Reflection of Annual Rossby Waves at The Maritime Western Boundary of the Tropical Pacific

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Baroclinic Rossby wave activity in the tropical Pacific was first observed by White (1977), who examined the annual cycle of subsurface thermal structure in the eastern region from 10°-20°N, finding westward propagating annual signals emanating from the eastern boundary, propagating eastward at baroclinic Rossby wave speeds to the longitude of the Hawaiian Archipelago. This work was followed by Meyers (1979), who observed westward propagation of the annual cycle in upper ocean thermal structure at baroclinic Rossby wave speeds near 6°N all across the Pacific. Subsequently, White (1983) examined interannual variability in the subsurface thermal structure in the western North Pacific, finding westward propagating signals traveling at baroclinic Rossby wave speeds from 6°-30°N. These interannual waves were associated with ENSO activity. Soon thereafter, interannual baroclinic Rossby wave activity in the latitude bands from 5°-15°N and from 5°-15°S was found to act as a precursor (e.g., White et al., 1985a; Pazan et al., 1986) to the onset of the 1982-1983 El Nino activity in the eastern equatorial Pacific; indirect evidence was found that the incidence of this baroclinic Rossby wave activity in these latitude bands was instigating baroclinic Kelvin wave activity in the equatorial wave-guide. Subsequently, White et al. (1987) used this information to demonstrate that both the observation and model simulation of baroclinic Rossby wave activity in the western tropical Pacific could be used to hindcast El Nino activity in the eastern equatorial Pacific for up to a year in advance over the 25-year period from 1962-1987.

The dynamical influence that baroclinic Rossby wave activity has upon in the maintenance of the quasi-cyclic behavior of ENSO (El Nino/Southern Oscillation) was discussed by Graham and White (1988). They established using a conceptual model that baroclinic Rossby wave activity influenced ocean/atmosphere coupling in the equatorial wave-guide by providing for a steady upwelling/downwelling (lasting from 3-9 months) through the excitation of upwelling/downwelling equatorial baroclinic Kelvin wave activity at the western boundary (via the Rossby wave reflection process). During the year prior to El Nino activity in the eastern equatorial Pacific, downwelling equatorial Kelvin wave activity was associated with a steady eastward advection of warm water from the western equatorial Pacific into the central and eastern equatorial ocean, advected by the associated zonal eastward current; this eastward displacement of warm SST's on the equator was hypothesized to instigate a coupled unstable wave in the ocean/atmospheric system along the equator, of the kind discussed by Schopf and Suarez (1988). Within this conceptual model, the reflection of baroclinic Rossby waves at the western boundary of the tropical Pacific also played an important role in turning the El Nino process off, leading to La Nina, as discussed in Graham and White (1988).

Sources of baroclinic Rossby wave activity in the tropical Pacific are fairly well known. White (1977) found evidence consistent with the idea that baroclinic Rossby waves were being generated at the eastern boundary through the action of the wind stress

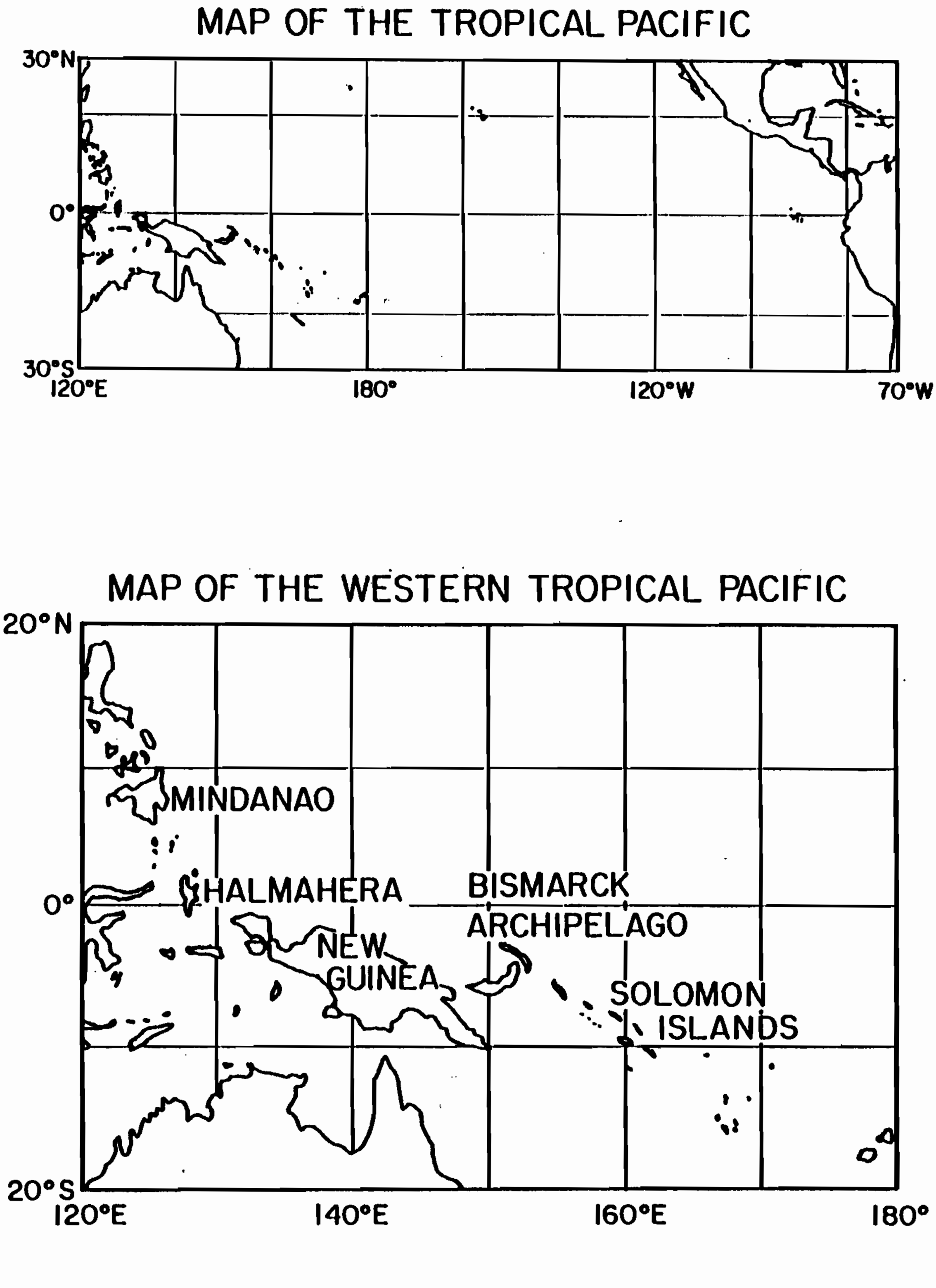

FIG.1. (Upper Panel): Map of the tropical Pacific from 30°N to 30°S where this study was conducted. (Lower Panel): Map of the western tropical Pacific, where the reflection of baroclinic Rossby waves by the maritime western boundary is emphasized.

curl. Meyers (1979) found baroclinic Rossby waves of annual period in the western and central tropical North Pacific to have been generated by the Ekman pumping in the Trade Wind field in the eastern tropical North Pacific. Later, White et al (1985b) demonstrated that the interannual baroclinic Rossby wave activity in the western North Pacific was wind-generated in response to Ekman pumping in the western and central tropical Pacific. Moreover, it was shown that baroclinic Rossby wave activity increased in amplitude toward the west in response to resonance with the wind stress curl forcing. Pazan and White (1987) computed a vorticity/volume budget for the western tropical Pacific from 4°-16°N, 160°E-160°W, finding the thermocline to have been pumped vertically in response both to the divergence of Ekman transport in the upper layer (i.e., indicative of Ekman pumping) and to the divergence of geostrophic transport in the region (i.e., indicative of Rossby wave propagation). More recently, White et al. (1989) demonstrated that it was possible to track interannual baroclinic Rossby wave activity all the way across the ocean from the eastern boundary to the western boundary in both observations and the models at approximately 12°N; regardless, Graham et al. (1989) found that most of the magnitude in this wave activity occurred in response to wind stress curl forcing in the central and western tropical Pacific, amplified by resonant forcing.

The reflection of baroclinic Rossby wave activity at the maritime western boundary of the tropical Pacific has been inferred from the statistical correlation between anomalous upper layer thickness in the off-equatorial western tropical Pacific (associated with baroclinic Rossby wave activity) and that in the central and eastern equatorial Pacific (associated with baroclinic Kelvin wave activity) in both models and observations (Pazan et al., 1986). Still, no direct observational evidence exists for the reflection of baroclinic Rossby wave activity at this western boundary and the subsequent generation of equatorial Kelvin wave activity emanating from it. A certain amount of skepticism surrounds the reflection process at the maritime western boundary (i.e., formed by the Philippines Archipelago in the northern hemisphere and possibly the Solomon Archipelago in the southern hemisphere, as shown in Fig. 1) because of the presence of gaps in the numerous archipelagoes that could conceivably constitute the maritime western boundary of the tropical Pacific. However, recently Clark (1989) and du Penhoat and Cane (1989) established that these particular archipelagoes were quite good reflectors of interannual Rossby wave activity, reflecting 70-80% of the incident baroclinic Rossby wave activity as baroclinic equatorial Kelvin waves.

In this study, the examination of GEOSAT altimetric sea level differences provides the first direct evidence of this reflection process operating at the maritime western boundary of the tropical Pacific in both the northern and southern hemisphere (see Fig.1). In an earlier study, Tai et al. (1989) analyzed altimetric sea level crossover differences from the first 17 months of the GEOSAT mission from November 1985-April 1987, finding statistically significant agreement with *in situ* measurements of sea level differences (i.e., island sea level and relative dynamic height). In that study the dominant time scale of variability that could be studied was the annual cycle; interannual variability could not be addressed because of the relatively short length of the record. So too in this study; only the wave reflection at the maritime western boundary of the tropical Pacific associated with the annual cycle is addressed.

In the earlier study of Tai et al. (1989), a time series of maps of altimetric sea level residuals in the tropical Pacific (17 months from April 23, 1985 to September 8, 1986, separated into thirty 17-day time steps) was generated. Verification of these maps was conducted by Tai et al., (1989) from the comparison of the maps of altimetric sea level residuals to similar maps of relative dynamic height (0/400 db) residuals, and time sequences of altimetric sea level residuals to similar time sequences of sea level residuals measured at island stations. Examples of these 17-day maps is given in Fig.2,

FEATURE TRACKING OF ALTIMETRIC SEA LEVEL
IN THE WESTERN NORTH PACIFIC

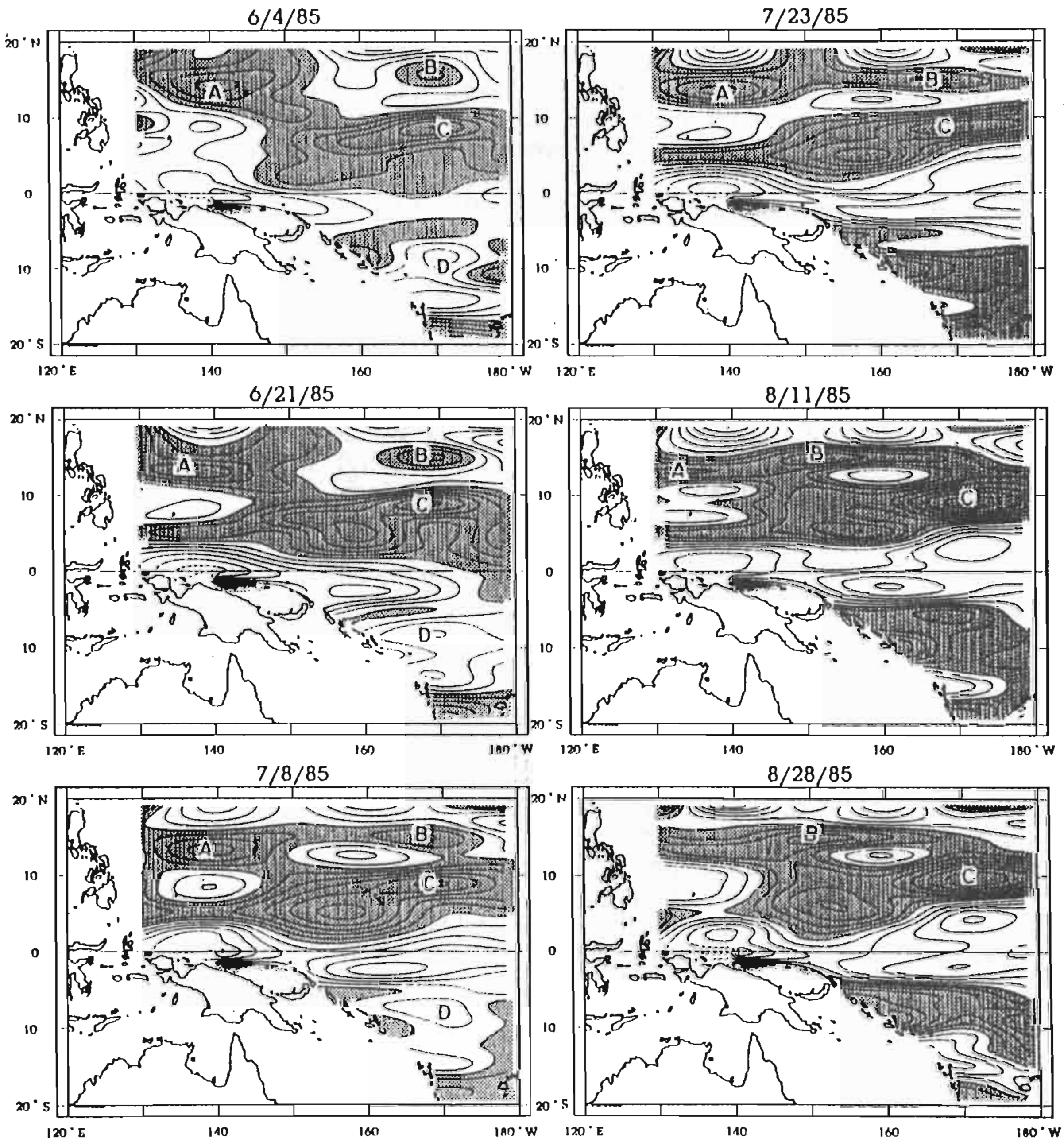

FIG.2. Altimetric sea level maps for June-August 1985. Features evolve smoothly and consistently through this series of maps (CI = 2cm, negative values are shaded).

concentrating on the western tropical Pacific, repeated from Tai et al. (1989). Notice that in the maps the Philippines Archipelago forms the maritime western boundary in the northern hemisphere from the equator to nearly 20°N, while the Solomon Archipelago/New Guinea complex forms the maritime western boundary in the southern hemisphere from the equator to approximately 10°S. Notice too that the sea level residual features labeled alphabetically in Fig.2 can be traced from map to map, often indicating (particularly A and B) westward propagation into the maritime western boundary of the tropical Pacific.

In order to extract the information that this time series of maps of altimetric sea level residuals has concerning the reflection of baroclinic Rossby waves from the maritime western boundary of the tropical Pacific, extended empirical orthogonal function (EEOF) analysis (Graham et al., 1987) was conducted upon the thirty-one 17-day maps of altimetric sea level differences about the mean for the 17-month period represented (i.e., from November, 1985-April, 1987). One EEOF analysis was conducted upon the entire field of altimetric sea level differences from South America to the maritime coast of Asia, allowing the evolution of the basin-scale redistribution of mass to be examined (Fig.4). Then, in order to focus on the reflection activity at the maritime western boundary of the tropical Pacific, the EEOF analysis was repeated upon the western tropical Pacific, from 20°S-20°N west of 160°W (Fig.5).

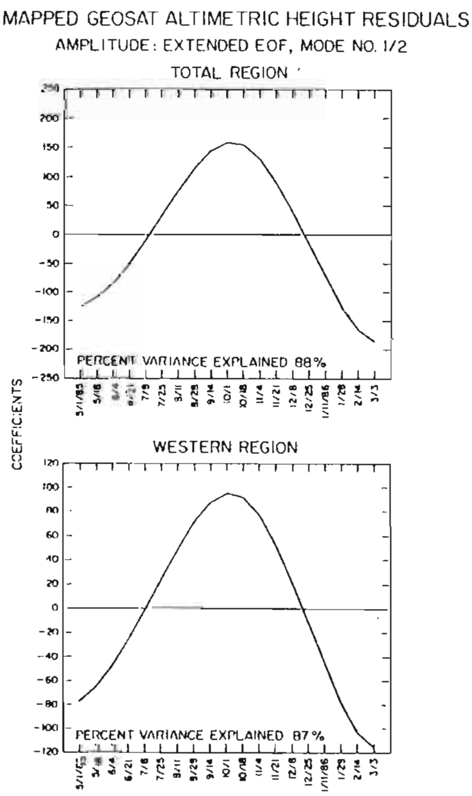

FIG.3. Time sequences for the fist extended empirical orthogonal functions of altimetric sea level residual computed for the tropical Pacific (Upper Panel) and the western tropical Pacific (Lower Panel),explaining 88% and 87% of the total variance in the 17-month time sequence. The evolutionary spatial patterns of these functions are given in Figs. 5 and 6.

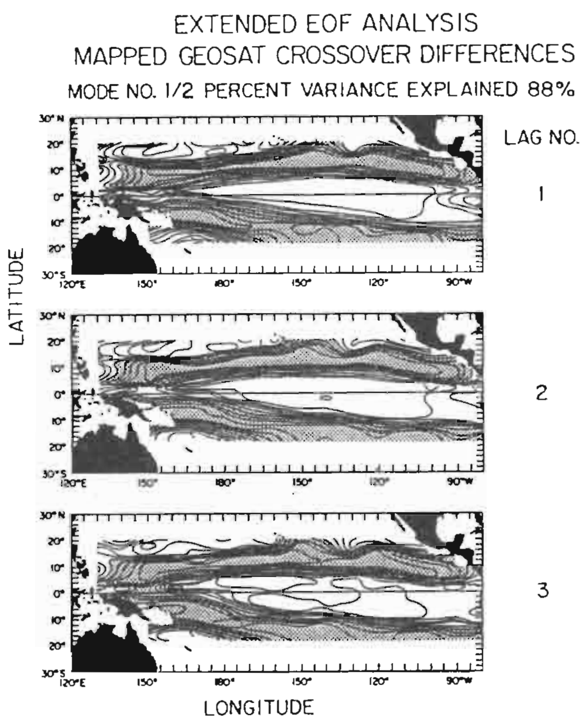

FIG.4.a Time sequence of the first three spatial patterns (1-3) of the first extended empirical orthogonal function for the tropical Pacific, the latter explaining 87% of the total variance in the 17-month time sequence. Shading indicates negative phase.

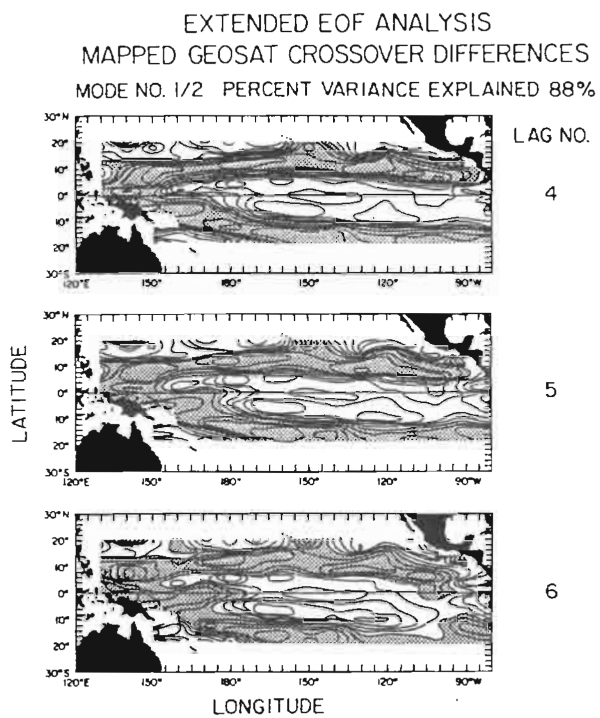

FIG. 4b. Time sequence of the second three spatial patterns (4-6) of the first extended empirical orthogonal function for the tropical Pacific, the latter explaining 87% of the total variance in the 17-month time sequence. Shading indicates negative phase.

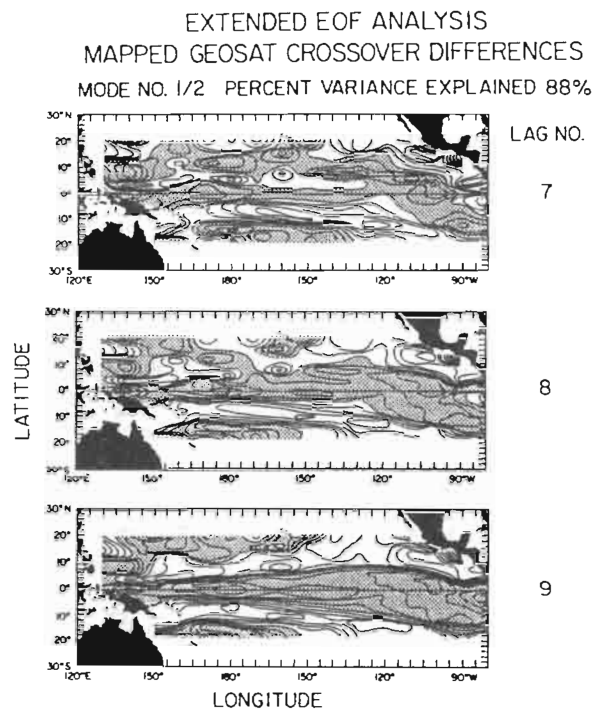

FIG. 4c. Time sequence of the third three spatial patterns (7-9) of the first extended empirical orthogonal function for the tropical Pacific, the latter explaining 87% of the total variance in the 17-month time sequence. Shading indicates negative phase.

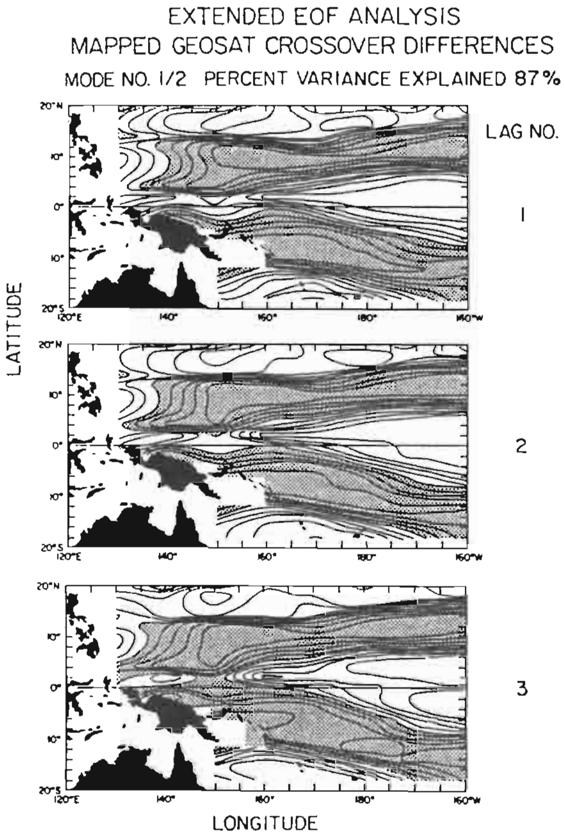

FIG. 5a. Time sequence of the three spatial patterns (1-3) of the first extended empirical orthogonal function for the western tropical Pacific, the latter explaining 88% of the total variance in the 17-month time sequence. Shading indicates negative phase.

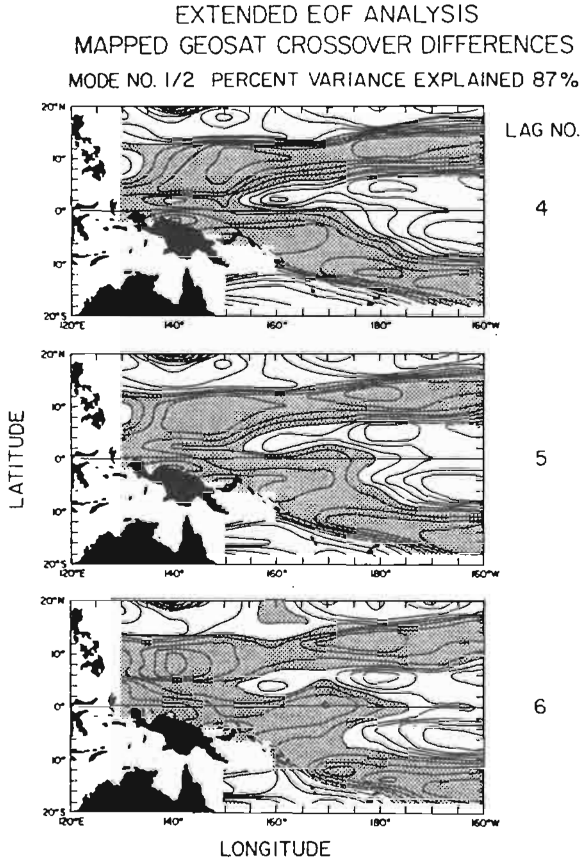

FIG. 5b. Time sequence of the second three spatial patterns (4-6) of the first extended empirical orthogonal function for the western tropical Pacific, the latter explaining 88% of the total variance in the 17-month time sequence. Shading indicates negative phase.

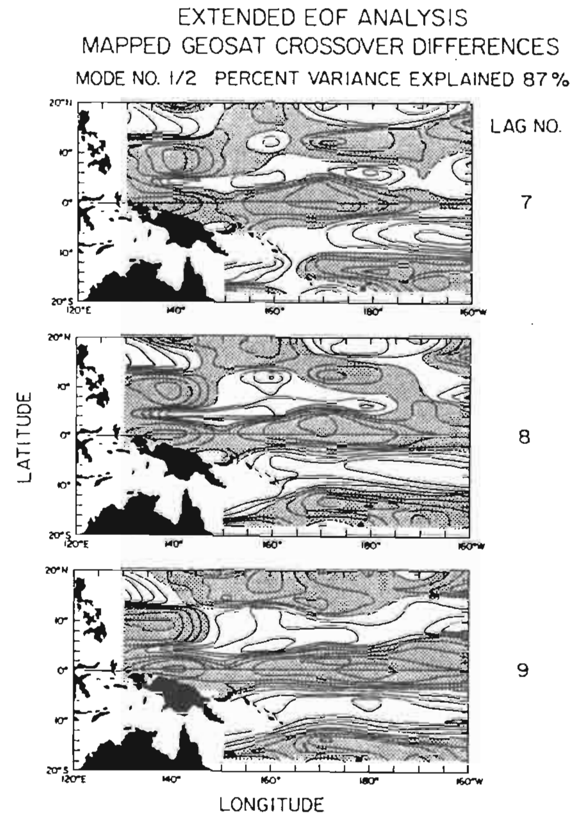

FIG. 5c. Time sequence of the third three spatial pattern (7-9) of the first extended empirical orthogonal function for the western tropical Pacific; the latter explaining 88% of the total variance in the 17-month time sequence. Shading indicates negative phase.

The time sequences of both of these EEOF's is given in Fig.3, with that for the entire tropical Pacific shown in the upper panel, and that for the western Pacific shown in the lower panel. Each are taken from the first two EEOF's, which together account for 88% and 87% of the total variance, respectively, in the tropical Pacific and the western tropical Pacific. The first two EEOF's tend to be redundant and, being orthogonal, give different phases of the same evolutionary behavior in the time sequence of maps (Graham et al., 1987). In this case, the time series of the two analyses are nearly identical, and strongly dominated by the annual cycle.

The EEOF on the basin scale maps is given in Fig. 4a, b, c, showing the evolution of the annual cycle through approximately half a cycle (i.e., 6 months), extending over nine 17-day time periods (i.e., 153 days). The time sequence of nine maps that constitutes the EEOF shows how the spatial pattern in Map 1 evolved into the opposite phase in Map 9 through the spring/autumn transition. In this series of maps, the latitude of approximately 8° marked a boundary where the phase of the annual cycle on the poleward side was opposite that on the equatorward side. This was also found to be true in the complex empirical orthogonal function (CFOF) analysis in Tai et al., (1989). In the present study, by following the evolution of the annual cycle through half a cycle with the aid of the EEOF analysis, from positive residual on the equator (i.e., Map 1) to negative residual on the equator (Map 9), this phase transition from winter to summer (and vice versa) can be understood.

In the first three maps of Fig.4a, the changes that took place in this phase transition are the following. In the off-equatorial region poleward of approximately 8° latitude, the negative residual propagated to the west, evident particularly in the western tropical Pacific. In the western equatorial Pacific, west of 160°E, the negative anomaly began to collapse onto the equator, more from the southern hemisphere north of the Solomon Archipelago/New Guinea complex than from the northern hemisphere. In the northern hemisphere, positive residuals along the coast of the Philippines Archipelago extended southward to the equator in Map 1 and Map 2, where they protruded eastward along the equator; this terminated in Map 3.

In the next three maps (i.e., Maps 4-6 in Fig. 4b), the transition in phase of the annual cycle found negative residuals on the equator in the western Pacific emanating from the Solomon Archipelago/New Guinea complex in the southern hemisphere. Moreover, these negative residuals began to extend eastward along the equator into the eastern equatorial Pacific forming a trough along the equator. In Map 4, this eastward extension of the trough along the equator seems to have originated from the southern hemisphere, but by Map 6 the trough seems to have originated from both hemispheres. At the eastern boundary, an interesting phenomenon began to occur in this sequence of three maps; apparent first-mode equatorial Rossby wave activity appears to have reflected from the eastern boundary (i.e., with local positive maxima at 4°N and 4°S) in response to incident Kelvin wave activity, associated with positive residuals extending northward along the coast of Central America. However, over these three maps (and, as we shall see, over the next three maps) this apparent first-mode equatorial Rossby wave activity did not propagate westward of the Galapagos Archipelago.

In the next three maps (i.e., Maps 7-9 in Fig. 4c), the autumn phase transition from positive to negative sea level residuals along the equator was completed. This occurred with the eastward propagation of negative sea level residuals, initially (Map 7) extending from the west along the equator and confined to the equatorial wave-guide, but later (Map 9) filling the entire region equatorward of approximately 8° latitude, apparently in response to wind stress forcing. In the northern hemisphere western Pacific, negative residuals can be seen to have propagated westward onto the Philippine Archipelago between 5°-15°N, extending equatorward along the maritime western boundary from 15°N to the equator, with negative residuals extending eastward along the equator from the maritime western boundary of the tropical Pacific. This is indicative of the reflection of incident baroclinic Rossby wave activity and the excitation of baroclinic Kelvin wave activity. In the southern hemisphere western Pacific, positive residuals in the western Pacific began to be found along the Solomon Archipelago, nudging equatorward to begin the process of another phase transition (i.e., this time the spring transition) from the present negative equatorial state to a positive one on the equator.

The EEOF of the western tropical Pacific (i.e., 160°W to the maritime coast of Asia) is given in Fig. 5a, b, c, showing the evolution of the annual cycle through half a cycle (i.e., 6 months), extending over nine 17-day period (i.e. 170 days). It shows in detail the western boundary reflection of the annual baroclinic Rossby wave activity, associated with phase transition of the annual cycle over the entire tropical Pacific shown in Fig. 4a, b, c, and the subsequent Kelvin wave excitation in the equatorial wave-guide.

In the first three maps of Fig. 5a, the changes that took place in the western boundary, associated with the phase transition of the annual cycle from positive to negative residual on the equator, were able to be observed. In the off-equatorial region of the southern hemisphere (i.e., poleward of equatorial wave-guide, the latter extending to approximately 3° latitude), the negative residuals can be seen to have propagated to the west, clearly riding up onto the Solomon Archipelago/New Guinea complex from 160°-150°E, penetrating as far west as 130°E. Over this approximately 1 1/2 month period, the negative residuals also penetrated equatorward and intersected the equator in a relatively broad longitudinal band equatorward of the Solomon Archipelago and New Guinea. Poleward of the equator wave-guide in the northern hemisphere, positive residuals along the coast of the Philippines Archipelago (i.e. from 3°-14°N) extended southward to the equator, where they protruded eastward along the equator, indicative of the excitement of equatorial Kelvin wave activity by the coastal Kelvin-Munk wave activity (Godfrey, 1976). It is important to remember that each altimetric sea level datum along 130°E is a 10° longitude mean, so the information at the longitude extends from 125°-135°, representing information in sea level variability next to the Philippine coast at 123°-127°E, but smearing it with information eastward of there. East of the positive residuals adjacent to the Philippines Archipelago, negative residuals were observed to extend all across the tropical Pacific from 4°-12°N; in Fig. 5a these negative residual can be observed to have propagated to the west over the three maps shown penetrating over the two month period to the maritime western boundary from 2°-6°N and extending toward the equator north of New Guinea.

In the next three maps (i.e. Maps 4-6 in fig. 5b), the transition in annual phase finds the negative residual that had penetrated to the equator north of Solomon Archipelago/New Guinea complex in the western Pacific (over the previous 1 1/2 month) extending eastward along the equator, forming a trough from 130°E-160°W. In Map 4, the eastward extension of this trough along the equator originated from the southern hemisphere, but in Maps 5 and 6 the trough began to originate from both hemispheres, with the contributions from both hemispheres clearly visible. The earlier contribution from the southern hemisphere seemed to have occurred because the maritime western boundary in the southern hemisphere was located 20°-30° of longitude east of that in the northern hemisphere, coupled with the fact that the negative residuals in the off-equatorial region as a whole were relatively symmetric about the equator. The lag of the northern hemisphere behind the southern hemisphere was only 1-2 months. In the last map (i.e., Map 6) of this series of three in Fig. 5b, the off-equatorial negative residuals in the northern hemisphere penetrated to the maritime western boundary at approximately 10°N, with maximum values continuing equatorward along the boundary and extending east along the equator.

In the next three maps (i.e., Maps 7-9 in Fig. 5c) the autumn phase transition in the annual cycle from positive to negative residuals along the equator was completed, with the equatorial wave-guide filled with negative residuals emanating principally from the maritime western boundary in the northern hemisphere from 5-15°N. During these times, the southern hemisphere had positive residuals in the off-equatorial region from 5°-15°S, nudging equatorward to begin the reciprocal process of the spring phase transition from the present negative equatorial state to positive; not so in the northern hemisphere where an intense negative anomaly from 5°-15°N continued to supply off-equatorial influence to the equator via the maritime western boundary, helping to maintain the negative residuals in the equatorial wave-guide. In this maritime western boundary of the northern hemisphere, regional maxima in negative residuals were found directly east of Mindanao near 8°N and east of Halmera near 2°N; smaller values were found in between where a gap occurs in the maritime western boundary. Yet still, indications are that the negative residuals bridged this gap, particularly evident in Maps 6-9 in Fig. 5.

An interesting side-light of this examination of the baroclinic Rossby wave reflection occurring at the maritime western boundary of the tropical Pacific is the reflection (or, rather, the lack thereof) of the incident equatorial kelvin wave activity at the eastern boundary of the tropical Pacific (i.e., South America) that can be seen in Fig. 4a, b, c. The incident annual equatorial baroclinic Kelvin wave was observed to have generated coastal Kelvin wave activity (i.e., traveling poleward) and equatorially trapped Rossby wave activity (i.e., traveling westward). It is observed that the equatorially trapped Rossby wave activity, thus generated, did not propagate westward past the Galapagos Archipelago.

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