Circulation and Low-Frequency Variability near the Chilean Coast: Remotely Forced Fluctuations during the 1991–92 El Niño

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ABSTRACT

Results are reported from the first long, recording current meter observations over the slope off Chile. These observations, at 30°S during the 1991–92 El Niño event, are analyzed together with observations of currents at a local deep sea site; local wind and sea level; sea level from the Peru and Chile coasts; and wind, temperature, and currents from the equatorial Pacific. Mean poleward flow of 12 cm s⁻¹ was observed within the Peru–Chile Undercurrent over the slope. Mean flow in the depth range of Antarctic Intermediate Water was not distinguishable from zero in the presence of strong, low-frequency (LF) variability, which dominated slope currents at all depths. The strongest LF fluctuations had periods of about 50 days, but periods of 10 and about 5 days were also observed. Significant, local wind forcing of slope currents was only found in the period band 6–10 days and may be related to coastal-trapped waves in the atmosphere.

Our analysis shows that free, coastal-trapped waves in the ocean, arriving from the north, dominated the LF variability over the shelf and slope off northern and central Chile during the 1991–92 El Niño event. Strong 50-day period fluctuations there started their journey about two months earlier—and 15 000 km farther up the coastal–equatorial waveguide—near the date line in the equatorial Pacific as equatorial Kelvin waves forced by westerly wind events of similar period. Upon reaching the South American coast, these waves forced coastal-trapped waves, which propagated along the Peru coast into the study region. Likewise, a scenario of equatorial-trapped waves forcing coastal-trapped waves may explain 10-day as well as 6-day and 4.5-day period coastal-trapped waves off Chile stemming from mixed Rossby–gravity and inertia–gravity waves trapped at the equator. Since the large, 50-day period, coastal-trapped waves may strongly modify coastal upwelling source water, such remotely forced waves may have a significant influence on the pelagic ecosystem off Chile, at least during El Niño events.

1. Introduction

The Peru–Chile Current, the eastern boundary current of the South Pacific Ocean, is the site of strong coastal upwelling and exceptionally high biological productivity (Ryther 1969) but also a site of extreme interannual variability of these phenomena due to El Niño events (Wyrtki 1975). Within the last two decades or so, several intensive studies of the Peru–Chile Current, and of the coastal upwelling system associated with it, have been carried out off Peru during "normal" and El Niño conditions. Results from these studies have confirmed the existence of a persistent, poleward undercurrent—the Peru–Chile Undercurrent (Wooster and Gilmartin 1961; Brockmann et al. 1980)—and have revealed poleward propagating, low-frequency (LF) fluctuations of currents and sea level (Smith 1978; Romea and Smith 1983; Brink et al. 1983). Furthermore, these results show significant changes in hydrography and currents and in the strength and nature of these fluctuations during El Niño events (Huyer et al. 1987; Huyer et al. 1991). Lowfrequency fluctuations off Peru appear to be mainly manifestations of free coastal-trapped waves (Brink et al. 1978; Brink 1982) propagating to the south, possibly forced by equatorial waves impinging on the west coast of South America (Clarke 1983; Enfield et al. 1987).

The Peru–Chile Undercurrent between 5° and 15°S off Peru flows poleward with speeds near 10 cm s⁻¹ centered at about 150-m depth over the outer shelf/inner slope and its transport there has been estimated to be

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about 1 Sv (Huyer et al. 1991). Although LF variability of the type discussed below significantly modulates the flow in this current, the Peru–Chile Undercurrent remains quite persistent over El Niño events (Huyer et al. 1991). Some of the present authors have recently reported observations of mean poleward flow of about 3 cm s⁻¹ within the undercurrent at a deep ocean site 150 km off the coast of Chile (30°S) during the 1991–92 El Niño event (Shaffer et al. 1995, to be referred to here as CH1). Direct observations of the Peru–Chile Undercurrent over the shelf and slope south of 15°S have been lacking [except for a few current profiles, Johnson et al. (1980)], but geostrophic calculations and property distributions indicate that the undercurrent probably extends well south of 40°S (Silva and Neshyba 1979).

Significant intraseasonal variability of sea level and currents off Peru at periods of about 40-70 days (to be referred to here as "50 day") may derive from eastward propagating, baroclinic, equatorial Kelvin waves impinging upon the South American coast. Such 50-day, equatorial waves are most prominent in the austral spring/summer and during the onset of El Niño events and are forced by eastward wind events of similar periods in the western and central Pacific (Enfield 1987; Kessler et al. 1995). Between September 1991 and May 1992, a train of four such waves was observed at the equator in the central and eastern tropical Pacific (McPhaden 1993). The wind events forcing these waves are apparently associated with the Madden-Julian oscillation, an atmospheric phenomenon of planetary scale (Madden and Julian 1971, 1972). An analysis of sea level records from the west coast of the Americas revealed convincing evidence for the poleward propagation of the 50-day oscillation as a coastal-trapped wave from the equator northward to central California (34°N) and southward to central Peru (12°S, Spillane et al. 1987). A comparable analysis for the west coast of South America south of 12°S has been lacking.

The strongest case for poleward propagating, coastaltrapped waves off Peru has been made for fluctuations at periods of one to two weeks (to be referred to here as "10 day"), observed in shelf and slope currents and sea level in 1976-77 during a moderate El Niño event (Smith 1978; Romea and Smith 1983). Observed poleward phase speeds and cross-shelf current structure calculated from this data agree rather well with predictions from coastal-trapped wave theory (Brink et al. 1978; Brink 1982). Analyses of multiyear sea level data from Ecuador and Peru indicate that amplitudes of the 10-day variability increase by a factor of 3 or more during El Niño events (Cornejo-Rodriguez and Enfield 1987). It has been suggested that these waves are forced mainly by eastward propagating, baroclinic, mixed Rossbygravity waves (or Yanai waves) trapped about the equator (Clarke 1983; Enfield et al. 1987). The fate of the 10-day, free coastal-trapped waves south of 15°S is not known.

In contrast to the situation off Peru, direct observa-

tions of currents have been lacking in the southern half of the Peru–Chile current off Chile. Recording current meter (RCM) observations in the deep sea 150 km off the coast of Chile (station 2 in Fig. 1) during a four to six month period at the onset of the 1991–92 El Niño event were presented in CH1. Neither 10-day nor 50-day variability was observed in these current records at this deep sea location. Near 100-m depth slow intensification of eastward flow occurred throughout the record and was interpreted as seasonal change.

Here we expand our analysis of the Peru-Chile Current system to address the circulation and LF variability near the Chilean coast. As part of the analysis we report moored current meter observations, which are apparently the first such long records on the slope or shelf off the west coast of South America south of 15°S. To aid in the interpretation of these observations and to address the possibility of remote forcing of LF fluctuations along the coast of Chile, we also consider our own CTD and wind data from the study area; sea level and atmospheric pressure data from a number of stations in Peru and Chile; and moored current meter, temperature, and wind observations at the equator in the central and eastern tropical Pacific. Predictions from a simple coastal-trapped wave model (Brink 1982) and from equatorial trapped wave theory (cf. Gill 1982) are compared with properties of the LF fluctuations as derived from the combined dataset.

2. Observations and methods

A recording current meter mooring was deployed at 30°19.0' S and 71°47.0' W (station 8, Fig. 1) on the slope on 4 November 1991 and recovered on 20 April 1992 (period 1, water depth 903 m). The mooring was redeployed six hours later at the same site and was recovered again on 29 November 1992 (period 2, water depth 863 m), more than one year after first deployment. This slope mooring was located about 10 km off the Chilean coast and lies about 70 km shoreward from the axis of the Peru-Chile Trench (about 6200 m deep in the study area). Depths of more than 500 m extend to within a few kilometers of the coast locally. The mooring had three Aanderaa RCM 7 current meters at nominal depths of 253, 514, and 777 m during period 1 and 213, 474, and 737 m during period 2. The deepest meter was nominally 126 m above bottom in both cases. All RCMs had speed, direction, and temperature sensors and the shallowest one had a pressure sensor. Good quality data were acquired with all current meters for each of the two periods except for the shallowest current meter for which all channels failed during period 2. Due to the difference in nominal current meter depth of about 40 m between the two periods, the RCM temperature records are only considered separately for each period. However, the current records from both periods at the deepest two RCMs have been merged to produce long records to which mean depths of 494 and 757 m have



FIG. 1. The study area and locations of moored current meter (open circles), sea level (large filled circles), wind (circle with cross), and CTD (dots) observations considered in this paper (wind at 170°W on the equator not shown).

been assigned. These choices are motivated by the relatively weak vertical shears of observed currents compared with rather strong vertical gradients in observed temperature (see below).

No attempt was made to correct the mooring data for "knockdown" although vertical displacements of up to 60 m, associated with strong current events, were observed by the pressure sensor of the shallowest RCM. Comparison of temperature from the RCMs with temperature profiles from CTD casts made adjacent to the mooring during the study suggested that vertical displacement amplitudes of the meter near 500 m were perhaps a factor of 3 less than those at the shallowest RCM and that the one near 750 m experienced only small displacements, probably less than 10 m. Based on observed vertical gradients, we expect that these knockdowns may significantly bias the RCM temperature data-our subsequent analysis relies little on this databut that associated bias in observed currents may not affect our conclusions significantly.

Spectra were calculated on hourly RCM data by the

fast Fourier transform. Before the spectra were computed, the mean and trend were removed and a Hanning taper was applied. Rotary energy spectra were calculated according to the method of Mooers (1973). The original hourly data were also low-passed filtered with a 121-point cosine–Lanczos filter with half-amplitude point at 0.016 c h⁻¹ (60 h) as in CH1.

During the study, a hydrographic section extending about 180 km from the coast (stations 1 through 8 in Fig. 1) was occupied during each of the following periods: 2–4 November 1991 and 6–8 February, 22–26 April, and 22–25 June 1992. CTD casts were made to 3000 m at stations 2 and 4 and to 1000 m otherwise. Salinity data from the November and February sections were corrected as described in CH1, where geostrophic profiles calculated from casts from stations 2 and 4 for the first three of these sections were also presented.

Wind speed and direction were measured on an automatic weather station at the coast at Lengua de Vaca Point immediately east of the mooring site (Fig. 1). Data gaps during the periods 17–20 April, 27 April–2 July,

and 30 July–26 September 1992 were filled by data from another automatic weather station at Punta de Toro, 60 km to the south at the coast (comparison between these two wind records during periods of common recording showed good agreement). Wind stress was calculated from this data using a drag coefficient of 1.3×10^{-3} and an air density of 1.2 kg m⁻³, and the results were low-passed filtered as for the currents above. Current, hydrography, and wind observations were made as part of an ongoing international study of the Peru–Chile Current system in association with the Joint Global Ocean Flux Study.

Sea level data from the study period were obtained from tide gauges at Lobos de Afuera and Callao (Peru) and Arica, Caldera, Valparaiso, and Coquimbo (Chile, Fig 1). Sea level data were obtained by ftp from the TOGA Sea Level Center, University of Hawaii (courtesy of G. Mitchum). These series have been subject to quality control by the Center (including compensation for change in zero level, potentially important in this seismically active region). These records contained no large gaps (days or more) during the study period; shorter gaps were interpolated using residuals and the predicted tides. The Coquimbo sea level series was obtained directly from Servicio Hidrografico y Oceanografico de la Armada, Valparaiso, Chile. It contained gaps over 3% of its record during the study period; the longest gap was over six days long. We filled the few multiday gaps using data from Caldera and Valparaiso; the shorter gaps were filled as above.

Atmospheric pressure data were obtained at Arica, Antofagasta (23°38' S at the coast), La Serena (near Coquimbo), and Santo Domingo (near Valparaiso) airports from Direccion Meteorologica de Chile. All series of sea level (hourly) and atmospheric pressure (three hourly) were low-passed filtered as above. Atmospheric pressure was added to the sea level records (scale factor 1 cm/mb) to form adjusted sea level (ASL), a measure of total subsurface pressure. Caldera sea level was adjusted with interpolated air pressure from Antofagasta and La Serena. The sea level data at the Peruvian stations were not adjusted but atmospheric pressure fluctuations there are weak and poorly correlated with sea level variations (Spillane et al. 1987; Huyer et al. 1991). For brevity, all sea level records used below, whether adjusted or not, will be referred to as ASL.

Series of daily means of equatorial winds (170°W) and equatorial temperatures and currents (110°W, observation depths: 1, 10, 25, 45, 60, 80, 100, 120, 140, and 200 m for temperature and 10, 25, 45, 80, 120, and 200 m for currents) from the tropical Pacific Ocean for the study period were obtained by ftp from the TOGA TAO Project Office, NOAA/Pacific Marine Environmental Laboratory, Seattle, Washington (courtesy of M. McPhaden). A description of the TOGA TOA sampling array is given by McPhaden (1993). These daily mean data were low-passed filtered as for currents described above.

3. Current characteristics and circulation

Based on the autospectra of the clockwise and anticlockwise kinetic energy density for the three RCMs at the slope mooring (Fig. 2), energy in the inertial band (1-day period at 30° latitude) decreases by about a factor of 10 and becomes less polarized from the shallowest to the deepest meter. Root-mean-square amplitudes in the inertial band were 10.7, 6.4, and 4.4 cm s⁻¹ at 253, 494, and 757 m at the slope site compared to 18.0 and 5.1 cm s⁻¹ at 100 and 3400 m, respectively, at the deep ocean site (CH1). The ratio of anticlockwise to clockwise energy ranged from 1.5 to 7 at the slope site and from 20 to 60 at the deep sea site. These results show the tendency for suppression of rotary motions near the ocean boundary. In contrast, energy in the semidiurnal band was greatest near the bottom at the slope site. Rootmean-square amplitudes in this band were 6.3, 10.1, and 11.0 cm s⁻¹ at 253, 494, and 757 m on the slope and 5.8 and 4.1 cm s⁻¹ at 100 and 3400 m, respectively, in the deep ocean. The results are consistent with the generation of baroclinic tides on the slope.

Figure 2 shows that current fluctuations over the slope are characterized by strong LF variability. In the long records at 494 and 757 m, which allow better spectral resolution, spectral peaks/plateaus emerge at periods near 5, 10, and 40 days. The record at 253 m, too short to permit resolution of the longest of these periods, is also consistent with enhanced current variability near the 5-day and 10-day periods. As shown in Fig. 2, anticlockwise rotation tends to exceed clockwise rotation at low frequencies, in particular at the shallowest RCM.

Means and variances of temperature and current as well as covariances and eddy kinetic energy for slope station current, all from the complete, low-passed RCM records, are shown in Table 1. Cross-shore and along-shore current components, u and v, are taken to be positive toward 105° and 015°, respectively. This orientation was chosen from the principal axes calculated from the low-passed current records (019.5°, 017.5°, and 013.9° for the complete current records at 253, 494, and 757 m, respectively) and to agree with the alignment of the hydrographic section (105°–295°, Fig. 1) used, for instance, for the geostrophic calculations in CH1.

Mean poleward flow of 12.0 cm s⁻¹ was observed during period 1 at 253-m depth over the slope. This flow is within the core of the Peru–Chile Undercurrent, identifiable in our February 1992 hydrographic section (Figs. 3a,b) by a nearshore, subsurface layer of rather uniform temperature (T~11.0–11.5°C) and, in particular, by a core of high salinity (S>34.65 psu). These properties mark equatorial subsurface water (Wooster and Gilmartin 1961) advected poleward in the undercurrent. When combined, this information leads to a transport estimate of about 1 Sv (Sv \equiv 10⁶ m³ s⁻¹) in the core of the undercurrent in the austral summer 1991/ 92. When the CH1 estimate of poleward flow on the same depth range farther offshore (their Fig. 5) is also



FIG. 2. Clockwise and counterclockwise spectra of kinetic energy density for current meter records from 253 m, 494 m, and 757 m depths at the slope station (8 in Fig. 1 inset). Spectral calculations were made with 10, 20, and 40 degrees of freedom for the frequency ranges 10^{-3} – 10^{-2} cph, 10^{-2} – 10^{-1} cph, and 10^{-1} –0.5 cph, respectively, and for the entire record length of each series (Table 1).

considered, the transport estimate for poleward flow in the undercurrent depth range in this time frame increases to about 1.5 Sv.

In the June 1992 hydrographic section, the nearshore quasi-isothermal layer and high salinity core have disappeared (Figs. 4a,b) but mean, poleward flow is now observed deeper down, 3.6 cm s^{-1} at 474 m during period 2 (Table 1). Earlier observations have shown that salinity values in the saline undercurrent core reach maximum values during summer off central Chile (Fonseca 1985). Between February and June 1992, horizontal density gradients nearshore have relaxed indicating less vertical shear in the geostrophic flow there. Although direct current observations are lacking in the undercurrent core during the austral winter 1992, the above hydrographic signatures imply that poleward flow there was weaker than observed during the preceding summer. Given these results and the work of Huyer and collaborators off Peru (Huyer et al. 1991), we suspect, but cannot prove, that this current and hydrographic signature of undercurrent structure and seasonal variation is rather representative for average conditions off central Chile despite the 1991-92 El Niño.

Observed mean alongshore flow below the core of the undercurrent is weak, typically a factor of 10 less than the rms velocities associated with LF variability (Table 1). Mean equatorward flow of 2.2 cm s⁻¹ at 777 m during period 1 may be compared with the estimate in CH1 of about 1 cm s^{-1} in the same period and depth range but 140 km farther seaward. It was suggested in CH1 that this deep equatorward current may be part of the eastern boundary current system off the west coast of South America and may be a significant conduit of Antarctic Intermediate Water from its formation region in the extreme southeastern Pacific to elsewhere in the Pacific. However, during period 2, weak poleward flow was observed at 737-m depth and the average over both periods was zero (Table 1). Given the strong LF current variability observed at our slope station, longer records will be needed there to resolve weak mean flows reliably and to resolve the issue of a deep equatorial undercurrent within the Peru-Chile Current System.

4. Low-frequency fluctuations: Local forcing

Low-pass time series of local alongshore wind stress, adjusted sea level, and temperature and currents from 253, 494, and 757 m at the slope station are shown in Fig. 5. Local winds are most favorable for coastal upwelling during the austral spring and summer and exhibit only a few northerly wind reversals in the austral fall and winter. This yearly cycle is strongly modulated by synoptic-scale wind events. This is confirmed by the spectrum of this wind stress record (Fig. 6) showing that energy in these events peaks at a period between 6 and 10 days. Much of this synoptic-scale wind variability is probably associated with "coastal lows," coastal-trapped waves in the atmosphere (cf. Gill 1982).

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TABLE 1. Means and variances of low-passed temperature and current component records as well as eddy kinetic energies ($K_e = (1/2)(u'^2 + v'^2)$) observed at the slope station (8 in Fig. 1 inset), calculated for the instrument depths and observation periods indicated in the table. The half-amplitude point of the low-pass filter used was 60 h. Positive *u* and *v* are toward 105° and 015°, respectively, and V_{max} is the maximum speed observed in the original records.

Depth (m)	<i>Т</i> (°С)	\overline{u} (cm s ⁻¹)	\overline{v} (cm s ⁻¹)	$ V _{\max}$ (cm s ⁻¹)	T^2 ((°C) ²)	$\overline{u}^{\prime 2}$ (cm ² s ⁻²)	$\overline{v}^{\prime 2}$ (cm ² s ⁻²)	u' v' (cm ² s ⁻²)	$\frac{K_e}{(\mathrm{cm}^2 \mathrm{~s}^{-2})}$	
4 Nov 1991–20 Apr 1992										
253	10.85	1.4	-12.0	75.8	0.28	55.3	457.9	37.3	256.6	
514	7.10	0.7	1.1	43.5	0.26	4.7	185.4	7.4	95.0	
777	4.79	-0.3	2.2	44.7	0.07	1.8	161.3	-3.1	81.6	
20 Apr-29 Nov 1992										
474	6.94	-0.5	-3.6	33.4	0.16	4.5	79.6	5.0	42.1	
737	4.78	-0.3	-1.7	43.8	0.04	1.9	86.2	-4.1	44.1	
4 Nov 1991–29 Nov 1992										
494		0.0	-1.6	43.5		5.0	129.6	7.5	67.3	
757	_	-0.3	0.0	44.7	—	1.9	121.5	-3.7	61.7	

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Off Chile, these waves propagate southward with phase speeds of about 10 m s⁻¹ and extend seaward about 150 km. Their influence extends along this coast from 20° – 25° S to about 40°S (Rutllant and Garreaud 1995).



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Adjusted sea level exhibits a yearly cycle modulated by 50-day fluctuations, particularly during the first 8–9 $0 \frac{1}{2} \frac{2}{3} \frac{3}{4} \frac{678}{678}$



FIG. 3. Vertical sections of temperature, salinity, and sigma-t based on CTD casts made during the period 6–8 Feb 1992 along the hydrographic line shown in the Fig. 1 inset. Station numbers are indicated in both figures. The large dots show the positions of the recording current meters during the period.

FIG. 4. Vertical sections as in Fig. 3 but based on CTD casts during the period 22–25 Jun 1992.



FIG. 5. Low-passed time series of alongshore wind stress (positive toward 15°) and sea level (adjusted by air pressure) at the coast near 30°S and temperature and vector current (upward \rightarrow 15°) at 233, 494, and 757 m at the slope station. The half-amplitude point of the low-pass filter was 60 h. Solid arrows on the time axis mark the periods of the two hydrographic sections shown in Figs. 3 and 4; the dashed arrow marks a "front" discussed in the text.

months of the record, and by shorter period fluctuations. The ASL spectrum (Fig. 6) shows enhanced energy in the band 4–7 days and for periods of two weeks or more. Significant coherence (at 80% confidence level, not shown) was found between currents at all depths at our slope mooring and local ASL for periods of several days or more. The corresponding phase was usually near 180°, as would be expected from coastal-trapped ASL fluctuations and geostrophy.

The vector representation of the low-passed currents in Fig. 5 shows vividly that the 50-day oscillations dominated the alongshore flow on the slope off the central Chilean coast during spring and summer 1991–92. Also, evident are the 10-day oscillations, most visible in the



FIG. 6. Autospectra of the low-passed local alongshore wind stress and adjusted sea level series shown in Fig. 5. Spectral calculations were made with 10 degrees of freedom.

757-m record during winter. A sharp transition to the highest observed ASL and very strong poleward flow occurred during 2–4 March (marked by the dashed arrow on the time axis). The temperature records from the RCMs also showed 50-day and shorter period fluctuations. However, mooring knockdown during events of strong currents, like the one in March, is certainly responsible for some of the observed temperature variations.

Cross-spectral calculations between the alongshore wind stress and alongshore currents from the three RCMs of the slope mooring are shown in Fig. 7. Coherence of local currents with local alongshore wind stress appeared to decrease with depth in general. At periods close to 5 days, at which considerable wind and current energy resided (Figs. 2 and 6), both cross-slope (not shown) and alongshore currents at 253-m depth were coherent with the local alongshore wind stress. In this band, alongshore flow tended to be in phase with the wind (Fig. 7b), suggesting that the 5-day current fluctuations in the upper water column at the slope station were largely forced by the local wind. As discussed above, the wind in this band is strongly associated with passing "coastal lows" in the study region.

Currents and local wind stress were also coherent in the 10-day band but alongshore flow at 253-m depth tends to lag the wind stress by 3–4 days, much longer than the local inertial period of 1 day. This phase lag is more consistent with "upstream" rather than local wind forcing: Wind events to the north of the study area may force perturbations that propagate southward (with coastal-trapped wave dynamics, see below) much slower than the wind events themselves. Those current fluctuations in the 10-day band that are coherent with local winds tend to have baroclinic structure (Fig. 7b). At 253-m depth in both 5-day and 10-day bands, crossslope flow tends to lead alongshore flow by about 90° (not shown), in agreement with observed anticlockwise rotation (Fig. 2).

5. Low-frequency fluctuations: Remote forcing

a. Local properties of coastal-trapped waves off Chile

As shown above, much of the LF variability in the study region did not appear to be related to local forcing. Here we investigate the possibility that this variability may be associated with coastal-trapped waves, shown to be important farther to the north off Peru (Romea and Smith 1983). Such waves are dynamical hybrids between baroclinic gravity waves trapped near a coast (baroclinic Kelvin waves) and barotropic vorticity waves trapped over the shelf and slope [barotropic shelf waves (Brink 1991)]. As the internal Rossby radius of deformation becomes large (small) compared to the shelf-slope topographic scale, coastal-trapped waves become more like Kelvin (shelf) waves. Since the Rossby radius is proportional to f^{-1} , where f is the Coriolis parameter, coastal-trapped waves tend to be more Kelvin-wave-like toward the equator and more shelf-wavelike toward high latitudes.

Figure 8 shows the distribution of alongshore flow associated with the first mode coastal-trapped wave (CTW) for our study region off Chile. Calculations followed the method of Brink (1982) using the computer programs of Brink and Chapman (1987). The bottom topography, Fig. 8, is an alongshore mean between 29° and 31°S from available depth data, and mode calculations used a mean vertical profile of N^2 based on data from the February 1992 (austral summer) hydrographic survey (Fig. 3c). An offshore depth of 4500 m was chosen for the calculation, shallower than the local depth of the Peru-Chile Trench (about 6000 m) but deeper than the abyssal plain farther offshore (about 4200 m, Fig. 1). The simple Brink (1982) model only permits monotonic depth changes, but the mode structure (Fig. 8), trapped quite close to the coast, would probably not change much in the presence of a trench in a more complex model. The current amplitudes have been scaled using the first EOF mode amplitudes (Table 2) to permit direct comparison with observed current speeds. Locations of our recording current meters with respect to the alongshore averaged topography are indicated by heavy dots.

The mixed baroclinic–barotropic structure of the first CTW mode reflects the importance of both stratification and topography for CTW dynamics off central Chile. The internal radius of deformation is 32 km based on the mean N^2 profile and the offshore slope scale is about



FIG. 7. Coherence (a) and phase (b) between local alongshore wind stress and alongshore current at 253 m (solid line, circles), 494 m (long-dashed line, triangles), and 757 m (short-dashed line, squares) of the slope station. The calculations were made for the longest common record length (Table 1) with ten degrees of freedom. Only phase results for coherence squared values above the 80% confidence level are plotted.

67 km (offshore position of 4500-m depth). For the firstmode long CTW, the poleward phase and group velocities are 2.78 m s⁻¹. For "long" periods of interest here, the local CTW is rather nondispersive. For instance, for wave periods of 10 and 5 days, the poleward phase (group) velocity decreases slightly to 2.73 m s⁻¹ (2.64 m s⁻¹) and 2.65 m s⁻¹ (2.47 m s⁻¹), respectively. Phase and group velocities for the long CTW are significantly faster than for either the first-mode internal (Kelvin) wave, 2.31 m s⁻¹, or first-mode long barotropic shelf wave, 1.37 m s⁻¹, all calculated for local stratification and topography. The second-mode long CTW (not



FIG. 8. Alongshore velocity structure of the first baroclinic mode CTW calculated from the model of Brink (1982) using mean local topography and a mean density profile based on Fig. 3c data. Velocity (in cm s⁻¹) has been scaled for comparison using the first EOF mode amplitudes computed from the moored current data (Table 2). Heavy dots indicate the positions of recording current meters relative to the mean topography. For this mode, the long-wave phase speed is 2.78 m s⁻¹.

shown) is more baroclinic with a zero crossover near 900 m on the slope and a phase speed of 1.38 m s^{-1} . For conditions near 15° S off the Peru coast, Brink (1982) found long CTW phase speeds for the first two modes of 2.28 and 1.25 m s⁻¹. A comparison with Brink's results show that lower, calculated phase speeds off Peru can be attributed mainly to the effect of a smaller Coriolis parameter there but also to Brink's choice of 4000 m, as opposed to our choice of 4500 m, for offshore depth.

To test the sensitivity of our results off Chile to seasonal variations of stratification and the choice of the offshore depth, we also calculated CTW modes for austral winter stratification and assumed offshore depths of 4000 and 5000 m. For a mean vertical profile of N² based on data from the June 1992 hydrographic survey (Fig. 4c) and offshore depth of 4500 m, we found a first-mode long CTW phase speed of 2.54 m s⁻¹ showing slower phase propagation due to weaker wintertime stratification (Figs. 3 and 4). For offshore depths of 4000 and 5000 m and the original austral summer profile of

TABLE 2. Amplitudes and percent variance explained for the three low-passed alongshore current records from the slope station by the first two alongshore current, EOF modes based on these records during period 1 (Table 1).

	Mo	de 1	Mode 2		
	(80	.9%)	(16.8%)		
Depth (m)	Percent variance explained	Amplitude (cm s ⁻¹)	Percent variance explained	Amplitude (cm s ⁻¹)	
253	91.1	20.4	8.6	6.3 - 3.9 - 9.0	
514	85.5	12.6	8.3		
777	47.0	8.7	49.8		



FIG. 9. Autospectra of first and second EOF modes of low-passed alongshore current from the three RCM records at the slope station during period 1 (Table 1). Spectral calculations were made with eight degrees of freedom.

 N^2 , we found first-mode long CTW phase speeds of 2.59 and 2.95 m s⁻¹, respectively.

To search for organized motion on the slope, we calculated empirical orthogonal function modes (cf. Davis 1976) for our three LF, alongshore current records during period 1. Table 2 lists percent variance explained and amplitudes for the first two EOF modes and the spectrum of the time series associated with each mode is in Fig. 9. The first EOF mode has a mixed barotropicbaroclinic structure and explains 80.9% of total LF variance (a still larger percentage of local LF variance is explained at 253 and 514 m). The second EOF mode has a baroclinic structure and explains only 16.8% of the variance but about one-half of the variance at 777 m. A comparison of the spectra of both modes indicates that the structures of the 5-day and 50-day period oscillations are well represented by the first EOF mode, which in these bands contained 10 times more energy than the second EOF mode. At a period of 10 days, this ratio decreases to about 2 and for periods between about 6 and 8 days, energy contained in both modes is similar. To test for local forcing of these modes, we calculated cross-spectra with local alongshore wind stress (not shown). Coherence of the local wind stress with the second EOF mode (at 90% confidence level) was found for periods of 7-12 days. As shown above, the relative importance of the second EOF mode is greatest within this period band. Also, this baroclinic mode lagged the wind by 3–4 days, in agreement with the analysis presented above (Fig. 7). The first mode was not coherent (at the 80% level) with the local wind stress.

These results may be compared with CTW predictions at the RCM position points on the slope. Both the model and the data-based first modes show decreasing amplitude with depth, whereas the first EOF mode shows considerably more vertical shear than predicted for the first CTW mode. A similar result was found by Brink (1982) off Peru. The simple CTW model used here neglects mean flow, although the Peru–Chile Undercurrent is a ubiquitious feature of the nearshore zone. Influence of this mean flow structure together with that of bottom friction may help explain observed large vertical shears in the LF variability over the slope.

We also calculated EOF modes from all five, lowpassed alongshore current records (three on the slope, two in the deep ocean) over their common record length (5 Nov 1991-22 Feb 1992). Over the slope, the first EOF mode was essentially identical to the one above, again explaining about 80% of the LF variance there. In contrast, this mode explained essentially none of the variance at the deep ocean site. Thus, as also confirmed by additional coherence calculations that we carried out, LF fluctuations on the slope are unrelated to those in the deep ocean 140 km farther seaward in the study area. The deep ocean site is located seaward of the zone within which significant LF fluctuations associated with coastal-trapped waves are possible. Finally, the motion of fluid particles associated with a coastal-trapped wave in the Southern Hemisphere would be expected to possess anticlockwise rotation due to the tendency to conserve potential vorticity (Gill 1982). The spectra of observed currents on the slope (Fig. 2) indicated rotation of this sense at low frequencies.

b. Propagation in the Peru–Chile coastal waveguide

Any free CTWs arriving in our study region will have passed through the coastal waveguide farther north, modulating coastal sea level in a systematic way. The ASL records for the study period from Lobos, Callao, Arica, Caldera, and Valparaiso, extending about 3500 km along the west coast of South America (Fig. 2), are shown in Fig. 10. The large sea level rise off Peru during late 1991 and early 1992 is an expression of the 1991-92 El Niño event. This event culminates with a large, steep sea level rise in the last week of February. This front can be easily traced southward to Valparaiso where it arrives in the first week of March. This is the same front observed in local sea level and currents in our study area during 2-4 March (Fig. 5). Significant poleward propagating 50-day variability is evident in the record, particularly at its most poleward extension off Chile where these fluctuations can be identified with the 50-day fluctuations observed in our study area (Fig. 5). Energy conservation in a long, free coastal-trapped wave implies increasing amplitude in the poleward di-



FIG. 10. Low-passed records of sea level along the west coast of South America from Lobos de Afuera and Callao (Peru) and Arica, Caldera, and Valparaiso (Chile, data adjusted with air pressure). See Fig. 1 for locations. Zero sea levels are means calculated for the period 1986–1993 and the separation between the different sea level axes is scaled by alongshore distance. The slope of the dotted lines following selected sea level features is based on a poleward phase speed of 266 km day⁻¹ (3.08 m s⁻¹) from the analysis in Fig. 11.

rection (Brink 1989). Such a tendency was also found for the poleward propagating 50-day fluctuations along the west coast of Central and North America south of 23°N (Spillane et al. 1987). Amplitudes were found to decrease greatly poleward of that latitude, probably due to dissipation of these waves within the Gulf of California. The coastal waveguide off western South America is topographically smoother than its northern counterpart, whereas coastal-trapped waves apparently carry more energy farther poleward off South America than off North America.

To help quantify the visual impression of poleward propagation from the ASL records, we calculated lagged correlation between all ASL record pairs from this data (Fig. 11). Based on the lag at maximum correlation plotted as a function of alongshore separation for each of these pairs, the correlations of all pairs were significant to the 99% level except the pairs Caldera–Lobos and Valparaiso–Lobos, which were significant to the 95% and 90% levels, respectively. A linear fit, passing through zero, yields an estimate of the poleward propagation speed of the ASL signals of $266 \pm 4 \text{ km } \text{d}^{-1}$ or $3.08 \pm 0.04 \text{ m } \text{s}^{-1}$. From a similar analysis of ASL data from 1976 to 1977 on the west coast of South America between 2° and 17°S, Romea and Smith (1983) found support for a poleward propagation speed of 240 km d⁻¹ they estimated from coherence analysis. A linear fit to their data (in their Fig. 8) yields a slightly higher speed, about 260 km d⁻¹ or 3 m s⁻¹.

These results indicate poleward propagation speeds near, but slightly higher than, theoretical phase speeds calculated and cited above for CTWs along the west coast of South America. As shown in Fig. 10, by the slanted lines with slope determined from our correlation analysis, individual features with timescales ranging from 10 to 50 days all seem to move poleward at similar



FIG. 11. Lag of maximum correlation versus alongshore distance for all pairs of sea level series in Fig. 10. Circles, squares, and triangles indicate correlations significant at 99%, 95%, and 90% levels, respectively. Also shown is the least-mean-square linear fit (forced to pass through zero) to these points. The inverse of the line slope gives a poleward propagation velocity of 266 km day⁻¹ (3.08 m s⁻¹).

speeds as expected for nondispersive CTWs (the large amplitude, long period fluctuations seem to move slightly faster, however). As indicated above, a choice of perhaps 5000 m for offshore depth for the Peru and Chile CTW calculations, a choice motivated by the presence of the Peru–Chile Trench, would lead to still closer agreement between theory and observations. In contrast to our results of the observed poleward phase propagation of 50-day period, ASL fluctuations off the coast of Central and North America appear to be slower than that expected for long CTWs in this region [~150 km d⁻¹ compared to ~190 km d⁻¹ (Spillane et al. 1987)].

To examine more closely the interrelationships among coastal ocean data at different alongshore positions off Chile in the frequency bands of interest, we made crossspectrum calculations between ASL at Arica and the three LF alongshore current records on the slope off Coquimbo (Fig. 12). High coherence between ASL and alongshore currents is found for all periods greater than about 8 days at depths of 494 and 757 m on the slope. Significant coherence of Arica ASL with the current record at 757 m is also found at periods near 6.0 and 4.6 days. If these high coherences were associated with poleward propagating long CTWs, phase lags of alongshore currents off Coquimbo after Arica ASL should increase linearly with frequency. The results in the phase plot (Fig. 12b) show this behavior. The slanted lines in the figure are the least mean-square best linear fit (forced to pass through 180° at zero frequency since ASL and alongshore currents are 180° out of phase in the simple CTW model) to all plotted phase results for periods



FIG. 12. Coherence (a) and phase (b) between low-passed records of adjusted sea level at Arica and alongshore current at 253 m (solid line, circles), 494 m (long-dashed line, triangles), and 757 m (short-dashed line, squares) of the slope station. The calculations were made for the longest common record length (Table 1) with ten degrees of freedom. Only phase results for coherence squared values above the 80% confidence level are plotted. Also shown in the phase plot is the least-mean-square linear fit (forced to pass through 180° at zero frequency, see text) to the plotted phase results at periods greater than 4.5 days. For the Arica–slope station alongshore separation of 1360 km, the slope of the linear fit yields a poleward phase speed of 250 km day⁻¹ (2.89 m s⁻¹).

greater than 4.5 days. Phase results for the 10-day and 50-day periods as well as for the 4.6-day and 6-day periods fall squarely on the best-fit curve. For the Arica–Coquimbo slope station alongshore separation of 1360 km, the slope of the best fit implies a poleward phase speed of 250 km d⁻¹ or 2.89 m s⁻¹, in good agreement with the first-mode long CTW phase speed of 2.78 m s⁻¹ calculated above. Arica ASL was coherent with

alongshore currents at 494 m at periods of 4.1 and 3.5 days, but with phase relationships inconsistent with CTW theory. Coherence between Arica ASL and the LF alongshore currents at 253 m over the slope was generally quite low with significant coherence only in the 10-day band. This may be explained by local forcing since the highest coherence between local wind and alongshore flow at 253 m (Fig. 7) is in the same band as the lowest coherence between Arica ASL and this flow. Recall that the energetic 50-day band cannot be resolved for the relatively short current record at 253 m.

Although improved models including wind forcing, mean currents, and friction should be considered in future work on LF variability off Chile, it can already be concluded from the above analysis and earlier work that free CTWs dominated the low-frequency variability on the shelf and slope off the west coast of South America at least from as far north as Lobos de Afuera to as far south as Valparaiso during the 1991–92 El Niño event.

c. Equatorial origin of the low-frequency fluctuations

Here we look to the equatorial waveguide of the tropical Pacific Ocean as a possible source of the free CTWs observed along the west coast of South America during the 1991–92 El Niño event. Solutions for equatorialtrapped waves can be derived for an equatorial β plane [$\beta = \partial f / \partial y$, where y is the north–south coordinate, cf. Gill (1982)]. One special solution is the equatorial Kelvin wave with zero meridional flow and the simple dispersion relation,

$$\omega - kc = 0, \tag{1}$$

where ω is the frequency, k is the zonal wavenumber, and c is the gravity wave speed. Equatorial Kelvin waves propagate eastward with maximum pressure and zonal current amplitudes on the equator. These amplitudes are in phase and decay poleward with the equatorial radius of deformation, $R = (2c/\beta)^{1/2}$. Solutions for the remaining classes of equatorial-trapped waves, also trapped on the scale R, follow the dispersion relation

$$\omega^2 - k^2 c^2 - \beta k c^2 \omega^{-1} - (2n+1)\beta c = 0, \quad (2)$$

where $n (=0,1,2,\cdots)$. The dispersion curves corresponding to Eqs. (1) and (2) are in Fig. 13.

The remaining classes of equatorial-trapped waves are inertia–gravity (IG) waves (upper curves in Fig. 13, $n=1,2,\cdots$), Rossby waves (lower left curves in Fig. 13, $n=1,2,\cdots$) and a mixed Rossby–gravity (MRG) wave (n=0). Even-numbered modes have (maximum) meridional velocity fluctuations but no pressure fluctuations at the equator, whereas odd numbered modes have (maximum) pressure fluctuations but no meridional velocity fluctuations there. These properties will be useful in the analysis below.

All equatorial-trapped waves exhibit eastward energy propagation (positive slope of the dispersion curves in Fig. 13) for some frequency–wavenumber combinations



(energy in Kelvin and MRG waves can only propagate eastward). When these waves reach the eastern boundary at the coast of South America, some energy may be reflected as westward propagating Rossby or IG waves and the remaining energy is passed to CTWs propagating poleward into the Northern and Southern Hemispheres (Moore 1968; Clarke 1983). However, within a range of intermediate wave periods—between 5.8 and 32 days for the dimensional choices in Fig. 13 neither Rossby nor IG waves are possible and wave energy reaching the eastern boundary will be deflected poleward as coastal-trapped waves.

A series of recent papers have shown that the 1991– 92 El Niño event started with a series of four equatorial Kelvin waves of about 50-day period in the central tropical Pacific (McPhaden 1993; Kessler et al. 1995; Kessler and McPhaden 1995). These waves were forced by westerly wind events of similar period in the western and central tropical Pacific and propagated eastward with the first baroclinic mode phase speed of 2.4 m s⁻¹. Each subsequent wind event advanced farther eastward than its predecessor, a behavior which suggests ocean– atmosphere feedbacks (Kessler et al. 1995).

The upper two curves in Fig. 14 show the zonal wind speed observed on the equator at 170°W and the depth





FIG. 14. Low-passed records of zonal wind velocity on the equator at 170° W, the depth of the 20°C isotherm, and the meridional current component at 45-m depth on the equator at 110° W and demeaned and detrended, alongshore current at the slope station (solid long-dashed and short-dashed lines refer to the currents at 253 m, 494 m, and 757 m, respectively). From top to bottom, arrows in the plots identify two westerly wind events, two equatorial-trapped Kelvin waves forced by the wind events, and two coastal-trapped waves at 30° S forced by the Kelvin waves impinging on the South American coast.

of the 20°C isotherm on the equator at 110°W for August 1991 to November 1992. The isotherm depth curve is based on linear interpolation from temperature observations at fixed depths. This marker of thermocline displacements on the equator has been used in the above papers to trace the eastward propagation of the baroclinic equatorial Kelvin waves across the tropical Pacific. The four wind events and the Kelvin waves forced by them are apparent in these curves although the first wind event (September–October 1991) still had not advanced east of the date line and the last downwelling Kelvin wave (April 1992) is partially masked by shoaling of the thermocline between March and June 1992. The two strongest downwelling Kelvin waves revealed in the 20°C isotherm depth are marked by arrows (Fig. 14), as are the two wind events that forced them. The The lower curve in Fig. 14 shows the detrended and demeaned, low-passed alongshore current observations from our slope station off Coquimbo. A comparison of the curves presented in Figs. 14 and 10 strongly suggests that the dominant 50-day period oscillations observed in sea level and alongshore currents off the coast of Chile during 1991–92 have their origins in the wind-driven Kelvin waves of the equatorial Pacific. Upon encountering the South American coast, these Kelvin waves force coastal-trapped waves, which propagate the signal into our study area. Above we presented strong evidence for this CTW connection.

Indeed, the two major 50-day events observed in alongshore flow over the slope off Chile (two arrows in last curve of Fig. 14) can be directly related to the two major Kelvin wave events at the equator [two arrows in the second curve; thermocline depression (sea level rise) accompanies eastward flow in the upper ocean on the equator for the first baroclinic, equatorial Kelvin wave, while depression (rise) accompanies southward flow in the upper ocean off Chile for the first baroclinic CTW]. The front (discussed in connection with Fig. 10 and in the text above as a marker of the abrupt onset of the second of these events) can be easily followed back along the equator. The date of the passage of this front at 110°W on the equator (5 Feb), the date of its arrival at Lobos (21 Feb), and the separation between these two locations of about 3600 km lead to an estimate for eastward propagation speed of the Kelvin wave in the eastern tropical Pacific of 2.5 m s⁻¹ similar to the phase speed estimate quoted above. Based on crossspectrum calculations between the depth of the 20°C isotherm on the equator at 110°W and Arica ASL (Fig. 15), significant coherence (and a phase relationship consistent with the above interpretation) was found for the lowest frequency spectral estimate (78-day period).

Clarke (1983) suggested that coastal-trapped waves of 10-day period observed off Peru (cf. Romea and Smith 1983) may have been forced by MRG waves at the equator impinging on the west coast of South America. Enfield et al. (1987) found considerable observational evidence to support this. Wunsch and Gill (1976) noted that large zonal-scale winds along the equator may excite such waves near the zero wavenumber crossover of the dispersion relation. Indeed, for our first baroclinic mode wave speed at the equator of 2.4 m s⁻¹, MRG waves at this crossover have a period of 10 days (Fig. 13). Since no westward propagating, equatorial-trapped waves exist at this period, all incoming MRG wave energy will continue poleward in the form of CTWs.

The third curve in Fig. 14 shows the meridional current component observed at 45-m depth on the equator at 110°W from August 1991 to November 1992. As discussed above, this current component on the equator should sample even modes, including the MRG wave, of the set of equatorial-trapped waves (Fig. 13). This record begins and ends with large fluctuations of about 20-day period to which we will return in section 6. At the start of the El Niño event in September 1991, the 20-day oscillations give way to longer period perturbations probably associated with the passage of the first Kelvin waves: Although Kelvin waves have zero meridional velocity in the simple model presented above, interaction with strong mean currents, like the equatorial undercurrent, at the equator may modify the ideal mode structure of the equatorial waves. From about January 1992, fluctuations of 10–15 day period became more apparent and grew in amplitude. These may be the MRG waves near the dispersion curve crossover. In July–August 1992 the 20-day fluctuations dominated again.

As a test of the scenario of MRG waves forcing coastal-trapped waves along the west coast of South America, we made cross-spectrum calculations between the meridional current component observed at 45-m depth on the equator at 110°W and Arica ASL. Significant coherence was found in the 10-day band (Fig. 15b), the MRG wave crossover period. Furthermore, the phase relationship in this band (Fig. 15d) supports this scenario: Near the zero crossover in the dispersion relation, MRG waves become essentially standing waves such that fluctuations associated with them would occur simultaneously at 110°W and at the South American coast near the equator. For the MRG wave, thermocline depression (sea level rise) south of the equator would lead meridional velocity at the equator by 90°. At the coast such a depression (rise) would propagate poleward as a CTW. Together with the alongshore distance of 2300 km from the point that such a depression (rise) would intersect the South American coast (about 2°S) to Arica and with the CTW phase speed off Peru of 240 km d⁻¹ [Romea and Smith (1983), close to our value of 250 km d⁻¹ off Chile], these properties predict the phase lines drawn in Fig. 15d. As shown, the observed and predicted phase differences agree well in the 10-day band. These results suggest that much of the significant CTW energy observed in the 10-day band off Chile during 1991-92 may derive from wind-driven, mixed Rossby-gravity waves impinging on the coast of South America at the equator.

Wunsch and Gill (1976) found evidence in sea level records for the western and central tropical Pacific of the several meridional modes in first baroclinic mode, IG waves trapped at the equator. Here we also consider the possibility that such waves may force CTWs that may be observable off Chile. Wunsch and Gill argued that either IG waves of zero group velocity (energy accumulation) or zero zonal wavenumber (large zonalscale wind forcing as for the MRG scenario above) would be excited. We prefer the second possibility for our scenario: Zero group velocity IG waves (wavelengths around 5000 km, Fig. 13) would tend to remain in their region of generation and would, in any case, be expected to reflect at the eastern boundary (Clarke



FIG. 15. Coherence and phase between low-passed records of adjusted sea level at Arica and 1) the depth of the 20° C isotherm (a,c) and 2) the meridional current component at 45-m depth (b,d), both from the equator at 110° W. The calculations were made for the period Nov 1991–Nov 1992 with ten degrees of freedom. Only phase results for coherence squared values above the 80% confidence level are plotted. The lines in the phase plots are phase predictions from the scenario of equatorial-trapped waves forcing coastal-trapped waves described in the text.

1983). On the other hand, IG waves at the dispersion curve crossover would be felt at the eastern boundary and for the slightly higher frequencies of the lowest of these modes (Fig. 13), most of the incoming wave energy could deflect poleward as coastal-trapped waves (Clarke 1983). For k = 0, Eq. (2) yields

$$\omega = [(2n + 1)\beta c]^{1/2}, \qquad (3)$$

which for c = 2.4 m s⁻¹ leads to wave periods of 10.0 days for n = 0 (MRG wave) and 5.7, 4.4, 3.7, and 3.3 days for n = 1, 2, 3, and 4 (IG waves).

The well-defined peaks of coherence between Arica ASL and alongshore current records at 757 m on the

slope off Coquimbo (Fig. 12) at periods near 6.0 and 4.6 days as well as phase relationships at these periods are consistent with CTW propagation. These periods correspond quite well with those of the first two IG wave modes. From the above theory and scenario, odd IG modes would show up in cross-spectra between Arica ASL and the depth of the 20°C isotherm on the equator (Figs. 15a,c), and even IG modes would show up in such spectra between Arica ASL and the meridional component of current on the equator (Figs. 15b,d). Two well-defined coherence peaks near 6.5 and 4.3 days are found for the first combination (Fig. 15a), and a coherence peak near 3.5 days is found for the second one

(Fig. 15b). Since the scenario we consider here is similar to that for the MRG waves above, the predicted phase lines in Fig. 15d should also be applicable to the present scenario. However, odd modes have maximum pressure fluctuations at the equator leading to a slightly longer path to Arica (2500 km) and slightly steeper phase lines in Fig. 15c. As shown, the observed and predicted phase differences agree reasonably well for the 6.5 day peak, which may be associated with the first IG wave mode. The phase results for the shorter period coherence peaks

in Fig. 15c. As shown, the observed and predicted phase differences agree reasonably well for the 6.5 day peak, which may be associated with the first IG wave mode. The phase results for the shorter period coherence peaks do not appear to be consistent with the scenario, but the assumption of nondispersive propagation on which the phase lines in Fig. 15 are based will break down for short enough periods. This evidence can only be considered to be suggestive. Strong equatorial currents may significantly modify the dispersion relation of these waves. Still, it appears possible that CTWs in some discreet frequency bands off Chile during 1991 and 1992 may have been forced by the first several meridional modes of equatorial trapped inertia–gravity waves arriving at the coast of South America.

6. Discussion and summary

Moored current meter observations on the slope near 30°S off Coquimbo, Chile, for a period of over one year in 1991-92 during a moderate El Niño event have been analyzed. Our current records from the slope are apparently the first such long records off the west coast of South America south of 15°S and north of high southern latitudes. In comparison with the results of current observations at a deep sea site 140 km farther seaward (Shaffer et al. 1995), we found more semidiurnal-band energy, less inertial-band energy, and much more low frequency energy at the slope site. In fact, LF variability at both sites was found to be unrelated. On the slope, large current fluctuations of about 50-day period with maximum currents approaching 75 cm s⁻¹ dominated the LF variability of the austral spring and summer 1991-92. Current fluctuations of 10-day period were particularly apparent during the austral fall and winter 1992. In addition, we found stronger mean poleward flow at the slope site within the depth range of the Peru-Chile Undercurrent, about 12 cm s⁻¹ compared to about 3 cm s^{-1} estimated for the deep sea site. Mean flow at greater depths on the slope was too weak to be determined with confidence from this data in the face of strong LF variability.

To determine the nature and source of these large LF fluctuations, we also analyzed our own CTD and wind data as well as sea level and atmospheric pressure data from a number of stations along the coast of Chile. Local winds were characterized by events of 6-day to 10-day periods often associated with "coastal lows," coastal-trapped waves in the atmosphere found along the Chile coast south of about 20°–25°S. We found that flow above 300-m depth on the slope was related to the local wind at periods of about 5 and 10 days but that at greater

depths these relationships degraded. In contrast, we found high coherence between adjusted sea level at Arica (18°28'S) and alongshore flow mainly below 300 m on the slope off Coquimbo at essentially all periods of 5 days or more. The phase relationships between these records clearly showed poleward propagation at these periods with a phase speed of about 2.9 m s⁻¹, determined by empirical fit to the phase results. This result agrees well with phase speeds of 2.55–2.95 m s⁻¹ calculated for the first baroclinic mode CTW from the model of Brink (1982) with local topography and stratification.

These results and an EOF mode analysis based on alongshore current observations on the slope support the following interpretation of the LF fluctuations over the shelf and slope off northern and central Chile: At least as far south as Coquimbo such fluctuations are dominated by free first baroclinic mode CTWs that enter from the north along the coastal waveguide. The frequency structure of the waves entering in the north, that is, off Arica, is determined by processes farther upstream that will be addressed below. Farther south along the coast, alongshore wind events in the 6- to 10-day band interact with the incoming, first baroclinic mode wave field and also would be expected to force higher baroclinic, CTW modes (Gill and Clarke 1974). Resonant interaction with the wind would not be expected since southward propagation speeds of the coastal lows, about 10 m s⁻¹, greatly exceed CTW propagation speeds. In future work, it would be useful to apply a wind-forced model of shelf and slope flow (cf. Clarke and Van Gorder 1986) to the Chile coast, whereas Arica ASL should provide a useful upstream boundary condition in the form of free, first baroclinic mode CTWs.

To assess the source of the free CTWs reaching Chile from the north, we also analyzed sea level observations from the coast of Peru and moored current meter, temperature, and wind observations in the central and eastern equatorial Pacific during our 1991–92 study period. We found strong support for an ultimate equatorial origin of the CTW energy in the period bands of 50, 10, and 5 days. These results are summarized in Fig. 13 (filled dots on the dispersion curves of equatorial trapped waves). These dots mark the type and characteristics of the equatorial trapped waves that in our analysis have been identified—with varying degrees of certainty—as the ultimate source of the free CTWs observed off Chile.

The large amplitude 50-day period fluctuations observed off Chile started their journey about two months earlier and more than 10 000 km to the northwest in the equatorial Pacific (about 15 000 km away as the trapped wave travels). Near the date line on the equator, first baroclinic mode eastward propagating, equatorial Kelvin waves of about 50-day period were forced during the austral spring and summer 1991–92 by westerly wind events of similar period (McPhaden 1993; Kessler and McPhaden 1995). Such wind events are associated

with the tropical Madden-Julian oscillation (Madden and Julian 1971, 1972) involving the first baroclinic mode equatorial Kelvin wave in the atmosphere (Parker 1973). With a phase speed of about 10 m s⁻¹, such a wave circles the globe in about 50 days. Equatorial Kelvin waves in the ocean are most prominent during the austral summer and during the initiation phase of El Niño events (Kessler et al. 1995). Upon reaching the South American coast, these waves forced CTWs that propagated along the Peru coast into our study region. It is a remarkable expression of the efficiency of the equatorial waveguide of the Pacific and the coastal waveguide of the west coast of South America that LF currents observed on the slope off Chile in 1991-92 were dominated by waves making this long journey and that these currents were strongly related to winds more than 10000 km distant but unrelated to observed currents just 140 km farther seaward.

Our results imply that much of the significant CTW energy observed at periods of about 10 days off Chile during 1991-92 may derive from wind-driven first baroclinic mode mixed Rossby-gravity waves impinging on the coast of South America at the equator. Likewise, cross-spectral properties between and among appropriate equatorial and coastal time series suggest that free CTWs at periods near 6 and 4.5 days off Chile may be remotely forced in a similar way by the first two meridional modes of first baroclinic mode equatorial trapped inertia-gravity waves. Our scenarios for the generation of both these wave types invoked selective forcing near the zero wavenumber crossover of the dispersion relation by large zonal-scale winds along the equator (Wunch and Gill 1976). At this crossover, such waves become standing waves whereby fluctuations associated with them would occur simultaneously all along the equator. Observed phase relationships were consistent with these relationships.

We noted large 20-day period oscillations of meridional current at 45-m depth on the equator at 110°W before October 1991 and after July 1992. Halpern et al. (1988) showed that 20-day waves dominate the LF energy of near-surface meridional flow in this region in the austral winter. The properties they observed for these waves—westward phase speed of 0.8–0.9 m s⁻¹, wavelength of 1300-1600 km-place them squarely on the first baroclinic mode, mixed Rossby-gravity wave dispersion curve (open dot in Fig. 13). The 20-day oscillations may be forced by barotropic instabilities extracting energy from the latitudional shear between the North Equatorial Countercurrent and the South Equatorial Current (Philander 1978). Indeed, these "tropical instability waves" appear in the austral winter when this latitudinal shear is largest. From the dispersion relation, the eastward group velocity of 20-day MRG waves is about 0.4 m s⁻¹, but we see no evidence in our analysis for an equatorial-trapped wave-CTW connection at the 20-day period. Perhaps they never reach the South American coast, about a three month's journey from 110°W, due to downward group velocity—expected for vertically propagating, mixed Rossby–gravity waves (cf. Gill 1982) and observed by Halpern et al. (1988)— and to dissipation.

It is interesting to speculate on the possible effects of strong, remotely forced LF fluctuations in the coastal zone off Chile, in particular the 50-day oscillations, on local ocean conditions and marine living resources. After all, strong trains of these 50-day events seem to be generated with preference in the austral summer and during El Niño events as was the case during our 1991-92 study period. Clearly, such waves could have a strong effect on the nearshore coastal upwelling system: While the amount of water upwelled is set by local alongshore wind stress via the seaward Ekman transport, the source of the upwelled water may be strongly modulated by the 50-day waves. During the wave upwelling phase, water transported up into the lighted surface layer would tend to be cool and nutrient rich; during the wave downwelling phase, water transported up into this layer would tend to be warm and nutrient poor. Since marine primary production can respond on timescales much shorter than 50 days, month-scale variability of local primary production off the coast of Chile may be significantly remotely controlled by the winds in the western and central equatorial Pacific during certain periods! It will be exciting to search for influence of this remote forcing on the local pelagic ecosystem, one of the most productive in the world.

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