0^\circ)/u(0^\circ 30^\prime)$ pair, at the wavenumbers of interest. Coherence amplitudes dropped below the 95% zero significance level for the pair $u(0^\circ)/u(1^\circ 15^\prime)$, implying latitudinal scales for the signals on the order of 100 km.

Power spectra calculations using off-equatorial records provided a more quantitative estimate of this meridional energy scaling. Equatorially symmetric and antisymmetric profiles of zonal velocity were obtained by adding and subtracting records from equal but opposite latitudes (Ponte 1988a). Symmetric power spectra at $0^\circ 30^\prime$ showed the same two peaks present at the equator (see Fig. 7). Antisymmetric power spectra had no special structure and contained in general less variance than their counterparts. The ratios of the symmetric power at $0^\circ 30^\prime$ and $1^\circ 15^\prime$ to that at the equator for the wavenumbers of interest are given in Table 1, together with the predicted values from Kelvin wave theory (zonal kinetic energy should scale as $\exp(-\{2\pi\beta y^2/N_0^2\})$, where $\beta$ is the planetary vorticity gradient and $\varphi$ is the stretched vertical wavelength). At $1^\circ 15^\prime$, energy at the signal wavenumbers has decayed to only 10% of equatorial levels. Agreement between the observed and Kelvin meridional energy scaling is reasonably good (especially for the 560 and 350 sm wavelengths), considering the size of the error bars on the spectral estimates.

b. Zonal scales and estimate of zonal wavenumbers

To estimate zonal decorrelation scales for the signals, as well as zonal wavelengths from any east–west tilt of phase lines, coherence over zonal separations was computed using casts collected at the equator and at $0^\circ 30^\prime$, where the signals were clearly present.

<table>
<thead>
<tr>
<th>$\varphi$ (sm)</th>
<th>0°30′</th>
<th>1°15′</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>SYM/0°</td>
<td>$\epsilon^*$</td>
</tr>
<tr>
<td>700</td>
<td>0.50</td>
<td>0.69</td>
</tr>
<tr>
<td>560</td>
<td>0.61</td>
<td>0.64</td>
</tr>
<tr>
<td>467</td>
<td>0.85</td>
<td>0.59</td>
</tr>
<tr>
<td>400</td>
<td>0.98</td>
<td>0.53</td>
</tr>
<tr>
<td>350</td>
<td>0.43</td>
<td>0.48</td>
</tr>
<tr>
<td>311</td>
<td>0.58</td>
<td>0.44</td>
</tr>
</tbody>
</table>
the longitudinal distribution of casts, several zonal lags were possible. At each zonal lag, usually more than one realization was available, so that generally no band averaging was used in the coherence calculations. We grouped together pairs which did not have exactly the same zonal lags (e.g., 2° of longitude separate E and I casts in Fig. 1, while K and P casts are separated by 3°, even though both pairs were taken together in the 2°30' lag calculations). Such procedure was justified since both signals have long zonal scales, compared with the lags under consideration. The lag actually assigned to each calculation is the average of the lags from each pair of casts used. Off-equatorial profiles were only used for the 7°15' zonally lagged coherences.

Results from calculations done using all the data are summarized in Table 2. The 95% zero significance levels at the 2°30', 5°15', 7°15', 9°45' and 15° lags were computed based on 24, 22, 28, 10, and 4 degrees of freedom respectively, treating all available realizations as independent. The zonal coherent character of the EDJ is especially seen in the 350 sm band, where records 1200 km apart were still significantly correlated (if more realizations at the 15° lag were available, we would probably find the EDJ to be zonally coherent over the whole PEQUOD array). This result contrasted with that of Eriksen (1981) near the Gilbert islands. He found zonal decorrelation scales for the deep jets shorter than 500 km, but the vertical wavenumber band averaging used in the coherence calculations and the proximity of the Gilbert island chain to the site of the measurements may have influenced his results. At 560 sm, values of Table 2 may be reflecting the very high coherence amplitudes observed during the last cruise, at lags as long as 10°. Individual cruise results suggested zonal decorrelation scales shorter than 2°–3° in January 1981 and February 1982, a fact perhaps related to the lower signal-to-noise ratios observed during the first two cruises for this particular wavenumber band (Ponte 1988a).

While in the EDJ band, we found mostly negative phases, meaning lines of constant phase slope downward to the east, the positive phases at 560 sm implied a zonal tilt in the other direction. Figure 8 shows phase as a function of zonal lag for the 350 sm and 560 sm wavelengths, using values of Table 2 and the April 1982 cruise results, respectively (these were the cases for which the zonal coherence was the most significant). Error bars drawn were computed as described in Koopmans (1974), based on the number of degrees of freedom previously given. The sign of the phase for the 560 sm vertical wavelength is indeterminate (but not for 350 sm). An estimation of the zonal wavelengths λ underlying the records was carried out by simply performing a linear regression on those plots. Results displayed in Table 3 suggest that both signals have basin scale zonal wavelengths, but the large uncertainty in the estimates given is clear.
TABLE 3. Estimates of zonal wavelength $\lambda$ in kilometers from the zonally lagged calculations, using data plotted in Fig. 8. $\lambda_{\text{min}}$ is obtained from lines with greatest possible slope still contained within the phase error bars of Fig. 8.

<table>
<thead>
<tr>
<th>$\vartheta$ (sm)</th>
<th>$\lambda$</th>
<th>$\lambda_{\text{min}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>560</td>
<td>12 400</td>
<td>6 200</td>
</tr>
<tr>
<td>350</td>
<td>15 800</td>
<td>10 100</td>
</tr>
</tbody>
</table>

c. Time scales and vertical propagation

To check whether the signals remained stationary or shifted in the vertical during our observational period, time-lagged coherences were computed between zonal velocity profiles from different cruises. Although we only have essentially three data points in time, temporal lags between cruises were long enough to check for vertical propagation at interannual time scales. Again only records collected at latitudes of 0° and 0°30' were used. Casts collected at one site during the same cruise were averaged, since coherences done between them usually showed zero phase lags at the wavenumbers of interest. Each of the equatorial and near equatorial sites provide one realization of coherence, giving a total of ten realizations at time lags of 2 months, and nine realizations at time lags of 13 and 15 months.

The 13 and 15 month time-lagged coherence results presented in Fig. 9 showed amplitudes significantly different from zero at the 95% confidence level (based on 18 degrees of freedom) for the 560 and 400–311 sm bands. The coherence phases for these bands (with their respective 95% confidence error bars) are shown in Fig. 10 as a function of time lag. Positive (negative) phase implies downward (upward) shifts with time in the records. Most striking is the occurrence of apparent downward propagation over the 400–311 sm band, and the clear upward propagation of the 560 sm signal, at time scales of at least several years. Even though the error bars on some of the phase estimates are not small, they do not embrace the zero phase line in general. A linear trend in the phase versus time lag plots of Fig. 10 is especially apparent for the 560 sm case. An attempt at establishing bounds on the wave periods suggested by the time-lagged coherence phases is deferred until the next section.

4. The wave interpretation

We sought the most simple explanation of results from previous sections in terms of linear equatorial waves (e.g., see Moore and Philander 1977 or Gill 1982 for the theoretical treatment of these waves). The zonal
extent, meridional resolution and duration of the records were not sufficient to determine with confidence the apparent long spatial and temporal scales of both signals. Nevertheless, it was important to establish whether the observations precluded the linear wave hypothesis.

a. The 560 sm signal

The apparent linear relation between phase and time lag (Fig. 10) suggested the presence of an extremely narrowband process in the frequency domain. Fitting a straight line through the phase points yielded an estimate of four years for the period $T$, with phase error bars giving the range

$$2.8 \text{ yr} < T < 6 \text{ yr}. \quad (4)$$

The long time scale of the signal is very pronounced. The sense of propagation is upward. This sense, coupled with the zonal tilt of phase lines downward towards the west, implies westward propagation (lines of constant phase are perpendicular to the phase velocity vector). Given the low frequency, low zonal wave-number character of the signal, only long Rossby waves could explain these results.

The dispersion relation for long Rossby waves, written in terms of period $T$, zonal wavelength $\lambda$, and vertical wavelength $\vartheta$, takes the simple form

$$T = 2\pi \left( \frac{2j + 1}{N_0} \right) \frac{\lambda}{\vartheta} \quad (5)$$

where $j = 1, 2, \cdots$ is the meridional mode number. Only odd-numbered meridional Rossby modes are relevant, since even modes have a node in zonal velocity at the equator. If the signal were to be interpreted in terms of a long Rossby wave, its time and space scales should be consistent with the respective dispersion relation, i.e., the range of periods defined in (4) should overlap with the range of periods computed from using zonal wavelength estimates of Table 3 in (5). This was indeed the case only for the first and third meridional modes, as shown in Table 4. Furthermore, the latitudinal structure of the first three odd meridional Rossby modes for waves with $T = 4$ yr and vertical wavelength of 560 sm, which is shown in Fig. 11, would imply a phase reversal of 180° between records at 0° and 0°30' for modes $j = 3, 5$ (in fact, latitudes of 0°30' almost coincide with the first node of mode $j = 3$). This was not seen in the data. Therefore, it was reasonable to conclude that we were most likely in the presence of a long first meridional Rossby mode.

The ratio of the symmetric energy at latitudes of 0°30' to that at the equator (see Table 1) was considerably larger (by roughly a factor of two) than what the theory predicted for a first meridional Rossby wave mode with the vertical scale of the signal (Ponte 1988a). Since the lower bound on the energy ratio, calculated using the 95% confidence error bars on the power density estimates, encompassed the theoretically predicted value, one can not reject the Rossby wave interpretation based on these results. In addition, apparent zonal inhomogeneities in the records [energy levels at site $Q$ (Fig. 1) were lower by a factor of 3 than at other sites to the east] may have contributed to a higher estimate of the energy ratio, because off-equatorial power estimates came from only two longitudes (138°W and 145°W), while equatorial estimates included casts from a wider range of longitudes (Fig. 1). Whether the higher than expected energy ratio reflects the uncertainty in the spectral estimates or a more fundamental inconsistency of the Rossby wave hypothesis with the records is an issue which can not be solved with these records.

b. The deep jets signal

In the most simple interpretation involving a single frequency process, the information in Fig. 10 seemed to imply downward phase propagation (and corresponding upward energy propagation) of the jets at a range of periods displayed in Table 5. If we consider that, by expressing the frequency as $\omega = c_mm$ where $m$
TABLE 5. Estimates of $T$, $T_{\text{max}}$, and $T_{\text{min}}$ (in years), based on linear least squares fits of the time-lagged coherence phases plotted in Fig. 10 for the jets band. $T_{\text{max}}$ and $T_{\text{min}}$ are calculated using lines with smallest and greatest slope still contained within the phase error bars.

<table>
<thead>
<tr>
<th>$\theta$ (sm)</th>
<th>$T_{\text{min}}$</th>
<th>$T$</th>
<th>$T_{\text{max}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>400</td>
<td>5</td>
<td>6.7</td>
<td>12</td>
</tr>
<tr>
<td>350</td>
<td>6</td>
<td>9.1</td>
<td>14</td>
</tr>
<tr>
<td>311</td>
<td>8.5</td>
<td>17</td>
<td>56</td>
</tr>
</tbody>
</table>

is the vertical wavenumber, the vertical group velocity $c_{g_z}$ (i.e., $\partial m \omega$) can be written as

$$c_{g_z} = c_z + m \frac{\partial c_z}{\partial m}$$

(6)

one can then estimate its value from the results of Table 5, given that $\left| c_z \right| = \theta / T$. The inverse relation between period and wavelength in Table 5 implies that $\partial m c_z > 0$ and in fact, a rough finite difference estimate of this slope evaluated at the middle wavelength yields positive values for $c_{g_z}$, a result which curiously points again towards upward propagation of energy.

A difficulty with the single frequency hypothesis was that it led to values for the period in Table 5 which were extremely long ($T_{\text{min}} = 5$ yr). The apparent downward phase propagation and the tilt of phase lines downward towards the east implied westward phase speeds, leaving again long Rossby waves as the only possible candidates for explaining the signal. Use of the dispersion relation (5) and zonal wavelengths estimates of Table 3 yielded an even longer value of $T_{\text{min}} = 9$ yr. In addition, the observed energy meridional structure was not consistent with a first meridional mode Rossby wave and higher modes would require phase reversals in latitude which were not seen in the data.

Another reasonable possibility was to have a finite frequency bandwidth underlying the process, in which case the behavior of phase with time is not strictly linear, and the scale of phase propagation can not be inferred from the results of Fig. 10 (Ponte 1988a). The fair agreement between observed energy meridional scaling and predicted Kelvin wave scaling (especially seen at 350 sm in Table 1) suggested a basic Kelvin wave character for the jets. From the dispersion relation, the zonal wavelength estimates of 10 000–16 000 km of Table 3 gave a range of periods from 3 to 5 yr, approximately. Kelvin waves and the observed zonal tilt of phase lines implied predominantly upward phase propagation in the records. A Kelvin wave packet of such finite bandwidth in frequency, propagating energy downward, could still exhibit the apparent "wrong" phase behavior displayed in Fig. 10 over relatively short times compared with the periods of the waves involved (Ponte 1988a), although for long enough records one should observe upward phase propagation. The Kelvin wave interpretation offered here can account for our observational results, but to test its validity, much longer records than the ones used here would be needed.

Other available profiler data collected roughly in the same region (Leetmaa and Spain 1981; Firing 1987) could be used to extend the records in time. Given the finite narrow band character of the jets and the energetic background, their long time scales and the opposite direction between phase and energy vertical propagation for equatorial waves, to obtain reliable results a careful Fourier analysis of all data sets would be needed, but this is beyond the scope of this paper.

Firing (1987) explicitly compares his data with the profile collected by Leetmaa and Spain (1981) at 159°W in 1980, and he finds a good match between them, if the profile obtained by averaging all his equatorial data is shifted downward by 130 m or upward by 220 m. If these depth changes were caused by vertical propagation, they would imply for a single frequency process either a 6.5 year period wave propagating downward or a 4.1 year wave propagating upward. He argues that the expected vertical displacements of the jets (65 m down or 110 m up) are not apparent from a visual inspection of his velocity records, thus concluding that "...the jets did not propagate steadily from 1980 to 1983." At the same time, he also notes that many of the jets are shallower by roughly 50 m at the end of his 16-month records. Firing's analysis not only suggests that the jets do indeed seem to shift vertically over long enough periods of time, but also hints at the presence of a process with a finite bandwidth $\Delta \omega$, in which case simple relations of phase with time will most probably not be observed (Ponte 1988a). Thus, the issue of bandwidth is very important in the interpretation of the records.

c. Forcing and nonlinearity

The energy sources for the observed signals and the mechanisms responsible for selecting their vertical scales remain an interesting question. One way to explain the appearance of energy at a particular dominant vertical scale in the ocean is to have a peak in the atmospheric forcing spectra at appropriate frequency and zonal wavenumber bands. Although knowledge of the atmospheric forcing functions is still inadequate to resolve their spectral character at low frequencies (Goldenberg and O’Brien 1981; Luther and Harrison 1984), and information on the zonal wavenumber spectral structure is virtually nonexistent, it is tempting to relate the observed signals with the strong basin scale interannual variability associated with El Niño events occurring every four years or so in the tropical Pacific.

In addition to peaks in atmospheric forcing spectra, instability processes may also give rise to energetic motions at particular scales for which maximum growth rates occur. In some cases the most unstable wave may
occur at a given frequency and zonal wavenumber, thereby setting the vertical wavelength of the flow, while in other cases the instability may directly favor a given vertical wavenumber. Present knowledge of equatorial instabilities is by no means satisfactory, but a few available studies on the subject (e.g., Philander 1976, 1978; Boyd and Christidis 1982, 1983; Hayashi and Young 1987) suggest (for parameter values which lead to a mean flow qualitatively resembling equatorial mean flows) zonal and time scales for the fastest growing disturbances which are too small to explain the signals observed in the PEQUOD data. It thus remains to be seen whether the deep energetic zonal currents found in the records are somehow related to instabilities of the equatorial ocean basin state.

Since low frequency, high vertical wavenumber equatorial waves have very small vertical group velocities (e.g., McCreary 1984), it is far from obvious how energy associated for example with El Niño-type variability in the upper ocean could penetrate to great depth in the presence of moderate amounts of damping. Although in some special situations, as in the presence of vertically sheared mean zonal flows discussed in Ponte (1988b), the vertical penetration of energy can be significantly enhanced, a simpler alternative is to hypothesize a deep energy source. Plausible mechanisms to excite the equatorial waveguide include fluctuations in cross-equatorial deep water mass fluxes (e.g., see Weiss et al. 1985), as supported by the recent work of Kawase (1987). The oceanic response to forcing mimicking these processes shows some similarity to the observed records (Ponte 1989).

The wave interpretation of the signals is perhaps indicative of their basic linear character. Given the values for the periods \( T \approx 4 \text{ y} \) and zonal wavelengths \( \lambda \approx 12,000 \text{ km} \) obtained from the wave analysis, we can estimate the degree of nonlinearity by comparing the magnitude of the terms \( uu_c \) and \( u_c \) in the zonal momentum equation. Choosing a velocity scale \( U \) is a little ambiguous, since it depends on whether a single or several vertical wavelengths are considered (i.e., although a single wave may have small enough amplitude to be linear in character, the sum of a number of such waves may be nonlinear since the resulting velocity field may become large enough, at least locally). If we take the amplitudes observed for individual vertical wavenumbers, then \( U \approx 5 \text{ cm s}^{-1} \), yielding a value of the ratio \( uu_c / u_c \approx 0.5 \). On the other hand, the actual amplitudes of some of the jetlike flows exceed 10 cm s\(^{-1}\) (e.g., see Fig. 2), implying this ratio to be of order one or larger. Thus, although the individual waves invoked here to explain the basic character of the signals may be reasonably linear in character, their interaction may be nonlinear.

5. The background spectra

Deep equatorial fields have been interpreted in terms of broadband sums of equatorial wave modes (e.g., Eriksen 1981). If deep motions are associated with wave processes, one expects, among other things, energy to be trapped to the equator and significant correlations between horizontal velocity and vertical displacement (e.g., Eriksen 1982). Study of the broadband background spectra seemed to confirm this idea.

Spectra of \( u, v, \) and \( \zeta \) computed using both piece-averaging over all realizations and band-averaging over four adjacent wavenumbers are shown in Fig. 12 as a function of latitude. In general, energy decreases away from the equator (perhaps most evident in the velocity records). The meridional decay scales for zonal current and vertical displacement are much broader than the Kelvin wave scaling at each vertical wavenumber (see Fig. 12), suggesting the existence of Rossby waves in the records.

For each equatorial wave mode (Kelvin, Rossby, mixed Rossby–gravity), there is a definite phase relation between the dynamical fields \( u, v \) and \( \zeta \) (e.g., Eriksen 1982; Ponte 1988a), but in the presence of a mixture of modes, the phases can take on any values. Thus, inferring the presence of a certain wave type in the data from observed coherence phases is ambiguous. To interpret results, one implicitly assumes that a particular wave type contributes predominantly to the records at a given wavenumber band.

Cross-spectral analysis (with no band-averaging) between \( u \) and \( \zeta \) profiles at both 0° and 0°30' latitude revealed significant coherence amplitude with \( u \) leading \( \zeta \) by \( \pi / 2 \) (within error bars) with depth, at the 933 sm band (see Fig. 13). This suggested a dominant presence of Rossby waves in the zonal velocity records at this wavelength. Fig. 13 also shows \( u \) lagging \( \zeta \) by \( \pi / 2 \) with depth at the equator, over the 400–140 sm band. Similar results were found at 0°30' latitudes for some of these wavenumbers. This phase relation between \( u \) and \( \zeta \) is typical of Kelvin waves. Since the 400–140 sm band includes the deep jets, one may think that the results at 400–311 sm are primarily a manifestation of the Kelvin wave signature of the jets signal. However, given the lack of energetic waves in \( \zeta \) power spectra at the jets wavenumbers (Ponte 1988a), perhaps due to poor resolution of the interannual variability in the \( \zeta \) records as explained in section 2, it is more likely that the coherence between \( u \) and \( \zeta \) is a reflection of the wave composition associated with the background energy.

6. Summary and discussion

Spectral analysis of the PEQUOD data set unveiled two narrowband peaks in zonal current, centered at vertical wavelengths of 560 sm and 350 sm. At the 560 sm band, coherence calculations yielded upward phase propagation at periods of about 4 years and basin-scale zonal wavelengths, with phase lines in the zonal–vertical plane tilting downward towards the west. These results suggested the presence of a first meridional mode.
long Rossby wave, but comparison of the observed and theoretically expected latitudinal energy structure was ambiguous. The apparent evidence for Rossby waves at interannual time scales is new in the deep ocean, but considerable Rossby wave energy at long periods has been observed in the upper ocean (e.g., Magaard 1983 found a peak in potential energy spectra at periods of roughly seven years, from records covering the area 20°–25°N and 130°–175°W in the north Pacific).

The equatorial deep jets, identified with the peak centered at 350 sm, were best defined as a finite narrowband process in vertical wavenumber (311–400 sm), and accounting for 20% of the total variance in the records. At the 350–311 sm wavelengths, the meridional structure of kinetic energy followed reasonably well the Kelvin wave scaling. Phase lines tilting downward towards the east yielded basin scale estimates for the zonal wavelengths. The jets shifted coherently in the vertical during PEQUOD, but definitive determination of their apparent interannual time scales would require records much longer than the ones used here (on the order of at least five years). Interpretation of results in terms of a single frequency Kelvin wave process led to inconsistencies, but finite bandwidth (in frequency and wavenumber) Kelvin wave processes of periods on the order of three to five years and basin scale zonal wavelengths remain a possibility.

The wave interpretation of the zonal current signals is limited by the relatively short spatial and temporal coverage provided by the data, when compared to the time and space scales involved. Although the simplest conceptual model discussed here (equatorial waves in a motionless basic state) remains a possible framework to explain the signals, some of the difficulties encountered in the interpretation may be due to the importance of physical processes which we have not contemplated (e.g., interaction of waves with background flows). Despite this possibility, it would be premature to consider more complicated models when quantitatively explaining the results, given the length of the records available.

The background spectra of velocity and vertical displacement records showed in general energy decaying away from the equator (especially in the velocity records). The decay scales for \( U \) and \( \xi \) were much broader than the Kelvin wave scaling at each band, suggesting the existence of Rossby waves in the records. Coherence phases between zonal velocity and vertical displacement suggested that Rossby and Kelvin waves dominated the zonal velocity records in the 933 sm and 140–400 sm bands, respectively. The latter result is strikingly similar to the findings of Eriksen (1981), using a western Pacific data set. Considering that the two experiments took place four years apart and at locations separated by roughly 2000 km, this broadband Kelvin wave activity seems to be a permanent feature of the deep equatorial Pacific fields.

In light of the long time scale of the narrow band signals, the broad band character of the energetic background and the opposing directions of phase and group velocity for equatorial waves, extremely careful judgement is needed when inferring vertical propagation from the data. Existence of more than one dominant vertical scale, each behaving differently in time as doc-
FIG. 13. Coherence between $\mu$ and $\xi$ at the equator. Positive (negative) phase means $\mu$ leads (lags) $\xi$ with depth.

umented here, complicates an already difficult task. It is perhaps not surprising that systematic and coherent vertical shifts of the generically called deep jets have been so elusive. Without settling the propagation issue (and that will require longer records), both wave and steady models, or wave-mean flow models, remain potentially important in the quest to understand the basic dynamics governing these deep flows.

The clear need for improved data sets in the future is one of the outcomes of this discussion. Collection of basinwide equatorial transects of deep zonal velocity and intensive observations along the western or eastern boundaries would shed some light on whether the interior flows can be traced to deep energy sources located at the oceanic side walls (Ponte 1989). More observations are necessary to resolve the mean and time dependent components of the deep circulation. This knowledge is crucial in studies of instability or wave-mean flow interaction phenomena which may be relevant to the existence of energetic motions at particular scales like the ones found in the records (e.g., McCready and Lukas 1986; Ponte 1988b). Only with a considerably improved overall picture of the deep equatorial circulation will some of the issues raised here be answered more decisively.

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