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SEA LEVEL AND NEAR SURFACE TEMPERATURE VARIABILITY AT THE GALAPAGOS ISLANDS, 1979-83.

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Abstract.

An array of near surface temperature and pressure gauges on the Galápagos Islands was used to describe changes in the thermal field and sea level which occurred during the 1982-83 El Niño event. These changes are contrasted with the variability observed during the three year period (July 1979 to July 1982) prior to the event. The location of gauges on the western side of the archipelago permitted monitoring of the strength of the Galápagos Front. In non El Niño times the cross front temperature gradient had a strong annual variation; it was largest in October. A station usually north of the front had pronounced semiannual variability while south of the front the annual component was dominant. The superposition of these two signals led to the observed annual change in frontal strength. During the El Niño event the front disappeared from the islands in October 1982 and did not reappear until July 1983. During the redevelopment, cold water occurred first at the southern stations; its occurrence north of the front was delayed by about 1 month. By October 1983 the cross frontal temperature gradient in the islands was near normal.

El Niño related sea level variations at the Galápagos reflected the large scale changes throughout the tropical Pacific. Maxima were observed in December 1982 and May 1983. The onset of rising sea level in early August 1982 followed by about one month the sea level rise in the central Pacific. This initial signal had properties consistent with equatorial Kelvin wave dynamics. The subsequent developemnt was not so easily described.

Resumen

Se utilizó un arreglo de sensores de temperatura y presión cercanos a la superficie en las Islas Galápagos, para describir los cambios del campo térmico y del nivel del mar que ocurrieron durante El Niño 1982-83. Estos cambios fueron comparados con la variabilidad observada en un período de tres años previo al evento (julio de 1979 a julio de 1982). La ubicación de los instrumentos en el lado occidental del archipélago permitió observar la intensidad del frente de Galápagos. En períodos en los que el fenómeno El Niño está ausente, la gradiente de temperatura normal del frente tuvo una fuerte variación anual; su máximo se observa en el mes de octubre. Una estación normalmente ubicada al norte del frente mostró una pronunciada variación semianual, mientras que al sur del frente la componente anual fue dominante. La superposición de estas dos señales produjo el cambio observado en la intensidad del frente. Durante el evento El Niño, el frente desapareció de las islas en octubre de 1982 y no reapareció hasta julio de 1983. Durante su reestablecimiento, agua relativamente fría apareció primero en las estaciones más al sur y un mes más tarde en la estación ubicada al norte del frente. Para octubre de 1983 la gradiente de temperatura normal del frente estaba cerca de su valor normal en las islas.

Las variaciones del nivel del mar en las Galápagos debidas a El Niño fueron un reflejo de los cambios en escalas grandes que ocurrieron en todo el Pacífico tropical. Sus máximos fueron observados en diciembre de 1982 y mayo de 1983. El comienzo del aumento del nivel del mar al principio de agosto de 1982 ocurrió aproximadamente un mes más tarde que la correspondiente elevación de la superficie del mar en el Pacífico central. Esta señal inicial tenía propiedades consistentes con la dinámica de ondas de Kelvin ecuatoriales. Su desarrollo posterior no puede ser descrito en una forma tan sencilla.

Introduction.

The large scale oceanic and atmospheric variability associated with the 1982-83 El Niño event are discussed in Wyrtki (1985). His article sets forth the changing physical environment in which the Galápagos Islands were situated and suggests interpretations of the large scale changes in terms of local and remote wind forcing of the ocean. In this paper we focus on variability at the Galápagos as observed in our array of near surface pressure and temperature gauges. Since our study began in 1979, we have a three year time period before the event (July 1979 to July 1982) to study the normal low frequency variability. This variability is contrasted with changes during the El Niño. Results are used to investigate some of the dynamical inferences based on the large scale analysis and to study some peculiarities of the inter-island differences at the Galápagos.

Much of our knowledge of the low frequency, large scale variability of the tropical Pacific

Ocean has been obtained through analysis of island and coastal sea level records. Non-tidal sea level variations reflect a variety of atmospheric and oceanic changes. In low latitudes, where atmospheric pressure perturbations are small and theoretical considerations suggest that low frequency barotropic (depth independent) oceanic pressure fluctuations are negligible, the dominant cause of non-tidal sea level change is thought to be internal readjustments of the density stratification of the water column. These readjustments could be associated with heating and cooling, precipitation and evaporation, large scale internal wave processes, or geostrophic ocean currents. Any process which replaces dense water with less dense water tends to make sea level rise.

In addition, a dynamic effect on sea level which is often ignored is the pressure perturbation associated with an ocean current flowing around an obstacle. Pressure is higher on the upstream than on the downstream side of the island. This process is potentially important in a swift current such as the Equatorial Undercurrent (EUC) and has been the subject of a recent study at Jarvis Island in the central Pacific (Roemmich, 1984). Results suggest that the strength of the EUC (assuming meanders are not important) can be simply monitored with a pair of pressure gauges spanning the island. While this result is encouraging from the standpoint of monitoring the flow, it reveals a complication in the use of island sea level records as an index of large scale ocean variability. The records may adequately describe the qualitative structure of the large scale changes; but, in regions of locally intense currents, quantitative studies may be severely contaminated.

Sea level variability can be determined either by directly measuring water depth relative to a fixed reference point (e.g., a surface tide gauge) or by measuring pressure fluctuations with a shallow, bottom moored pressure gauge. The two measurements will differ because of atmospheric pressure changes and density variations in the water above the gauge. Since pressure change is the parameter of most interest to oceanographers, at mid and high latitudes one generally assumes the "inverse barometer effect" and adds atmospheric pressure to sea level to obtain "synthetic subsurface pressure." As noted above, in tropical regions atmospheric pressure changes are generally small and can be ignored; sea level and subsurface pressure are considered equivalent. One further point to keep in mind is that both direct sea level and near surface pressure measurements only establish the relative change in pressure between two locations. The absolute pressure or height difference cannot be determined without knowing the level of the gauges relative to the geoid.

In order to study the sea level variations near the Galápagos Islands we have maintained an array of near surface pressure and temperature gauges since July 1979. Our specific interests were interpretation of the measurements in terms of (a) low frequency equatorial trapped waves, (b) changes in upper ocean heat content, and (c) fluctuations in the near surface zonal velocity. To investigate these processes gauges were deployed on the western side of the archipelago (figure 1) at Caleta Iguana (0°59'S, 91°30'W) and Vincente Roca (0°3'S, 91°28'W) on Isabela Island (referred to as SI and NI, respectively) and at Wenman (Wolf) Island (WI; 1°24'N, 91°50'W). A fourth gauge was added on the eastern side of the islands in November 1980 at Bahia Hobbs (SC; 0°42'S, 89°18'W) on San Cristóbal Island. The mean depth of the instruments are SI: 14 m, NI: 16 m, WI: 25 m, SC: 16 m. These gauges were recovered and deployed annually, normally in November using the M/V BEAGLE III. Technical details of the gauges and the data processing are discussed in

Hayes et al. (1978), Ripa and Hayes (1981) and Hayes and Halpern (1984). In this report pressure changes are referred to in terms of equivalent sea level fluctuations assuming 1 mbar = 1 cm.

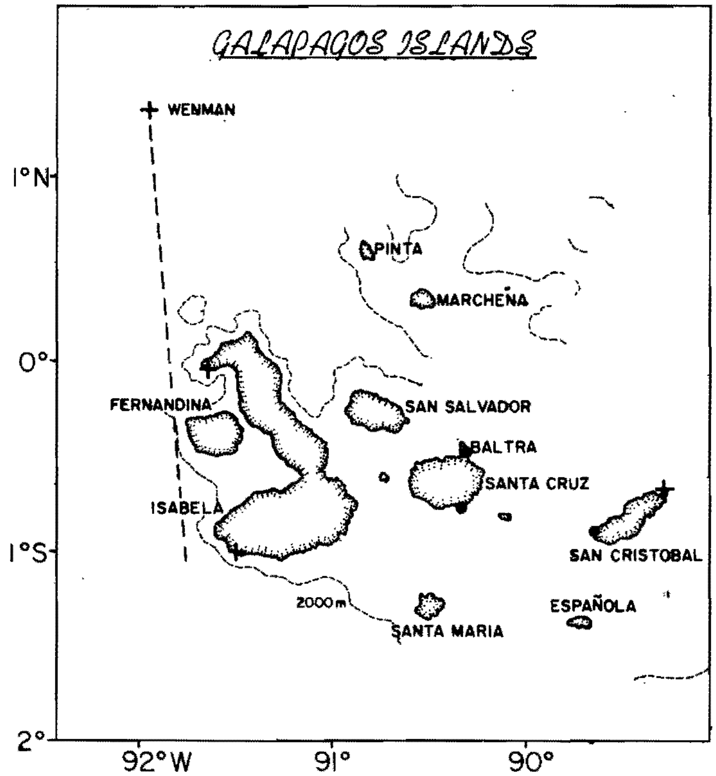

Figure 1. Location of subsurface pressure and temperature gauges (+) in the Galápagos Islands. The location of historical (Baltra and San Cristóbal) and current (Santa Cruz tide gauge measurements (●) are also shown. The trackline of the XBT sections west of the islands is indicated by the dashed line.

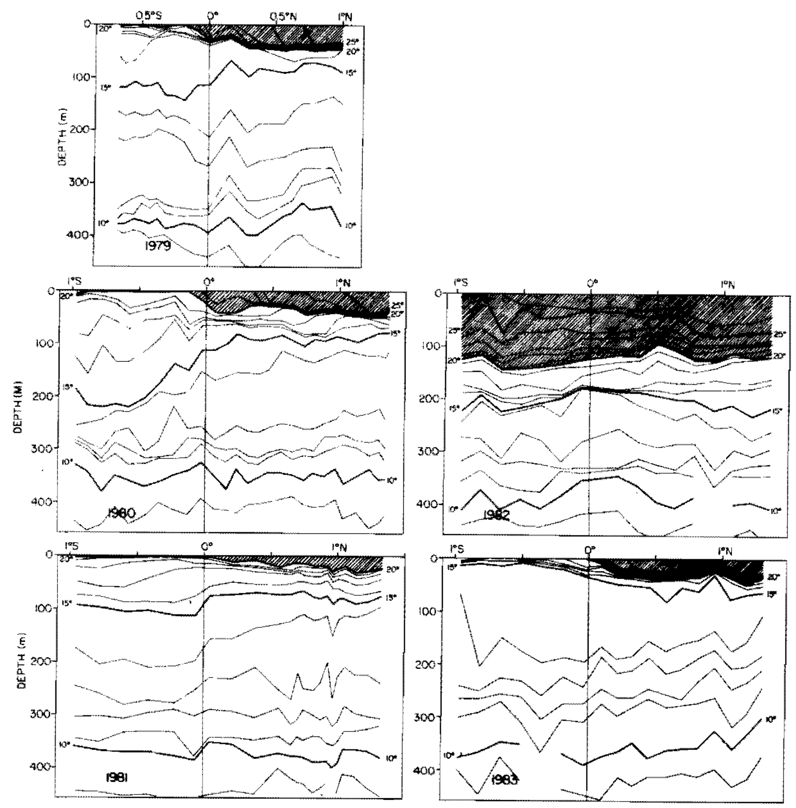

Figure 2. Temperature sections taken in November along the trackline shown in figure 1 for the five years indicated.

Temperature Variations.

On deployment/recovery cruises XBT temperature sections (figure 2) were made along the trackline shown in figure 1 on the western side of the islands from 1°S to 1.5°N. Near surface structure on these sections is dominated by the Galápagos Front. This feature of the large scale oceanic circulation in the eastern Pacific crosses the equator near the Galápagos and extends southeastward intersecting the South American coast at about 5°S. It separates warm, fresh tropical waters on the north from the cooler, more saline water of the Peru Current. The thermal front varies annually (Wyrtki, 1966). It is strongest from May to November when temperature differences of up to 5°C and salinity changes of up to $1^{\circ}/_{oo}$ in 50 km are observed. From January to March when the southern waters warm the thermal front disappears; however, the salinity gradient remains. During El Niño events when the northward wind stress relaxes, there is some evidence (Wooster and Guillen, 1974) that the warm fresh water flows southward and the front disappears or moves far to the south.

In our November temperature sections (water warmer than 20°C is cross hatched in figure 2) the Galápagos Front was well developed north of the equator in 1979, 1980, and 1983. Surface temperature changed by about 5°C from 0° to 1.3°N. The mixed layer depth was about 25 m at the northern limit and negligible at the equator. In November 1981 the Front was displaced to the north and cold water extended north of Wenman Island. The greatest change in near surface structure was apparent in November 1982. This section occurred about four months after the initial sea level rise associated with the 1982-83 El Niño (see figure 7). No evidence of the Galápagos Front was seen. Waters warmer than 26°C extended from the north to about .7°S and no surface water cooler than 25°C was observed. The 20°C isotherm which typically is within 25 m of the surface on the equator was 120 m deep. A massive change in upper ocean thermal structure had occurred.

No salinity measurements were made on our BEAGLE III cruises. However, on 7-8 November 1983, about two weeks prior to the XBT section in figure 2 the NOAA Ship RESEARCHER occupied a CTD section from 2°N to 2°30'S along 92°30'W about

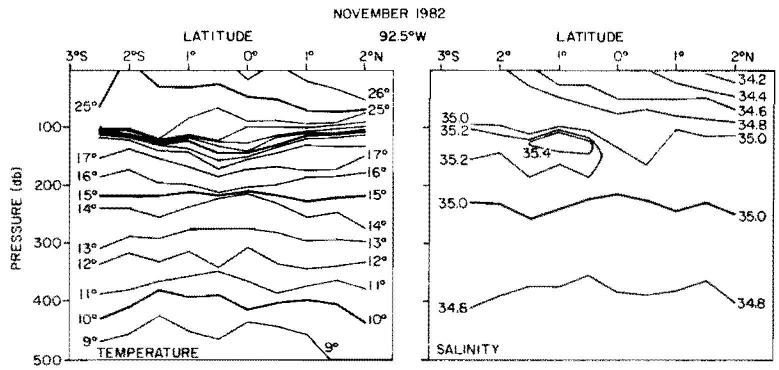

Figure 3. Temperature and salinity distribution along 92°30'W in November 1982.

100 km west of the Galápagos. Temperature and salinity distributions on this section are shown in figure 3. We note that although surface temperature was fairly uniform, a strong meridional salinity gradient was still observed. The warm water near the equator does not have the low salinity ($<34^{\circ}/_{oo}$) characteristic of waters north of the Galápagos Front. It thus appears that, initially, the warm waters near the Galápagos in the 1982-83 event were not caused by southward advection of the warm fresh tropical waters of the north. Rather, their characteristics are more nearly associated with waters to the west of the islands. In contrast, figure 4 shows the April 1983 temperature-salinity (T-S) diagram at 0°, 92°30'W overlaid on the November 1982 data. In April, surface waters were very warm (T_o = 30.4°C) and fresh (S_o = 32.9 $^{\circ}/_{oo}$). This low salinity water was in a thin (~15 m) mixed layer which ended in a sharp halocline. The temperature quasi-mixed layer was thicker; water warmer than 29°C extended to 40 m. However, with our data we cannot determine whether the low salinity water was the result of locally heavy rainfall or advection from the north.

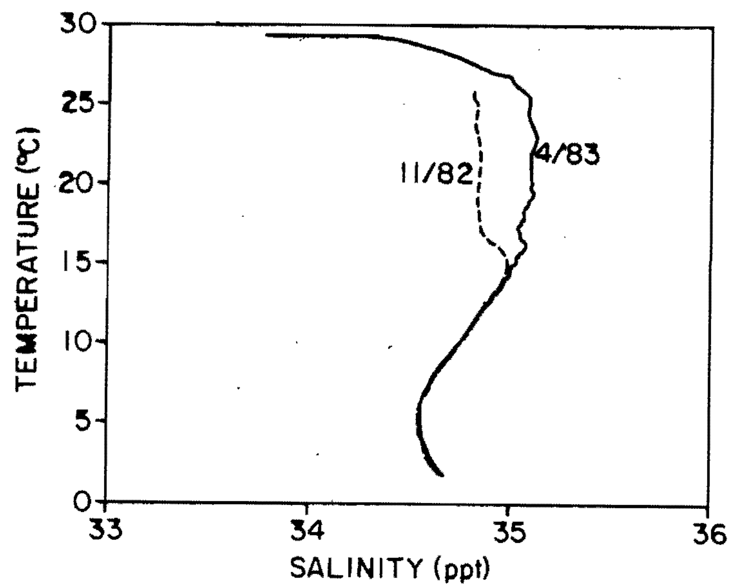

Figure 4. Temperature-salinity diagrams at 0°, 92°30'W for November 1982 and April 1983.

The general position of the Galápagos Front just north of the equator west of the islands suggests that our island temperature records, particularly at NI and WI, could provide an effective monitoring of frontal location and strength. Daily average values of temperature at the four gauges are shown in figure 5. Considerable high frequency structure is superimposed on the dominant low frequency variations. The development of the 1982-83 event is seen by the onset of warming in mid-August 1982 and a continually increasing temperature until July 1983. The high frequency variance levels were similar at all gauges. Spectral analysis of the

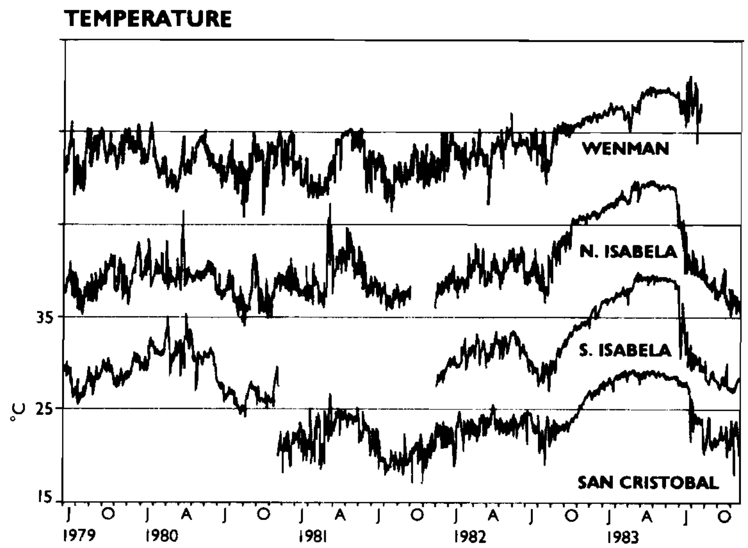

Figure 5. Daily averaged values of temperature at the four sites shown in figure 1 for the period July 1979 to November 1983. The temperature scale for San Cristóbal is indicated, successive plots are offset by 10°C.

records revealed no particular frequencies with statistically significant variance peaks. Rather, the variance density spectra in the period band from 180 to 2 days decreased monotonically from low to high frequency with a shape approximated by frequency to the -1.3 power. During the El Niño event high frequency variance decreased. This result presumably reflects the greater mixed layer depth and lateral homogeniety during the warm episode.

The low frequency variability is more easily discussed in terms of low pass filtered data (figure 6a) with a 30 day half-power point. The occurrence of the Galápagos Front between NI and WI is obvious in this figure. NI and SI lie south of the front and generally track one another. WI

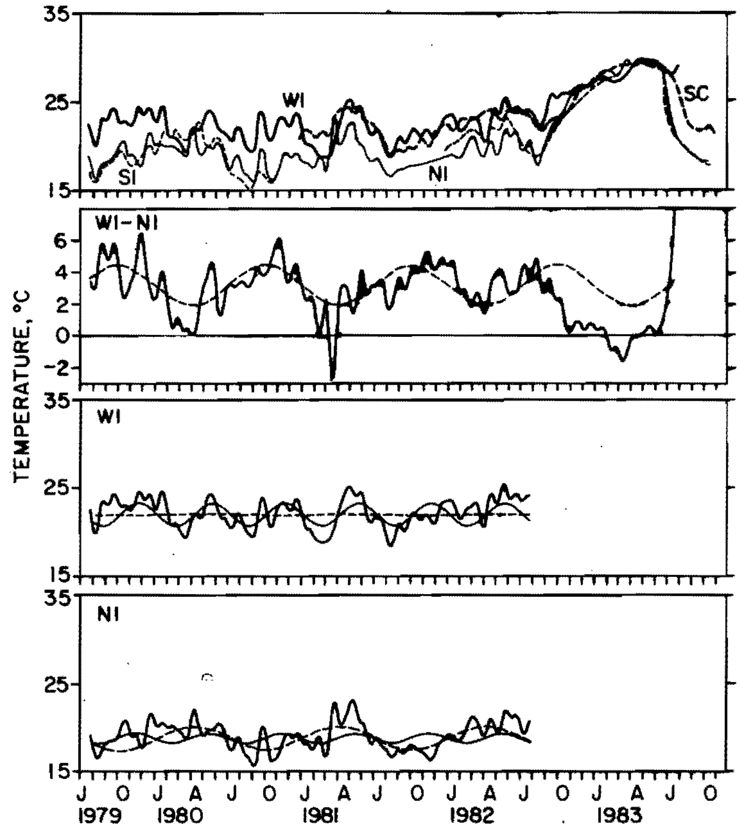

Figure 6. (a) Low pass (30 d half power point) filtered temperature data at all sites. (b) temperature difference between WI and NI, dashed curve indicates annual component fit to data from July 1979 until July 1982. (c) temperature record at WI dash curve is annual component, solid curve is semiannual component fit to non El Niño period. (d) same as (c) for NI.

and SC also follow each other reasonably well and appear to be on the warm side of the front. It is surprising that SC at 0°42'S agrees better with WI (1°24'N) than with the southern stations. Apparently, the front angles sharply across the equator in the Galápagos. This result is qualitatively in agreement with the suggestion that the EUC flows southward around the islands (Stevenson and Taft, 1971; Lukas, 1981). Upwelling associated with this current flowing eastward along upward sloping isopycnals would keep the western side of the Galápagos cool near the equator and could contribute to maintaining cooler water south of the islands.

The annual variation in the strength of the front is apparent in figure 6b where the difference WI-NI is plotted. The gradient weakens or disappears each January to April and is strongest in about October. This cycle was interupted by the El Niño event which reduced all gradients in November 1982 and the water remained relatively homogeneous until June 1983. The smooth curve on figure 6b is a fit of the three years of data July 1979 to July 1982 to an annual cycle. This fit explains 37% of the variance. The amplitude is 1.3°C and the phase of the maximum relative to 1 January is 274° (i.e., maximum temperature difference in October). This annual amplitude corresponds to a meridional gradient of 8.5×10^{-6} °C m^{-1}.

Close examination of figure 6a shows that the annual cycle in frontal strength does not result solely from a warming in the south. The water at WI north of the front also cools down. The difference between WI and NI temperature records is further exemplified by figures 6c and 6d where these data and the annual and semi-annual components are plotted. These components are again based on the three pre-El Niño years July 1979 - July 1982. At WI the annual component was negligible (explained 0.2% of the variance); however, the semi-annual component was significant (explained 32% of the variance). The semi-annual amplitude and phase are 1.3°C and 287° (maxima in May and November). At NI only the annual component is significant accounting for 37% of the variance with an amplitude of 1.4°C and a phase of 100° (maximum in March). The semi-annual component only accounted for 7% of the variance. The minimum frontal strength in March corresponds to maximum annual component at NI and a near minimum semi-annual component at WI. Similarly, the maximum frontal strength in October occurs at a time when these two components coincide to yield warmest water at WI and coolest at NI.

The reasons for the differences in thermal variability at the equator and at 1°24'N are not known. West of the Galápagos at least two processes (oceanic advection and reduced wind stress) contribute to the warming in March-May. Halpern (1984) has shown that surface currents at 0°, 110°W reverse annually nearly in phase with the warming. The February - May period is associated with eastward flow; the October - November cold period is associated with strong westward flow. Lukas (1981) has suggested that when the eastward current is strong, it flows through the Galápagos Archipelago (rather than to the south) carrying warm relatively salty equatorial waters. This could lead to the initial warming on the equator but a relative cooling at WI, which had been in the warm fresh northern tropical waters. Further developments in the February - May period include reduction in the SE trade winds and a near vanishing of the meridional wind component. These effects would contribute to continued warming on the equator and to the north, essentially removing the forcing mechanism which appears to be responsible for the Galápagos Front. Constructing a plausible scenario for the temperature changes at WI in July through December is somewhat more difficult. The equatorial cooling is likely a response to the increase in the SE trade winds and the SEC. These winds reestablish the front; however, there is presently insufficient information on the interaction of the winds and the ocean currents to describe in detail which processes lead to the observed thermal time series.

The onset of the warm event is most clearly marked at the equatorial station, NI, where temperatures (figure 5) rose sharply in late August (5°C from August 15 to 25). A similar, essentially simultaneous temperature jump occurred at Wenman Island; however, this change appeared as part of a high frequency fluctuation superimposed on a general warming trend. In fact, at WI this warming extended fairly persistently from August 1981 to July 1983. The El Niño warming was not seen as a unique signal at this northern station. At the other island locations an identifiable temperature peak was observed. It rose gradually from August 1982 until February 1983. It then remained relatively constant until mid June. From June 20 to July 13 temperatures south of the normal position of the Galápagos Front fell 9°C at SI and NI. North of the front, WI and SC remained warm. It was not until one month later that SC temperatures dropped (the WI gauge failed at this time so we must use the SC record as an index of the waters north of the front). During this one month period the surface temperature difference from SI to WI was nearly 10°C (figure 6) which was the most intense cross frontal gradient observed in our 52 month record. The timing of the temperature changes suggests that advection of cool surface waters from the south may be important in the reemergence of the Galápagos Front after the warm event. However, it is not known what limits the northward flow of cold water and produces the strong thermal front observed.

Sea Level Variations.

Sea level fluctuations at each island site are shown in figure 7. Records have been detided using response analysis techniques (Munk and Cartwright, 1966) and low pass filtered using a 30 hour half power point filter. The mean sea level at each gauge has been subtracted. Visually, it is clear that the coherent sea level variations are much larger than the incoherent changes. All records appear similar. There were a series of high sea level events from July 1979 to May 1980, then a relatively quiet period until March 1981. Another sea level rise was seen in April 1982. In the three year pre-El Niño period the three highest sea

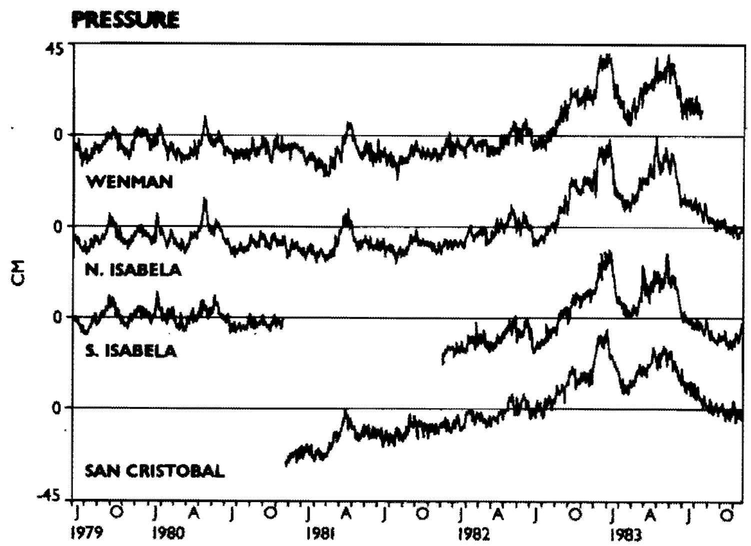

Figure 7. Daily values of pressure at the four sites shown in figure 1. Data have been detided and filtered (30 h half power point). The mean pressure over the time period shown has been removed from each series. Units of pressure are mbar.

levels seen were in May 1980, April 1981, and May 1982. These events amounted to changes in sea level of 15-20 cm. Near the end of July 1982, the first sea level changes clearly related to the warm event occurred; by October sea level had risen 30 cm. It rose another 15 cm in the next two months before falling nearly 30 cm to a relative minimum in February 1983. A second sea level peak extended from March to June. After this, sea level fell at all sites and was near pre-event levels in October 1983.

The frequency composition of the pre-El Niño variability was examined by computing power spectra for the two longest records, NI and WI. These spectra (figure 8) were estimated by averaging Fourier coefficients from the three one year pieces July 1979 to July 1982. The main difference between the two locations occurred near periods of 5 days. WI had a pronounced peak at this frequency while the NI spectrum was featureless. At lower frequencies the two locations were fairly similar. A statistically significant peak occurred at about 12 d periods; below this the variance density increased with decreasing frequency until the semi-annual frequency. The relative drop off in variance density at the annual period indicates that the semiannual component is more important than the annual component at both islands.

To look more closely at the annual (η_a) and semiannual (η_s) variability we fit these constituents to the three years July 1979 to July 1982. The amplitude (in centimeters) and phase of these two signals at NI were $\eta_a = 2.3$, $\phi_a = 124°$ (maximum on 4 May) and $\eta_s = 3.0$, $\phi_s = 251°$. At WI these parameters were $\eta_a = 0.7$, $\phi_a = 121°$ and $\eta_s = 2.9$, $\phi_s = 283°$. The standard errors of the amplitude estimates were all about 1.2 cm. The annual component accounted for about 10% of the low pass filtered (30 d half power point) variance at NI and

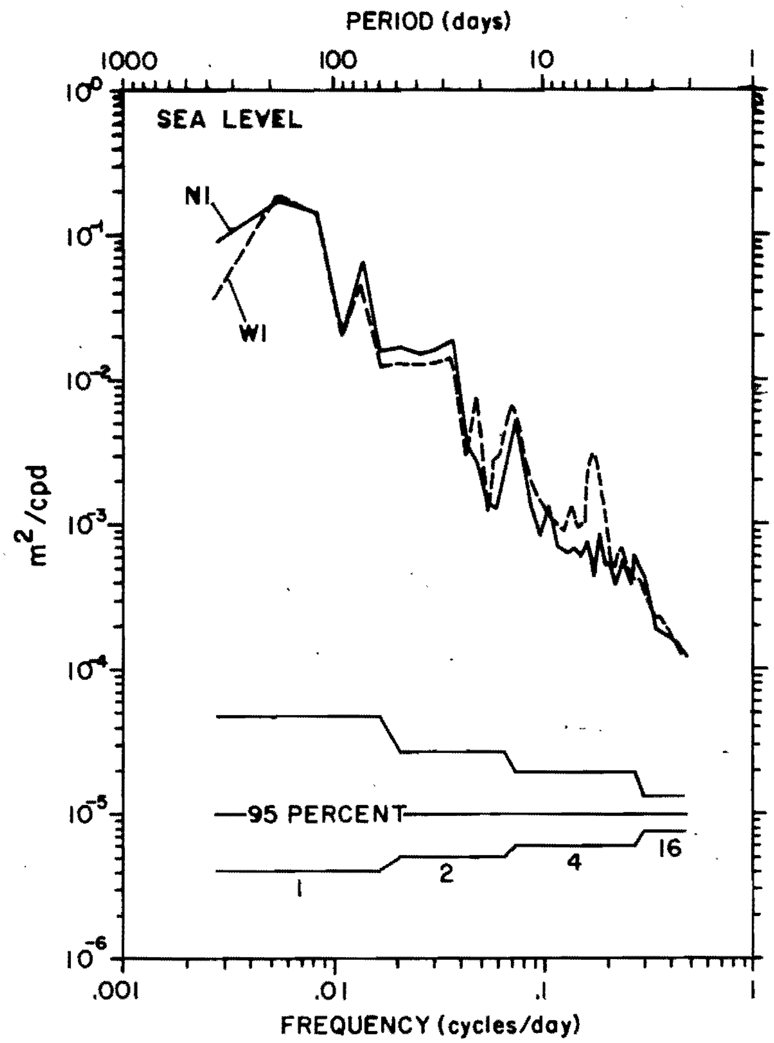

Figure 8. Power spectra of pressure variations from July 1979 to July 1982 at NI and WI.

less than 3% at WI. On the other hand, the semi-annual component accounted for about 18% of the low pass filtered variance at both locations. These results for the three years studied here contrast with the longer historical records at Baltra (1968-77) and San Cristóbal (1959-68) (see figure 1) which Wyrtki and Leslie (1980) analyzed. They found annual (semiannual) components at Baltra of $\eta_a = 3.0$, $\phi_a = 84°$ ($\eta_s = 1.2$, $\phi_s = 101°$) and at San Cristobal of $\eta_a = 2.4$, $\phi_a = 93°$ ($\eta_s = 1.2$, $\phi_s = 7°$). As indicated by the relatively large standard errors of the regression coefficients, our records are too short to quantitatively study variations in the annual and interannual sea level components among the islands. Nevertheless all analyses indicate that these components are small compared to the event-like signals seen in boreal spring or to the El Niño sea level rise.

As noted in figure 7, the low frequency changes at all the islands were highly correlated. Comparison with Wyrtki (1984) indicates that our pressure records were also well correlated with the tide gauge at Academy Bay on Santa Cruz Island. The relation of these data to the large scale sea level changes are discussed in Wyrtki (1985) and a discussion of events which appear to propagate between the central and eastern Pacific is presented in Lukas *et al.* (1984). The latter paper finds evidence for equatorial waves in the cross correlation functions between Galápagos Island and Jarvis Island (0°23'S, 160°W) or Christmas Island (2°N, 157°W) sea level. Correlation peaks were found with lags corresponding to eastward propagation of first and second vertical mode Kelvin waves and westward propagation of first vertical, first meridional mode Rossby waves. In figure 9, NI and Jarvis sea level are over plotted with the Jarvis record delayed to 30 d which corresponds to a phase speed of c = 2.9 m/s which is approximately the speed of a first vertical mode equatorial Kelvin wave. With this time shift many of the peaks in

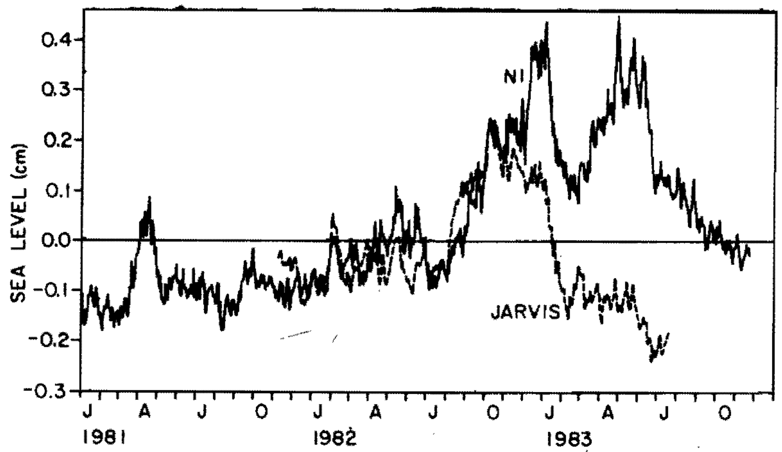

Figure 9. Galápagos sea level (NI) and Jarvis Island sea level delayed by 30 d.

the two records line up. In particular, events in February and May 1982 at the Galápagos are clearly seen 30 d earlier at Jarvis. The early August 1982 sea level rise at NI also corresponds to a dramatic rise at Jarvis which began in early July; however, the rise time in the central Pacific was much shorter. Lukas *et al.* (1984) attribute this difference to modal dispersion. If the sea level signal is composed of equatorial Kelvin waves with different vertical structures, then each mode will propagate eastward with its respective phase speed. The higher modes travel slower and could lead to the change in shape observed. Lukas *et al.* (1984) show that the measured rise time of NI could be reproduced by assuming that the Jarvis record consisted of equal parts mode 1 ($c = 2.8$ m/s) and mode 2 ($c \simeq 1.3$ m/s).

Following the initial sea level rise, the amplitudes of the time adjusted sea levels at Galápagos and Jarvis agreed in about October 1982. Jarvis sea level then began to drop while Galápagos sea level continued to rise. By December the Galápagos change was 25 cm above the Jarvis change. The difference between these two signals increased in 1983 since the second sea level peak at NI was not present at Jarvis. Sea levels, adjusted to agree before and at the onset of the event, disagreed by nearly 50 cm in April 1983. Since the mean sea level increase from Galápagos to Jarvis is only about 35 cm (relative to 1000 db), the sea level slope was likely reversed during this period (Firing *et al.*, 1983). The structure of the sea level records after November 1982 suggest that local wind forcing may be quite important in determining the Galápagos sea level at this time. Indeed, Arkin *et al.*, (1983) show that in early 1983 the zonal wind anomaly had extended quite far into the eastern Pacific and remained significant until at least June 1983. A simple description of the sea level changes in terms of free waves propagating along the equator as discussed in Lukas *et al.* (1984) only appears valid at the start of the event.

Hayes and Halpern (1984) showed that during the period March 1980 to July 1981 sea level at the Galápagos was highly correlated with currents measured 2000 km to the west near 0°, 110°W. Because of the lag between current and sea level (10.5 d) and their relative amplitudes, this correlation was interpreted as further evidence for first vertical mode Kelvin waves. Given the similar interpretation for the onset of El Niño conditions, it is of interest to compare these variables during this event. Figure 10 shows low pass filtered (30 d half power point) sea level at NI and zonal current and temperature at 100 m depth near 0°, 108°W (Halpern, personal communication, 1983). The latter two variables have been delayed by 10.5 d. In March through June sea level,

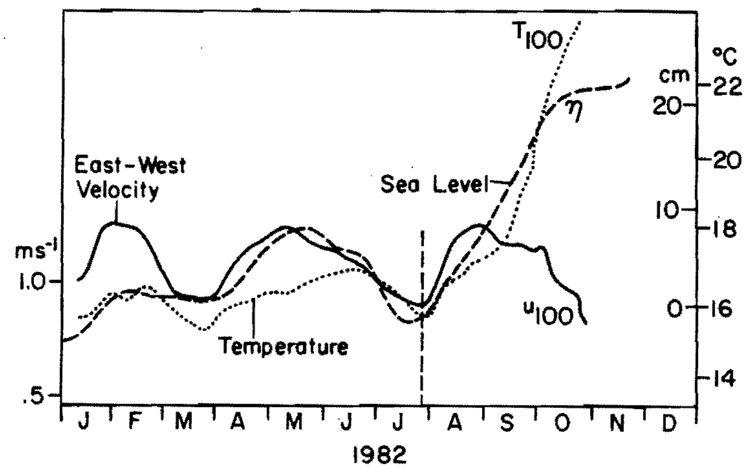

Figure 10. Low pass filtered (30 d half power point) sea level at NI and zonal current and temperature at 100 m depth near 0°, 108°W (Halpern, personal communication, 1983). The 108°W data have been delayed by 10.5 d.

current, and temperature all increased together. The current and sea level changes are consistent with the first vertical mode Kelvin wave interpretation. These records continued to track each other through August 1982 (about one month after the event onset. However, subsequently sea level and temperature continued to increase while zonal velocity decreased. Thus, the simple single vertical mode Kelvin wave model only works reasonably well during the first month after onset. This result is consistent with other observations (Lukas *et al.*, 1984) and model studies (Busalacchi and Cane, 1984) which suggest that other waves become important at about this time. In addition, part of the change in current at 100 m is due to vertical movements of the EUC. Toole and Borges (1984) found that by October the mixed layer at 0°, 110°W had deepened to nearly 100 m. This depth is below the mean position of EUC core (75 m; Halpern, 1984). In November, Hayes and Mangum (1983) found that the EUC core was at 120 m depth. These large changes in the background stratification and currents make interpretation of the last few months of figure 10 difficult.

Comparison of the Galápagos sea level and temperature records through the event shows an interesting feature. The near surface temperature remained relatively constant in January - February 1983 when the sea level dropped nearly 30 cm. As noted in the introduction sea level is a measure of the heat content of the water column (ignoring salinity effects on density). Thus, although surface temperature remained warm, the internal thermal structure must have readjusted to cause a reduction of heat content during this period. Comparison with upper ocean moored temperature measurements at 0°, 95°W (Halpern, 1984) show that this is, indeed, the case. The 15°C isotherm was at its deepest level (greater than 200 m) in December 1982. Water warmer than 25°C extended to 100 m and surface temperatures were 26° to 27°C. In March 1983 the 15°C isotherm rose to about 150 m and the 25°C isotherm was at 50 m. However, surface warming, confined to a thin layer above a near surface thermocline, continued and surface temperature was greater than 29°C. The estimated heat content in the upper 200 m reflected the change in depth of the 25°C to 15°C isotherms. Maximum heat content occurred in December 1982 and it fell rapidly in early 1983. Qualitatively sea level and upper ocean heat content appear reasonably well correlated throughout the event.

The above discussion has focussed on the large scale changes which are coherent throughout the Galápagos Islands, we now turn to details of the pressure distribution around the islands. Ripa and Hayes (1981) used principal component analysis (empirical orthogonal functions) to study the correlated sea level fluctuations of the three gauges on the western side of the islands for the first five months of data (July to November, 1979). They interpreted the results in terms of equatorial waves. The first principal component explained 95% of the variance and had a meridional shape consistent with a first vertical mode equatorial Kelvin wave. A similar analysis was done for the period from 1 January 1982 through 22 July 1983 which includes the warm event. Low pass filtered (30 d half power point) data at the three western gauges were highly correlated during this period (all linear correlation coefficients exceeded 0.98). The first principal component explained 99% of the variance and had amplitudes (in cm) at the three locations of WI: 13.7, NI: 14.2, and SI: 12.8. The time series of this component looked quite similar to the NI sea level record. Fitting the meridional shape to a Gaussian in latitude, and assuming that the sea level response is due solely to a single vertical mode Kelvin wave gives an estimate of the wave phase speed as 7.5 m s^{-1}. This speed is about a factor of three greater than the gravest mode Kelvin wave; higher modes have even smaller speeds. Therefore, the average pressure distribution across the islands is not well represented by a single Kelvin wave mode although for some pieces of the data this simple description still holds.

Recent studies (Eriksen, 1982; Hayes, 1982; Lukas and Firing, 1984) have pointed out the utility of the geostrophic approximation even at the equator. The geostrophic component of the zonal current u is given by

$$fu = \beta yu = -\frac{1}{\rho}\frac{\partial p}{\partial y} \tag{1}$$

where $f = \beta y$ is the Coriolis parameter, y is the north-south coordinate, ρ is the water density, and p is the pressure perturbation. At the equator ($y = 0$) this expression becomes indeterminant if $\partial p/\partial y = 0$. Then the derivative of (1) provides an estimate of current:

$$\beta u = -\frac{1}{\rho}\frac{\partial^2 p}{\partial y^2} \qquad (2)$$

This approximation is discussed in the above references; both Hayes (1982) and Lukas and Firing (1984) had reasonable success in estimating the zonal velocity at the equator using equation 2. Poorest agreement was near the surface where current variability is large and ageostrophic terms are likely to be important.

In spite of the possibility of additional terms in the near surface momentum balance, the three western pressure gauges were used to estimate the time variability of the meridional pressure gradient and the geostrophic current at approximately 15 m depth. With only three measurements a finite difference estimate of $\partial^2 p/\partial y^2$ can be obtained or equivalently the equatorial velocity can be estimated as the average of the velocity north and south of the equator. Computations were on daily values of the 30 d low pass filtered data. Two estimates of zonal velocity are shown in figure 11. The first estimate was obtained from equation 1 using the pressure gradient fluctuations north of the equator between WI and NI. The second estimate used equation 2; finite difference estimates of $\partial^2 p/\partial y^2$ were computed at each time step. These velocity estimates are compared in figure 11 with the measured zonal velocity at 15 m at 0°, 95°W (Halpern, personal communication, 1984). The two pressure gradient estimates appear fairly similar. These estimates show some of the general low frequency pattern of the measured velocity. For example, in February through May 1982 and 1983 both estimated and measured velocity were eastward; in June to September they were westward. The overall correlation coefficients of the estimated and measured velocity were $r_1 = 0.5$ for the first estimate and $r_2 = 0.3$ for the second.

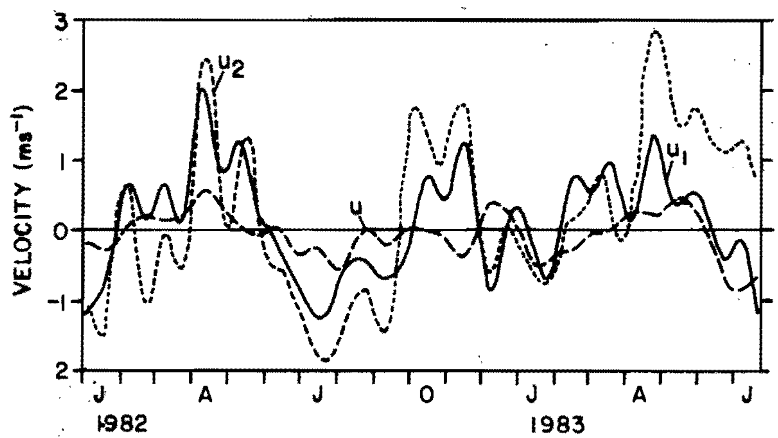

Figure 11. Low pass filtered (30 d half power point) measured zonal velocity (u) at 0°, 95°W (Halpern, personal communication, 1984) and geostrophically estimated velocities using equation 1 (u_1) or equation 2 (u_2) as discussed in the text.

The major discrepancies between computed and measured velocity were the amplitude of the fluctuations and the structure in October - November 1982. Fluctuations of the computed currents were on average a factor of two greater than those in the measured currents. This result implies that equatorial (NI) sea level changes were too large relative to the off equatorial changes. It should be recognized that the sea level differences involved are very small. A 1 cm change in sea level between ±1° and the equator corresponds to a geostrophic velocity of about 70 cm s^{-1}. Given the uncertainties, the remarkable result is that low frequency pressure gradient variations actually were correlated with measured zonal velocity. Many reasons for deviations are possible. These reasons include the pressure field distortions associated with flow around the islands.

Roemmich (1984) reviewed the theory of pressure perturbations caused by high Reynold's number flow around small cylindrical islands. In a uniform, steady flow with speed u_o the pressure difference between the upstream and downstream side of the island can be written as:

$$\frac{\Delta p}{\rho} = 0.5\ (1+\gamma)\ u_o^2 \qquad (3)$$

The form of this expression was chosen to facilitate comparison with laboratory experiments where γ is referred to as the base pressure coefficient. On the upstream side of the island, pressure is elevated relative to its free stream value by $0.5\ u_o^2$ and on the downstream side it is reduced by $0.5\ \gamma u_o^2$. Roemmich found that for Jarvis Island $\gamma = 0.3$ produced a best fit with his velocity data; laboratory studies (Roshko, 1961) yield $\gamma = .85$. The Galápagos Islands do not satisfy the scaling criteria appropriate for this theory. They are sufficiently large that assumptions of uniform flow and small perturbations by the islands are unlikely to be valid. Nevertheless, it is of interest to examine pressure deviations across the archipelago and attempt correlation with u^2 in order to estimate the importance of this effect.

The low pass filtered pressure difference between SI (1°S) on the western side of the islands and SC (0°42'S) on the eastern side is shown in figure 12 for the 18 month period January 1982 to July 1983. The time averaged pressure difference between the two sites has been subtracted. This pressure difference time series has considerable low frequency structure. Positive deviations (~0.03 dbar) occurred in November - December 1982

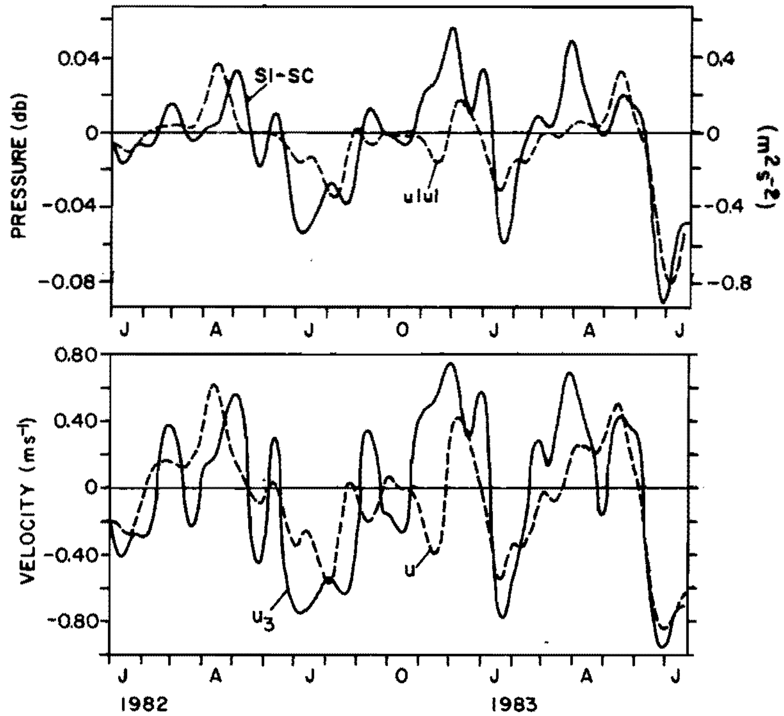

Figure 12. Low pass filtered (30 d half power point) pressure difference between SI and SC and the zonal velocity function u|u| computed from measurements at 0°, 95°W (Halpern, personal communication 1984). The lower frame shows measured (u) and estimated (u_3) zonal velocity using equation 3 with $\gamma = 1.0$.

and negative deviations were found in June to August 1982. For comparison the zonal velocity function u|u| measured at 0°, 95°W is also shown in figure 12. Although this velocity measurement was about 400 km west of the islands, correlation between u|u| and Δp is visually clear. The linear correlation coefficient between these series was $r = 0.71$ (about 50% of the variance explained). Based on the integral time scale of these series there were about 25 degrees of freedom in this correlation. Thus, r is significant at the 99% confidence level. The regression coefficient between $\Delta p/\rho$ and $u|u|$ was 1.0. Comparison with equation 3 yields $\gamma = 1.0$. That this value is close to the laboratory values is probably accidental considering the uncertainties in our estimation of the flow field and the likely errors in applying potential flow theory. Zonal velocity (u_3) estimated from equation 3 with $\gamma = 1.0$ is compared with the measured equatorial velocity in the bottom panel of figure 12. The correlation coefficient of these two series is $r = 0.67$ and the general trends of the measured velocity are fairly well estimated although the computed velocity has more high frequency structure.

Given the significant correlation between cross island pressure difference and zonal velocity, the question arises as to how significant this island drag effect is on our interpretation of low frequency sea level. As noted above, Ripa and Hayes (1981) interpreted the meridional shape of the sea level fluctuations on the western side of the islands as an equatorial Kelvin wave. The amplitude of this signal was about 0.04 m which in a first vertical mode Kelvin wave implies a surface zonal velocity signal of $g\eta/c = .17 \text{ m s}^{-1}$ (assuming $c = 2.3 \text{ m s}^{-1}$). Such a zonal velocity causes an upstream sea level elevation of only 0.003 m; i.e. about 10% of the Kelvin wave amplitude. Thus, the island pressure effect is not a serious contamination. A similar result holds for the velocity-sea level correlations discussed in Hayes and Halpern (1984). In general, since the sea level elevation in a Kelvin wave depends on u and the island pressure perturbation depends on u^2, the latter will dominate at high velocities. For the first vertical mode these two effects yield equal sea level perturbations at an unreasonably high velocity of 4.6 m s^{-1} (again ignoring mean flows). Therefore, the island pressure deviation is likely to be unimportant in many applications; nevertheless, this term cannot always be ignored in quantitative comparisons. Roemmich (1984) has pointed out the potentially important effects which the nonlinearity of this term contribute to high frequency fluctuations.

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