Long range Kelvin wave propagation of transport variations in Pacific Ocean equatorial currents

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ABSTRACT

Two 100 km scale arrays of moored upper ocean current meters, one near 0, 152W, the other near 0, 110W, were used to study the zonal transport of the strong equatorial currents in and above the thermocline. At long periods (several days), fluctuations in the vertically integrated zonal velocity (transport per unit meridional distance) at a single equatorial mooring were highly correlated with fluctuations in the total transport across the section 0-250 m and 1N-1S, which includes most of the Equatorial Undercurrent. Transport values at 152W varied from 15-55 Sv over three months. A pronounced pulse or surge in transport propagated from 152W to 110W and subsequently was evident in sea level records at the Galapagos (91W). The phase speed compared favorably with that of the first baroclinic mode equatorial Kelvin wave, with appropriate allowance for Doppler shifting by the mean flow. This is a singularly unambiguous example of long range Kelvin wave propagation in the equatorial ocean, a process of central importance in many models of equatorial adjustment to unsteady wind forcing.

1. Introduction

Studies of the transport of major ocean currents are of intrinsic dynamical interest and often have broader implications as well. In the case of the Pacific equatorial currents the principal such implication is climatic. Wyrtki (1975) proposed that El Niño, the episodic sudden warming of the coastal waters of Peru and Ecuador, was due to a relaxation of the southeast trades in the central Pacific, allowing the equatorial zonal pressure gradient to accelerate eastward flow of relatively warm water from the central and western Pacific. In his words, the "water flows eastward, probably in the form of an internal equatorial Kelvin wave. This wave leads to the accumulation of warm water off Ecuador and Peru and to depression of the usually shallow thermocline." The coastal warming gives rise to a local biological upheaval: upwelling of nutrient-rich water is cut off, commercial fisheries are devastated, and bird populations dependent on fish for food suffer enormous losses.

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The rise in sea surface temperature (SST) leads to heavy rainfall and sometimes to flooding near the coast. There may be larger-scale climatic effects beyond the immediate coastal region. Atmospheric general circulation models (Rowntree, 1972; Chervin *et al.*, 1976) show high sensitivity to eastern tropical Pacific SST, in the sense that an SST anomaly in that region, prescribed as a lower boundary condition for the model, gives rise to a stronger anomaly of subsequent atmospheric circulation than does a similar SST anomaly in other regions.

It is not clear whether Wyrtki's (1975) Kelvin wave hypothesis is the predominant mechanism for all El Niño events; the subject is one of energetic research activity at this time. Rasmussen and Carpenter (1981), by compositing several El Niño events, show that the major wind relaxation takes place some months *after* the onset of coastal warming; this seems to eliminate the wind relaxation as a precursor phenomenon. But Luther (private communication) points out that the morphology of El Niño varies considerably from event to event; in particular, the wind variations prior to and during the event may have rather different space and time patterns from case to case. Compositing therefore may not be the best way to treat the data, and it may obscure precursors in some events. Certainly in the 1957–58 case the wind data presented by Wyrtki (1975) show a complete relaxation in the central Pacific by the end of 1956, prior to the onset of warming in early 1957.

The range of possible equatorial responses to a relaxation of the trade winds is well illustrated by the model calculations of Philander (1981). Abrupt relaxation over the entire basin generates strongly accelerating currents which transport warm western water eastward, raising the eastern SST rapidly. Kelvin and Rossby waves generated at coasts at the instant of relaxation eventually arrest the acceleration, and a new equilibrium, with substantially higher eastern SST, results. If the relaxation extends over only a portion of the basin, say the center, then this portion responds much as described above, while to the east of the relaxation a more modest increase of SST is accomplished by an increase of eastward advection which propagates as a Kelvin wave. Should the relaxation be slow rather than abrupt the Kelvin waves become less discernible in the response, but the acceleration of surface currents still raises eastern SST. For an extremely slow relaxation, the ocean remains in equilibrium balance throughout (zonal pressure gradient balances zonal wind stress) and advective changes of SST are very slight.

The variety of responses possible in models is not likely to be more tightly constrained until (i) better representations of the wind field are available, and (ii) more direct observations of contemporaneous variations in equatorial currents and SST are available to test model results. In this paper we present one relevant observational contribution to the subject. We have found a strong increase or pulse in eastward transport of the equatorial currents, and have seen this increase propagate for 60° of longitude at Kelvin wave speed.

2. Experiment

As one component of the NORPAX Equatorial Experiment, also called the Hawaii to Tahiti Shuttle (Wyrtki *et al.*, 1981), we deployed a one-year array of moored upper ocean current meters and wind recorders in the central equatorial Pacific near 152W. The main objective of the Shuttle Experiment was to measure low frequency variations of the complete system of tropical and equatorial quasi-zonal currents, attempting to test and validate simple monitoring devices such as XBT sections and sea level stations on tropical islands for their effectiveness as indicators of such fluctuations. Our component of the Shuttle Experiment was designed to monitor the strength of the Equatorial Undercurrent and the overlying portion of the South Equatorial Current in a narrow zone ($\pm$ 1° of latitude) centered on the equator. Here geostrophy fails and direct measurements are essential. Simultaneously one of us (DH) maintained a similar array of instruments in the eastern equatorial Pacific, near 110W. A chart showing locations of these arrays is given in Figure 1.

Full descriptions of the data sets, which encompass many more phenomena than the transport fluctuation to be discussed here, will be published separately (Halpern and Knox 1982). Included in these descriptions will be thorough discussions of our techniques. For summary purposes the following comments suffice. Taut surface moorings with faired cable in the upper sections where currents are strong were used throughout. With careful attention to component testing and conservative sizing of components, deployments of 4-6 months are routinely realizable in the equatorial high-current regime. AMF (now EG&G Sealink) vector-averaging current meters (VACMs) were used on all moorings. The accuracy of VACMs on surface moorings has been a subject of discussion in the literature since the discovery by Gould *et al.* (1974) of erroneously high energy levels at all frequencies in records from deep VACMs on such moorings. Considerable instrument and mooring intercomparison work since then has refined knowledge of the size and source of the error. A key lesson is that for shallow instruments on surface moorings, such as those in our equatorial arrays, errors at low frequencies are modest. For example, Halpern *et al.* (1981) found that 80 m VACMs on adjacent surface and subsurface moorings differed by only 15% in rms velocity fluctuations at frequencies below 0.3 cph. This difference was obtained under open ocean mid-latitude conditions and is probably an upper bound error for our equatorial instruments, where strong mean flows prevent the rapid wave-driven direction reversals which cause the error.

3. Observations

The complete 152W data set is displayed in Figure 2. All records have been lowpass filtered to remove periods shorter than about two days. Gaps in the records

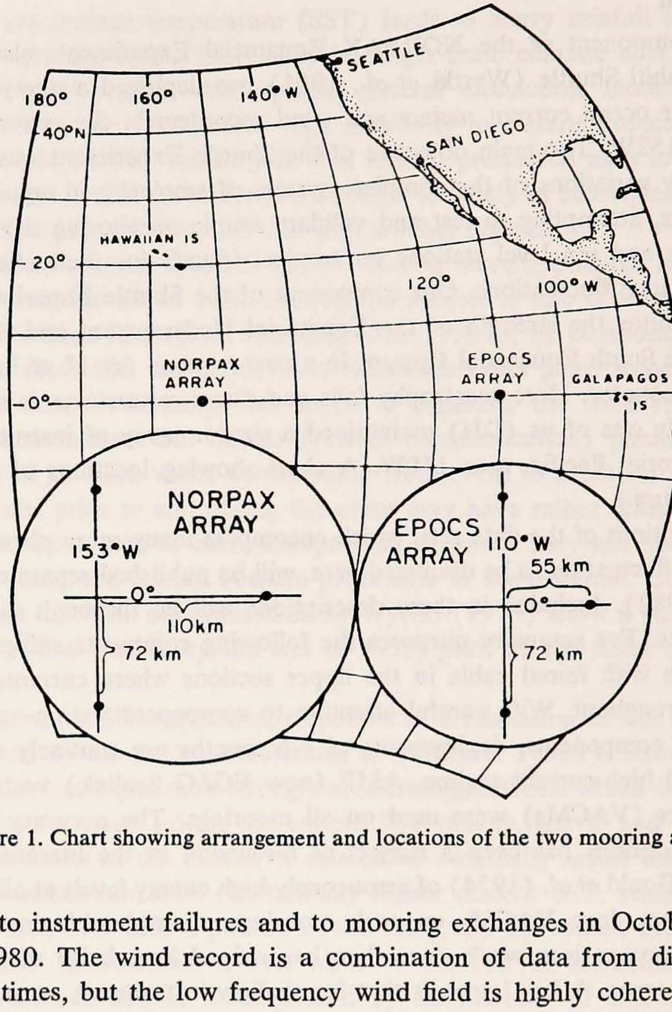

Figure 1. Chart showing arrangement and locations of the two mooring arrays.

correspond to instrument failures and to mooring exchanges in October 1979 and February 1980. The wind record is a combination of data from different buoys at different times, but the low frequency wind field is highly coherent across the entire array (Halpern and Knox, 1982), and the record of Figure 2 can be viewed as representing the entire region of the array for periods longer than a few days.

Examination of Figure 2 (dotted lines) shows that in April 1980 eastward velocity increased suddenly throughout the array. At the shallow levels (15 m-100 m) the current became more eastward as part of an apparent annual cycle. At 150 m and 250 m a pronounced transient, of one month or less duration, is observed. In order to form an integrated measure of this fluctuation, and to see how representative of the full array a single mooring is, we computed quantities related to "transport." As will be seen, two different methods of computing "transport" give equivalent results, but because of confusion surrounding the term (flux through a fixed section vs. flux of water carrying certain specified properties) we will explain both calculations in some detail. The basic data are daily-averaged values of zonal velocity at each instrument. From these we first computed a vertical integral at each day and mooring, limiting the integration to the eastward-flowing portion of the profile. If the 15 m current meter showed eastward flow, the shear between 15 m and 50 m was extrapolated toward 0 m until zero crossing occurred or the surface was reached, and the integration was extended to this depth. Extrapolation below 250 m when necessary was accomplished similarly. The resultant time series of vertical integrals is referred to as "eastward transport per unit width," in $m^2\ s^{-1}$. To compute a total transport for each day we multiplied the eastward transport per unit width at each mooring on that day by a latitude span of 40′, i.e., we conceive of each mooring representing the flow within ± 20′ of its position and ignore the fact that all 3 moorings are not at the same longitude. This time series is the "1N-1S Undercurrent transport," in Sv. The underlying idea is to count only the eastward-flowing water, which corresponds to our qualitative concept of the Undercurrent. In this sense, it is an attempt to calculate a transport of water with a certain property (eastward flow).

The second technique proceeded from the same basic data and used the same latitude integration, but carried out the three vertical integrals between fixed limits of 15 and 250 m. This technique computes transport through a fixed section. The upper and lower depth limits chosen are not far from zeroes of the mean flow; *c.f.* Figure 3.

Both calculations were applied to the data of the third deployment when all instruments worked without failures. The first result, of some importance for future measurements, is that the fluctuation is sensed about as well by a single mooring as by the full array. Figure 4a shows a comparison between the eastward transport per unit width at the equatorial mooring and the 1N-1S Undercurrent transport as defined above; the relationship between the two quantities is very tight. Firing's (1981) regression curve relating transport per unit width at the equator to the total transport of the Undercurrent determined from closely spaced profiles along 153W was similar to the curve shown in Figure 4a, indicating that our index of total transport in fact coincides with a spatially well-resolved measurement. A clear implication is that much could be learned about long period current fluctuations with single equatorial moorings at several longitudes, an economy of resources over setting out horizontal arrays. The second result is that both techniques of computing total transport are nearly equivalent. In Figure 4b the time series from both total transport computations, as well as that of the eastward transport per unit width at the equatorial mooring, are shown. The transport surge is revealed equally well by all three curves. The surge appears at the end of a longer-term rise, during which transport changes from 15 to 55 Sv.

There is no abrupt change in the local wind of Figure 2 at the time of the transport surge, so we suspect a propagating event and we turn to our other data set for further evidence. In Figure 5 the equatorial eastward transport per unit

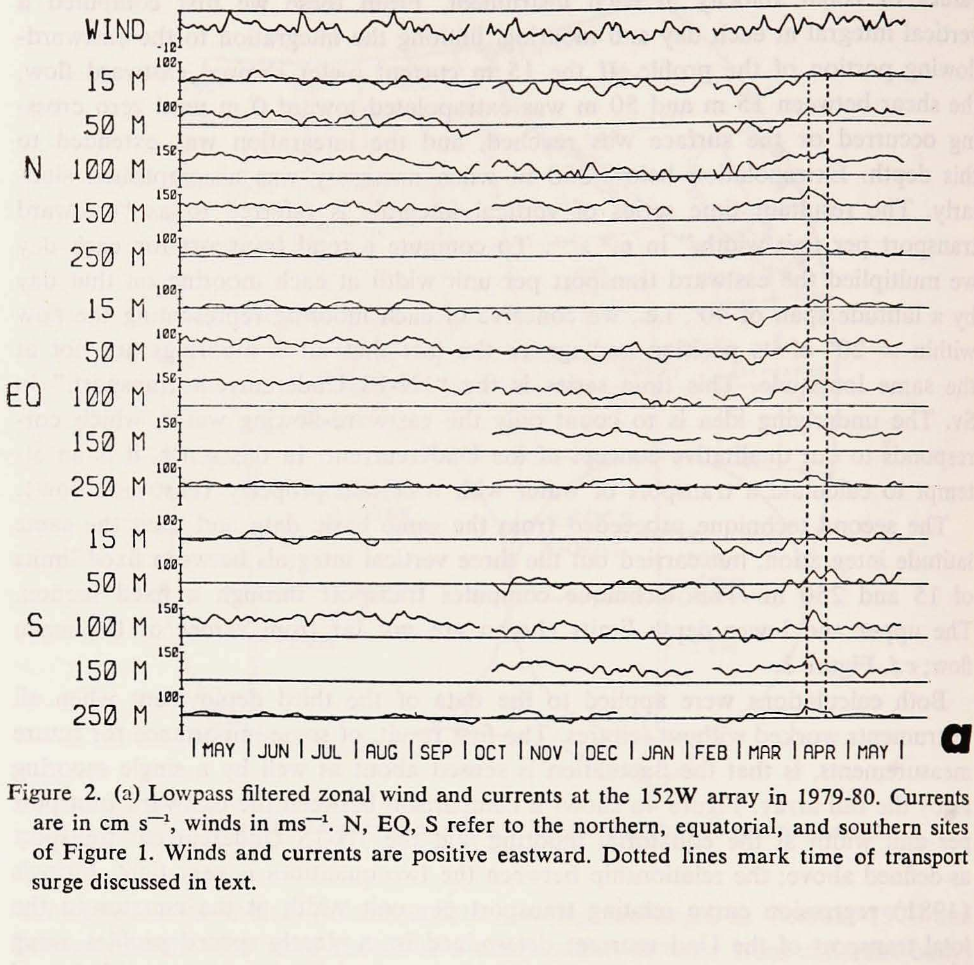

Figure 2. (a) Lowpass filtered zonal wind and currents at the 152W array in 1979-80. Currents are in cm s^{-1}, winds in ms^{-1}. N, EQ, S refer to the northern, equatorial, and southern sites of Figure 1. Winds and currents are positive eastward. Dotted lines mark time of transport surge discussed in text.

width of Figure 4 is replotted along with a similar curve from the 110W array. The same surge appears at 110W about 18 days later. This corresponds to a propagation speed of about 3.0 ms^{-1}. Still later, the surge appears as an elevation in sea level at Isabela Island in the Galapagos, in a record kindly provided by Drs. S. Hayes and P. Ripa. The propagation speed from 110W to Isabela is about 2.7 ms^{-1}. Luther (1980) has computed first baroclinic mode Kelvin wave speeds in the equatorial Pacific from hydrographic data. Values range from 3.0 ms^{-1} at 180W to 2.3 ms^{-1} at 90W. Linearly interpolating these values over the two longitude spans 152-110W and 110-91W yields average propagation speeds for this mode of about 2.6 ms^{-1} and 2.4 ms^{-1} respectively, or 15% and 13% less than our observations. This discrepancy may well be due to Doppler shift of the wave by the mean eastward current, as discussed by McPhaden and Knox (1979). In their Figure 9, for Rossby number $R = 0.2$ (which is roughly appropriate to their

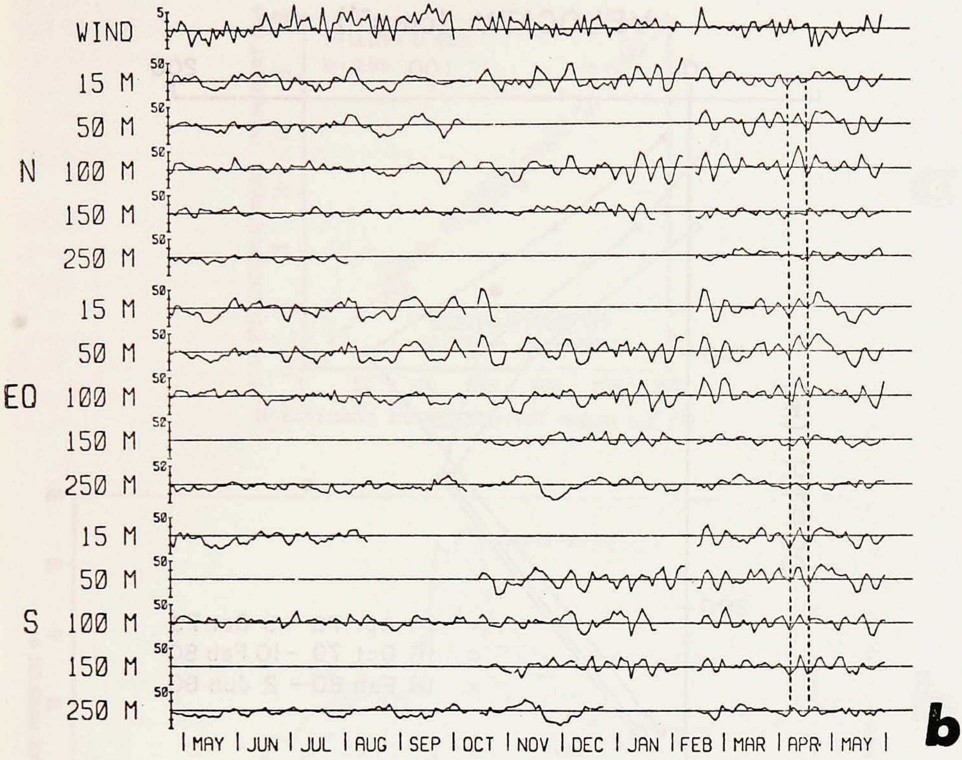

Figure 2 (b) Like (a) for meridional wind and current, positive northward.

Pacific mean flow profile), and for a wide range (0-3) of the parameter α (which sets the mean flow length scale), the Kelvin wave phase speed σk^{-1} is increased by 10-25% over its value in the absence of mean flow.[3] Considering the severe approximations of their model and of our interpolations to Luther's (1980) phase speed estimates, our observed higher phase speeds are quite consistent with the Doppler effect.

During an intense one-week storm near Tarawa (1° 30′N, 173E) in the latter part of March the zonal wind changed by about 7.5 ms^{-1} as normal trade winds were replaced by westerlies (C. Eriksen, private communication). This is the kind of abrupt forcing event which efficiently generates Kelvin waves (Philander, 1981).

3. They modeled the mean zonal flow as vertically constant in the upper layer of a two layer system and of Gaussian shape meridionally. For a first baroclinic mode horizontal length scale $L = (C/\beta)^{1/2}$ with $C = 2.5$ ms^{-1}, their Rossby number $R = \bar{U}/\beta L^2 = 0.2$ implies a mean flow amplitude $\bar{U} = 0.5$ ms^{-1}, or roughly the vertical average of the profiles in Figure 3. Their parameter $\alpha = 2\, L^2/\sigma^2$ where σ is the (dimensional) standard deviation of the Gaussian, so $\alpha = 0$ corresponds to an infinitely broad flow, while $\alpha = 3$ corresponds to $\sigma = 135$ km.

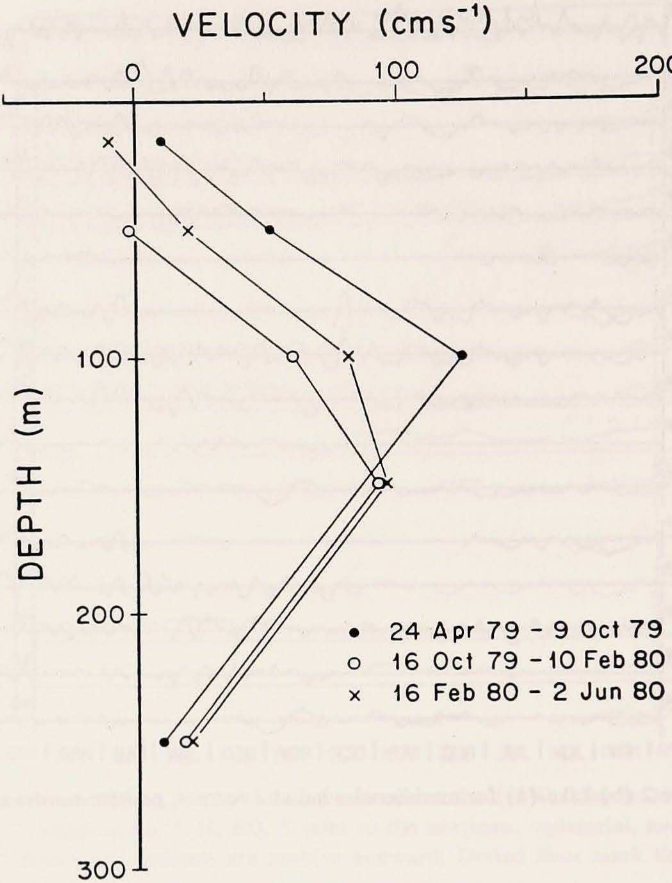

Figure 3. Mean zonal velocity profiles (positive eastward) at the equatorial mooring during the three settings of the 152W array.

If the observed pulse was forced by this storm it should have arrived at the 152W array about 15 days later, and this is consistent with our measurements.

4. Discussion

We have observed one clear example of a sharp increase in the transport of the equatorial currents in the Pacific and have seen it propagate eastward as a Kelvin wave in the absence of any similarly abrupt forcing event in the local winds. It is noteworthy that the nondispersive Kelvin wave does indeed preserve its shape for 6000 km; the sharp leading edge of the pulse is evident in all three time series of Figure 5. While this was a solitary short-term event, not part of a massive equatorial adjustment (the year of our experiment was not an El Niño year), it

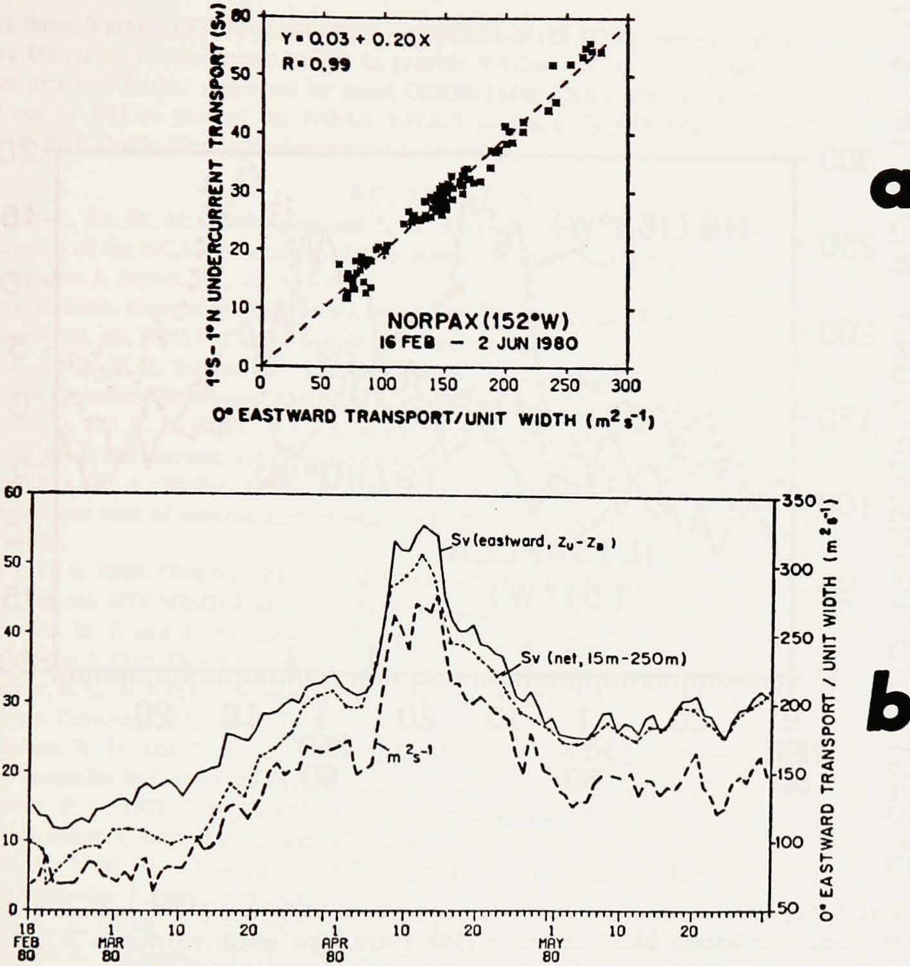

Figure 4. (a) Linear regression between eastward transport per unit width at equator and 1N-1S Undercurrent transport, as defined in text, third deployment at 152W. R is correlation coefficient of the regression. (b) Time series of the two quantities in (a) and of fixed-section transport in Sv (dots) as defined in text.

was still a major fluctuation. The current meter records at 110W subsequently were extended to August 1981 (not shown). The May 1980 transport surge was associated with 100 m zonal currents 30% greater than at any other time in the extended record. The pulse apparently was forced by winds to the west of our arrays; the total distance traversed was over 10,000 km. It would clearly be of interest to document such changes in transport, and their propagation (or lack of it), during a full El Niño. The fact that a single equatorial mooring measures the transport variation adequately (Fig. 4b) leads us to think that a sparse array

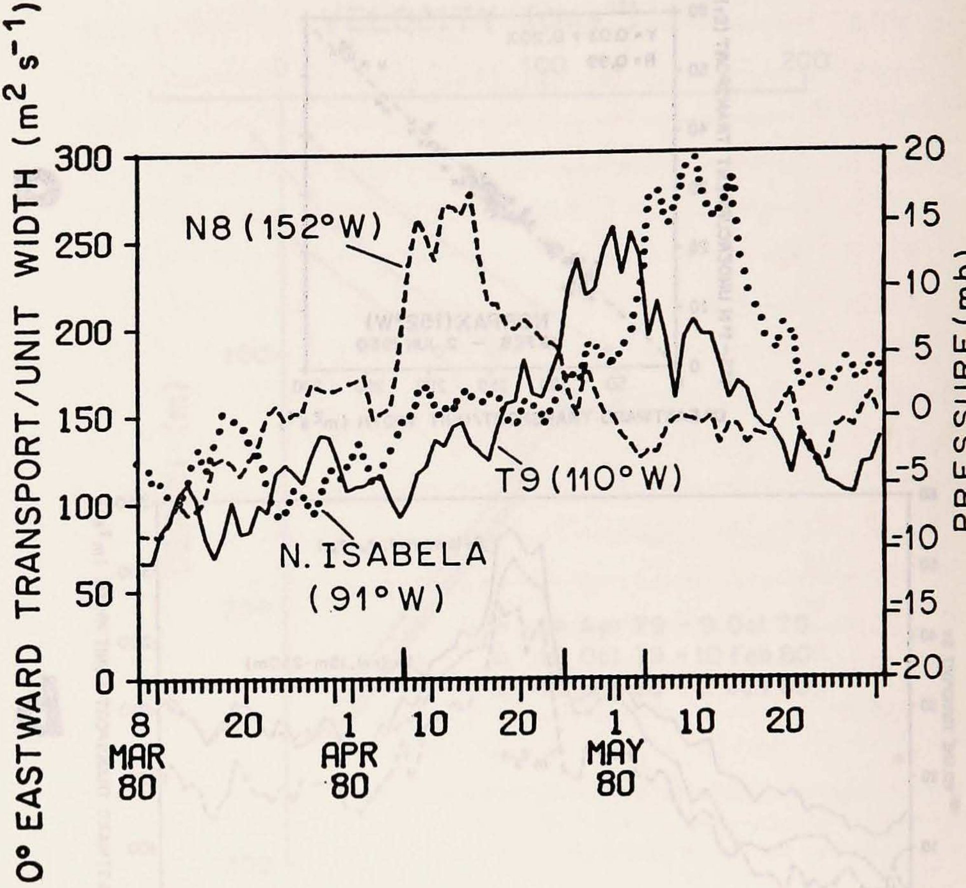

Figure 5. Equatorial eastward transport per unit width at 152W and 110W, and sea level on the equator at Isabela Island, in spring 1980. Propagation speeds referred to in text are judged from times of leading edge of transport surge, marked by vertical lines.

of such moorings (one at each of several longitudes on the equator), maintained for a long time, could do an excellent job of acquiring this sort of data. If a good description of the tropical wind field could be obtained at the same time we would then have a much more complete description of one El Niño case history over the entire equatorial Pacific than presently exists.

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