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PROPAGATIONS AND REFELCTIONS OF LONG EQUATORIAL WAVES IN THE TROPICAL PACIFIC OCEAN DURING THE 1986-1989 AND 1992-1993 ENSO EVENTS AS OBSERVED FROM GEOSAT, TOPEX/POSEIDON AND TOGA-TAO DATA

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I. INTRODUCTION

A potentially important mechanism of El Nino/Southern Oscillation is commonly referred to as the delayed action oscillator (Schopf and Suarez, 1988; Battisti, 1988). In this mechanism, long equatorial Rossby waves are generated by the ocean-atmosphere coupled system in the central-eastern Pacific and propagate freely westward in the western Pacific. These waves eventually reflect at the western boundary into eastward propagating Kelvin waves, suggested to be crucial for the termination of warm ENSO events.

Li and Clarke (1994), examining sea-level and wind historical data in the equatorial Pacific, concluded that this mechanism may not be at work. However, Mantua and Battisti (1994) partially revisited the delayed action oscillator mechanism and argued that upwelling Kelvin waves were observed in Li and Clarke's data at the end of each warm event. Therefore, although the delayed action oscillator may not explain the onset of El Nino events, it may actually be at work in the equatorial Pacific to terminate these warm events. In order to examine the validity of such a mechanism in the real ocean, one needs to access high quality synoptic data sets of the equatorial Pacific ocean that encompass ENSO events.

Fortunately, during the TOGA program, the GEOSAT and TOPEX/POSEIDON satellite altimeters were launched and provided the oceanographic community well suited date for the observation and understanding of ENSO events (see Picaut *et al.*, 1995). Besides, a major success of the TOGA program was the deployment of the TOGA-TAO array (Figure 1). therefore, we now possess those remotely-sensed and *in situ* synoptic data to examine, in the real ocean, the potential wave propagation and western boundary reflection hypothesized by the delayed action oscillator theory.

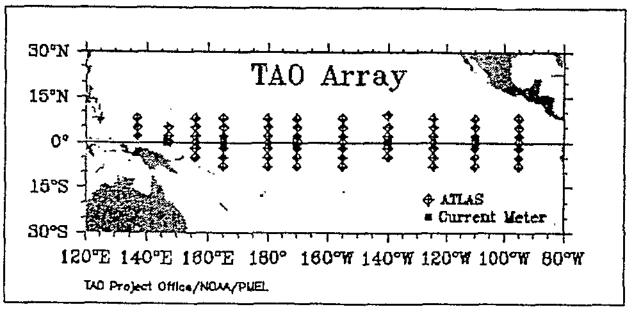

Figure 1: The TOGA-TAO array in its final configuration.

II. LONG EQUATORIAL WAVES IN GEOSAT DATA

GEOSAT sea-level data were first processed and validated. Geostrophic zonal current anomalies were then derived and compared fairly well with *in situ* measurements in the western and central Pacific ocean. Then, GEOSAT data (sea-level and geostrophic zonal current anomalies, restricted to 160°E-90°W/28°N-28°S) as well as the FSU wind were projected onto long equatorial waves (Delcroix *et al.*, 1994). As shown in Figure 2, Kelvin and first meridional Rossby waves are observed propagating at theoretical phase speeds throughout the equatorial Pacific ocean during the November 1986-February 1989 ENSO period. The first Rossby mode forcing is mostly a mirror of the Kelvin mode forcing and is therefore not shown here. Both Kelvin and Rossby coefficients are strongly related to their corresponding forcing coefficients (see Figure 2 for the Kelvin mode) which indicates that wind forcing is a major component of wave triggering.

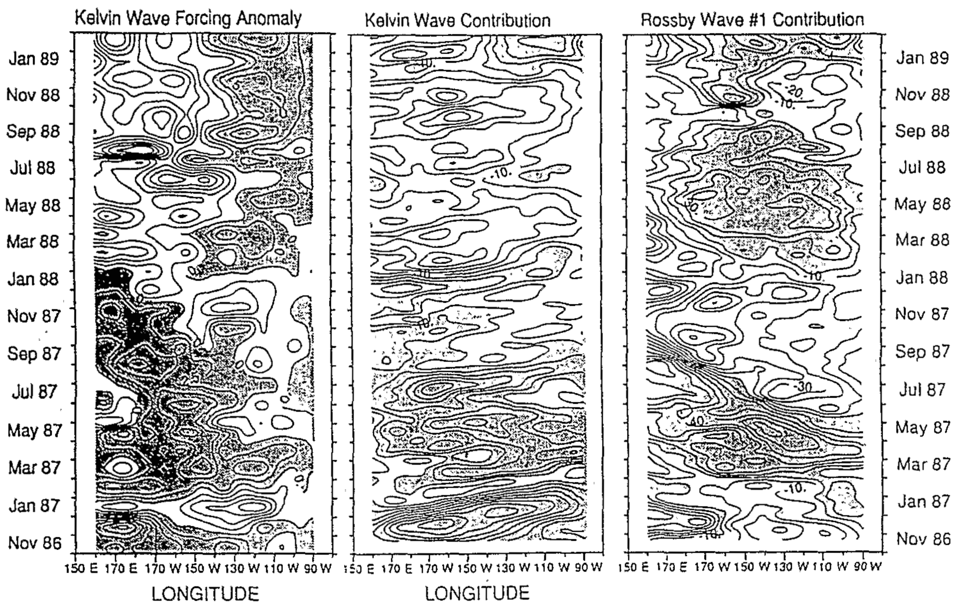

Figure 2: Left panel: Longitude-time plot of the projection of the zonal wind anomalies onto the Kelvin mode. Units are non-dimensionalised. Note that 2.5 units roughly correspond to 10 m^2/s^2 zonal pseudo stress at the equator. Middle and right panels are longitude-time plots along the equator of zonal current anomaly contributions of respectively the Kelvin and first meridional Rossby modes. Units are cm/s and shaded areas denote eastward current anomalies. Contour intervals are 10cm/s.

To investigate the role of western boundary reflection versus wind forcing for wave generation in the western Pacific, SSM/I and FSU Kelvin wind forcing averaged over 130°E-160°E are compared to the oceanic Kelvin coefficient calculated from GEOSAT data at 160°E (Figure 3). This comparison suggests that wind forcing, rather than reflection of long Rossby waves at the western boundary, is the main trigger of Kelvin waves during the 1986-1989 El Niño-La Niña period which does not favor the dynamics involved in the hypothesized delayed action mechanism (Battisti, 1988).

To extend the wave propagation and reflection results to the second 1992-1993 warming, the TOPEX/POSEIDON period is now examined.

III. LONG EQUATORIAL WAVE PROPAGATION AND REFLECTION IN TOPEX AND TOGA-TAO DATA

To investigate long wave propagation and reflection at both eastern and western boundaries, high quality data over the entire basin are first required. Thus, we used TOPEX sea-level and TOGA-TAO dynamic height data during the November 1992-December 1993, validated by Menkes *et al.* (1995a) in the equatorial band.

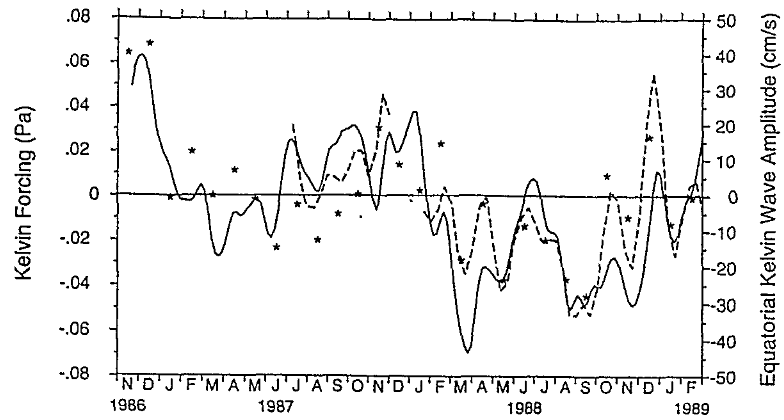

Figure 3: Time series of 130°E-160°E averaged SSM/I (dashed line) and FSU (stars) Kelvin forcing and oceanic Kelvin coefficient wave contribution to zonal current anomalies at the equator and at 160°E (solid line).

The projection method used previously (Delcroix *et al.* ,1994) requires the knowledge of both sea-level and derived geostrophic zonal current anomalies. Moreover this method is useful if the meridional extension of the data is large compared to the turning latitude of the wave coefficient to be calculated (*e.g.* 5°, 6.5° and 8° respectively for the first, second and third Rossby waves). This condition does not apply in the western Pacific where lands (Papua-New Guinea) are close to the equator, nor is it applicable to the TOGA-TAO data in the entire basin as the TAO array domain does not extend poleward of 8°N-8°S. Therefore a new method is developed to project TOPEX/POSEIDON sea-level and TOGA-TAO dynamic height fields onto long equatorial waves throughout the whole basin and to study wave propagation/reflection even in the western Pacific where land is close to the equator (Boulanger and Menkes, 1995). Furthermore, in order to investigate relationships between oceanic wave generation and wind forcing, a similar method is developed to project a gridded field of TAO wind stress (Menkes *et al.*, 1995b) onto wave structures throughout the entire basin. Finally, this method is used to study wave propagations and generations that may partly be due to reflections at the maritime boundaries and/or to wind forcing.

Wave propagation: The Kelvin and the first three meridional long Rossby modes are observed to propagate throughout the basin in both TOPEX and TOGA-TAO data at theoretical phase speeds characteristic of the first baroclinic mode (Boulanger and Menkes, 1995). For example, the third meridional Rossby wave coefficient is shown on Figure 4 (see also next paragraph). Although its propagation is less clear in TOGA-TAO data than in TOPEX data (due to the coarser meridional resolution of the TOGA-TAO array), this result confirms that long equatorial waves do propagate individually throughout the equatorial Pacific ocean.

Wave reflection: Kelvin and Rossby wave coefficients calculated from TOPEX sea-level data are represented on Figure 5.

Eastern boundary reflection: We first consider eastern boundary reflection by focusing on the left and middle panels. Theoretically, a downwelling incoming Kelvin wave would reflect into a downwelling first Rossby wave. As shown in Figure 5 (a and b), there is not any clear wave agreement between Kelvin and Rossby coefficients and therefore reflections at the eastern boundary during the 1992-1993 El Niño event.

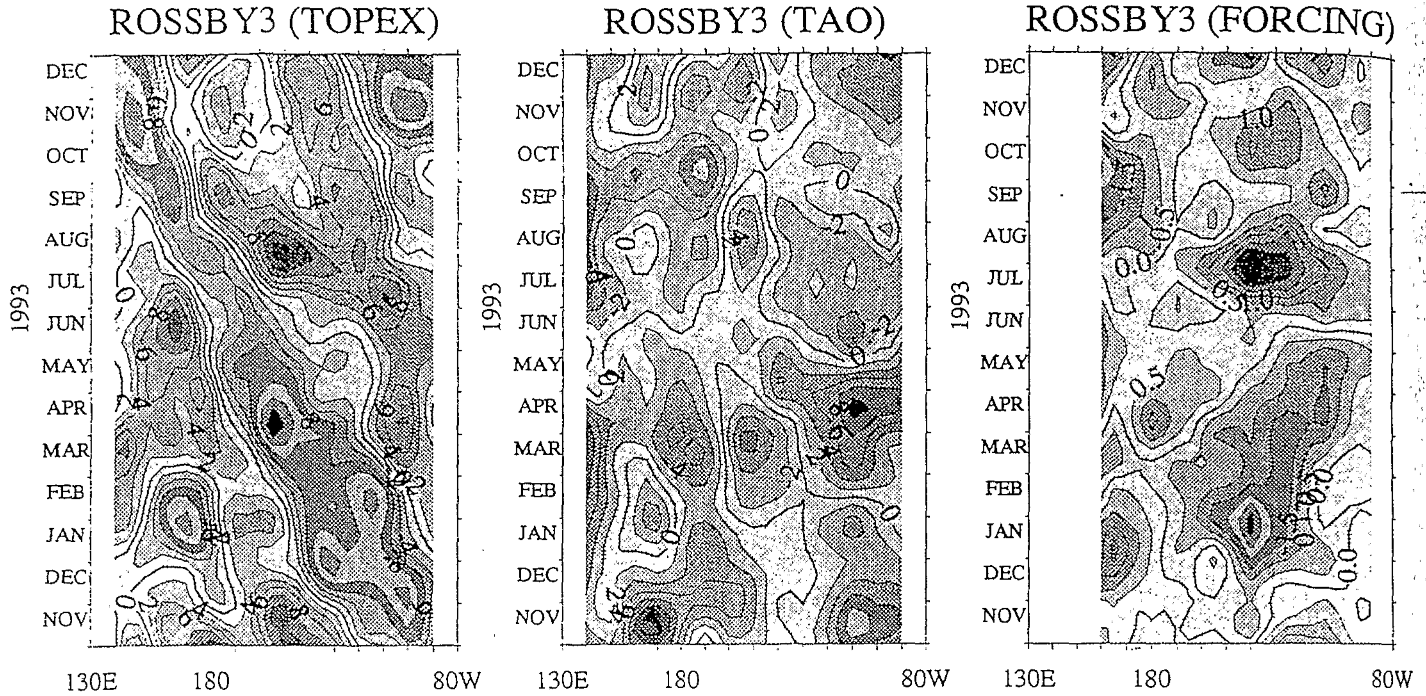

Figure 4: Longitude-time plots of (a: left) TOPEX third meridional Rossby coefficient, (b: middle) TOGA-TAO third meridional Rossby coefficient, (c: right) third meridional Rossby forcing calculated from TOGA-TAO zonal wind stress anomalies (Menkes *et al.*, 1995b). Note that 2 units roughly represent an amplitude of 0.8cm for the oceanic coefficient at 7°N (where the third meridonal Rossby structure has a maximum), and that 1 unit, an amplitude of 0.015 dyn/cm^2 at the equator.

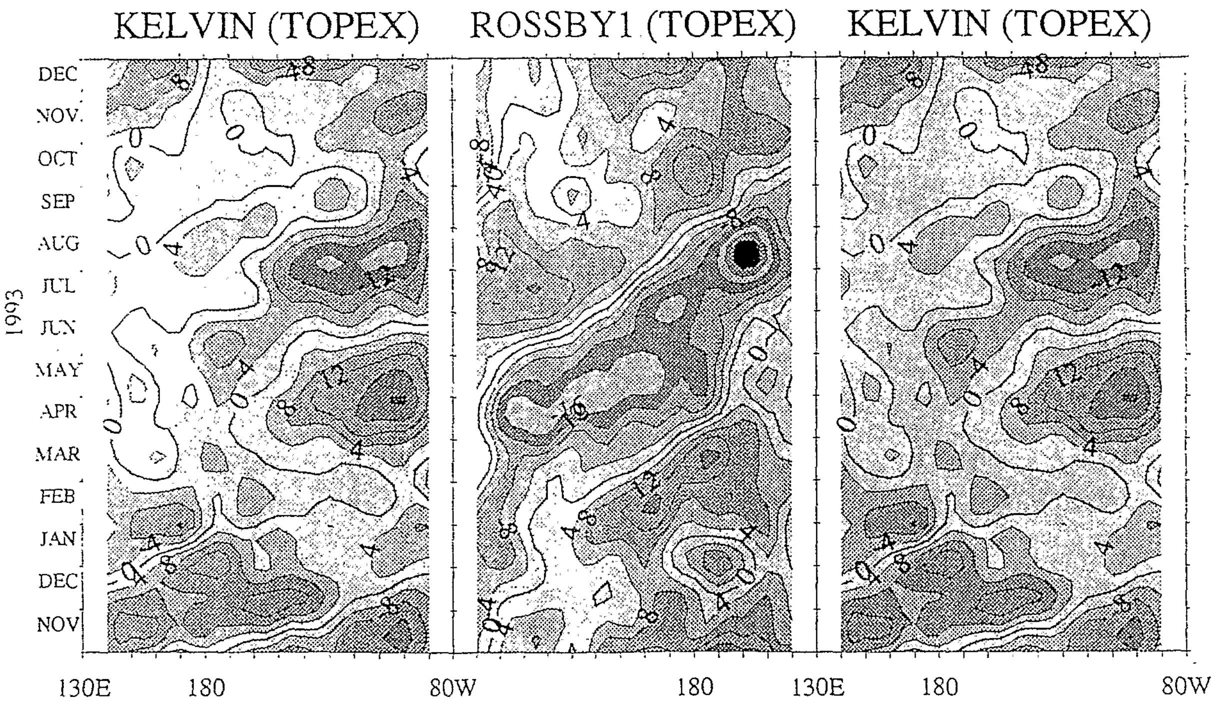

Figure 5: Longitude-time plots, from 130°E to 80°W, of TOPEX Kelvin coefficient (left and right panels) and (middle panel) TOPEX first meridional Rossby coefficient (plotted in reverse from 80°W to 130°E). Coefficients are adimensionalised. Note that 2 units roughly represent an equatorial sea-level amplitude of 1cm for the Kelvin wave and 0.4cm for the first Rossby mode, and an equatorial zonal current amplitude of 3cm/s for the Kelvin wave and 4cm/s for the Rossby wave.

Western boundary reflection: Consider now western boundary reflection from left and right panels. Theoretically, both Kelvin and first Rossby wave coefficients should be in phase, but the amplitude of the reflected Kelvin wave should be smaller than the incoming Rossby (a reflected amplitude of 0.4 for an incoming amplitude of 1). Yet, here, there is not any clear coherence between both coefficients on Figure 5 (b and c). A low amplitude upwelling Kelvin wave is seen propagating throughout the basin during the September-October 1993 period simultaneously to the termination of the warm event. This wave follows the arrival of an upwelling Rossby wave generated in the eastern Pacific in March-April 1993 (Figure 5). However, the upwelling Kelvin wave is observed to be linked to a strong easterly anomaly occuring in the western Pacific near 140°E-150°E in September 1993 (Boulanger and Menkes, 1995). Thus, Kelvin waves propagating toward the central Pacific do not seem to come from first meridional Rossby wave reflection at the western boundary.

To further estimate if Kelvin wave coefficients were preferentially forced by wind forcing rather than by Rossby wave reflections, we compared Kelvin wave amplitude at 160°E calculated from TOPEX sea-level data, the 140°E-160°E averaged Kelvin forcing and a Kelvin wave coefficient at 160°E calculated from the total reflection of an hypothetical Rossby wave reflection at 130°E. Our result, similar to the one obtained during the GEOSAT period, strongly suggests that wind forcing, rather than western boundary reflections, is the main trigger of Kelvin waves propagating to the central Pacific.

III. CONCLUSION

Both GEOSAT, TOPEX/POSEIDON and *in situ* TOGA-TAO data are well suited to investigate theoretical ENSO mechanisms in the real ocean. Projection of remotely-sensed sea-level and in-situ dynamic height data onto long equatorial waves over the entire basin (*i.e.* even in the western Pacific ocean where lands are close to the equator) allows to investigate for wave propagation and reflection mechanisms critical for the hypothesized delayed action oscillator.

Observation of individual wave propagations both during the 1986-1989 and 1992-1993 events, prove that the low-frequency linear dynamics of the equatorial Pacific ocean may be interpreted in terms of individual wave propagation. During both ENSO periods, and contrary to the delayed oscillator mechanism, our results strongly suggests that wind forcing, rather than Rossby wave reflection at the western boundary, is the main trigger of Kelvin waves propagating to the central Pacific.

However, a wind-forced upwelling Kelvin wave is observed to propagate throughout the Pacific ocean simultaneously with the termination of the unusually long-lasted 1991-1993 warm ENSO event. Although this wave is not generated by a Rossby wave reflection, its propagation is coherent with some features of the most recent works on the delayed action oscillator (Mantua and Battisti, 1994). Yet, it is not clear whether this upwelling Kelvin wave propagation is fortuitous or plays an active role in this warm event termination.

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