ENSO Seasonality: 1950–78 versus 1979–92*

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ABSTRACT

ENSO-related seasonal-to-interannual variability in the Pacific basin is documented, based on marine surface observations of monthly mean sea surface temperature, sea level pressure, and wind, together with satellite-based estimates of rainfall and mean tropospheric temperature. Anomalies in these fields are linearly regressed onto simultaneous values of an index of equatorial Pacific SST anomalies. The analysis is performed separately on the data for earlier (1950–78) and later (1979–92) epochs of the record. The analyses are further stratified in terms of the climatological-mean warm and cold seasons in the equatorial Pacific, which correspond to January–May and July–November, respectively. Composite SST, wind, and rainfall fields for the warm and cold seasons that fall within typical warm and cold ENSO episodes are also presented.

Despite the dramatic differences in the sequencing of ENSO warm episodes with respect to the annual march in the two epochs, the anomaly patterns are found to be remarkably similar and generally consistent with the Rasmusson and Carpenter composite. SST and zonal wind anomalies were quite comparable in strength in the warm and cold seasons, although the distributions were somewhat different. Rainfall anomalies and the associated anomalies in surface wind convergence and mean tropospheric temperature were much stronger during the warm season (and particularly during January–February) than during the cold season. Some aspects of the observed rainfall anomaly seasonality can be explained on the basis of simple thermodynamical considerations.

1. Introduction

In a seminal study, Rasmusson and Carpenter (1982, hereafter RC) characterized the spatial distribution and evolution of low-level atmospheric circulation, rainfall, and sea surface temperature anomalies associated with six El Niño–Southern Oscillation warm episodes between 1950 and 1975. Rasmusson and Carpenter composited conditions before, during, and after the occurrence of positive SST anomalies along the Peruvian coast during the calendar months March through May. The key features of the ENSO warm episodes in this epoch, as documented by RC, include a strong seasonality of the anomalous conditions and a rather consistent evolution of the phenomenon from one episode to the next. In particular, the occurrence of anomalous warm SSTs at the Peru coast in March–May is preceded in November–January by westerly surface wind anomalies in the far western equatorial Pacific and followed in August through the following February by positive SST anomalies and anomalous surface wind convergence throughout much of the equatorial Pacific. The ENSO warm episodes typically lasted from 12 to 18 months, with significant cold episodes often following about a year behind warm episodes.

Subsequent studies based on data centered in the epoch studied by RC filled out the description of the ENSO phenomenon in these years. These efforts presented analyses of additional meteorological fields not considered by RC, including sea level pressure (SLP) (Barnett 1983, 1984a, b, 1985; Trenberth and Shea 1987), cloudiness, and outgoing longwave radiation (Heddinghaus and Krueger 1981; Liebmann and Hartmann 1982). They employed a wide variety of analysis techniques, some of which considered data during both warm and cold episodes of the ENSO cycle. The results of these studies largely corroborated RC’s findings of a strong seasonality and a regular evolution of the ENSO phenomenon. Wright et al. (1988), Deser and Wallace (1990, hereafter DW), and Nigam and Shen (1993) emphasized the seasonality of the ocean–atmosphere coupling in the tropical Pacific, with strong interactions between the anomalous basinwide low-level atmospheric circulation and SST during the climatological cold season for the equatorial eastern Pacific (centered on August–October) and much weaker coupling during the warm season (March–May). Barnett (1983, 1984a, b, 1985), Graham et al. (1987), and Meehl (1987) examined the role of interannual vari-

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ability originating over the Indian Ocean in the ENSO phenomenon. These studies identified physically consistent patterns of SLP, wind, and rainfall anomalies that form in some years in June through August over the Indian Ocean and Indian monsoon region, and propagate eastward into the Pacific by the following December–February. Van Loon (1984), van Loon and Shea (1985), Rasmusson et al. (1990), and Rope-
lewski et al. (1992) characterized the distinctive biennial component of the ENSO signal during portions of the historical record.

The ENSO warm episodes since the late 1970s have evolved somewhat differently than those in the epoch documented by RC. The unusually strong 1982–83 warm episode was not preceded by a "buildup phase" with easterly wind anomalies in the equatorial western Pacific, which was believed to be a prerequisite for El Niño events (Wyrtki 1975), and the basinwide warming preceded, rather than followed, the warming along the Peruvian coast (Rasmusson and Wallace 1983). The subsequent ENSO warm episode exhibited positive equatorial SST anomalies from July 1986 through February 1988, significantly longer than the warm episodes in the earlier epoch documented by RC. The most recent episode that began in 1990, by some measures, had a duration that is unprecedented in the his-

Fig. 4. SST and surface wind fields regressed upon normalized values of the CTI for the earlier epoch 1950–78 in the (a) warm (January–May) and (b) cold (July–November) seasons. Contour interval 0.25°C per standard deviation of the CTI; the zero line is thickened and negative contours are dashed.

Fig. 5. As in Fig. 4 but for the later epoch, 1979–92.
historical record (Trenberth and Hoar 1996). As a group, the later episodes did not exhibit the same sequencing of SST anomalies relative to the annual march as the warm episodes of the ENSO cycle during the earlier epoch.

The inability of the ENSO paradigm based on the RC composite to adequately describe the ENSO-related SST variability during the later part of the record suggests the need for a reassessment of what constitutes a typical warm phase of the ENSO cycle and how it relates to the annual march, in light of more recent data. To that end, the present study compares the structure of ENSO-related anomaly fields during “earlier” (1950–78) and “later” (1979–92) epochs, based on linear regression analysis. The earlier epoch includes all the warm episodes in the RC composite, while the later includes the major warm episodes that have occurred since that time and also happens to span the period of continuous satellite observations of ongoing infrared and microwave radiation, which define the rainfall anomalies much more clearly than the surface observations available to RC. Dividing the record in this manner and using separate reference climatologies for the two epochs substantially reduces the impact of the interdecadal-scale variability associated with the 1976–77 “regime shift” (Nitta and Yamada 1989; Trenberth 1990; Graham 1994; Parker et al. 1994) upon the inferred structure of the ENSO cycle. In order to capture the gross features of the seasonality documented by RC, the data are binned separately for the cold, dry (July–November) and warm, rainy (January–May) seasons in the equatorial Pacific. The cold season encompasses the “transition” phase of the RC composite, and the warm season begins during RC’s “mature” phase. In recognition of the diverse evolution of individual warm episodes, no attempt is made to relate the structure of warm episodes during
these seasons to antecedent or subsequent conditions as in the analysis strategy employed by RC.

The paper is organized as follows. Data and analysis techniques are described in the next section. Anomaly fields characteristic of warm episodes of the ENSO cycle are presented in section 3. Reconstructed total fields (climatology plus anomaly) representative of ENSO warm and cold episodes for 1979–92 are presented in section 4. The final, discussion section includes a brief synopsis of corresponding results for the mature phase of the RC composite.

2. Datasets and analysis procedures

a. Datasets

Unless otherwise noted, all the data used in this study were obtained from the NOAA/NCEP Climate Prediction Center. Surface observations comprise monthly mean gridded fields of SST, SLP, and surface wind for the years 1950–92 derived from the Comprehensive Ocean–Atmosphere Data Set (COADS) (Woodruff et al. 1987, 1993). This dataset was provided by the NOAA Climate Diagnostics Center. All SST analyses in map form are based on EOF-filtered SST fields (Smith et al. 1996). Both the COADS fields and the EOF-filtered SST are means for 2° latitude–longitude boxes, centered on odd-numbered latitudes.

The Southern Oscillation index (SOI), based on normalized values of normalized Tahiti SLP anomalies minus normalized Darwin SLP anomalies (Chelliah 1990), is used as an indicator of the strength and polarity of the anomalous ENSO-related zonal wind stress in the equatorial Pacific. Negative (positive) SOI values are associated with eastward (westward) surface wind stress anomalies in the western and central part of the basin.

Several sets of satellite observations are analyzed in this study. Microwave Sounding Unit (MSU) estimates of rainfall (Spencer 1993) and mean tropospheric temperature (Spencer and Christy 1990, 1992, 1993) for...
the period January 1979 to August 1992 were provided by Roy Spencer of the NASA Marshall Space Flight Center. All satellite data are monthly means for 2.5° latitude-longitude boxes, centered at 1.25, 3.75, 6.25, . . . degrees of latitude in both hemispheres.

b. Processing

Climatology and anomaly fields are calculated separately for the years 1950–78 and 1979–92 in the following manner. A climatological mean is calculated for each field and calendar month. The climatologies based on COADS fields are subjected to a three-point running mean along latitude circles.

The climatological mean fields are subtracted from the original data to produce anomalies. The anomaly fields based on the COADS are smoothed with a seven-point running mean along latitude circles, weighted by the number of observations in that month and year. The running mean takes the average of all available data for the seven points and as such acts to fill 2° latitude-longitude boxes for which there are no values. The effective resolution of the smoothed and interpolated data is 28° longitude and 4° latitude. The spatial distribution of tropical Pacific SST observations for the years 1946–76 for a dataset comprising many of the same data as the COADS is given in RC (their Fig. 14).

c. Analysis

Monthly mean gridded anomaly fields of surface wind, rainfall, and SST are linearly regressed onto the time series of cold tongue SST anomalies, defined later in this subsection. For further discussion of this analysis technique, the reader is referred to section 2 of DW.

The regression analysis is performed separately for the epochs 1950–78 and 1979–92. The differences in the annual-mean climatologies of the two epochs are documented in Fig. 1. The later epoch was characterized by higher SST in the tropical southeast Pacific, weaker equatorial tradewinds to the west of the date line, and a lower mean value of the SOI. As noted in the introduction, these difference patterns are largely a reflection of the 1976–77 "regime shift," which is the most outstanding interdecadal-scale feature in the segment of the historical record considered in this study.

The reference variable for the regression analysis is the cold tongue index (hereafter denoted CTI), defined as the SST anomalies (relative to the means of the respective epochs) averaged over the region 6°N–6°S, 180°–90°W, which corresponds to the dotted rectangular box in Fig. 2. The term "cold tongue" refers to the bent-back shape of the isotherms, indicative of a temperature minimum along the equator across most of the Pacific. The variance of monthly mean temperature also exhibits a relative maximum within this region, as indicated by the shading in Fig. 2. The CTI has been used in a number of previous studies, including that of DW. It is a linear combination of the NINO-3 and NINO-4 SST indices used by RC, and it is highly correlated with both of those indices.

The values of the CTI used in the regression calculations are anomalies with respect to the climatology

Fig. 8. Surface wind and MSU rainfall fields regressed upon the normalized CTI for 1979–92 in (a) warm (January–May) and (b) cold (July–November) seasons. The solid contour corresponds to the zero line. Positive anomalies in excess of 6 (12) cm mo⁻¹ are indicated by the lighter (heavier) shading. The dashed contour corresponds to negative anomalies of 6 cm mo⁻¹.
for each epoch. The resulting regression maps are scaled to indicate the typical anomalies in each field for a one standard deviation positive anomaly in the CTI in that season and epoch. The regression maps based on COADS surface vector wind and SLP are subjected to a three-point running mean along latitude circles and an additional “light” two-dimensional smoother is applied to the regression analyses of COADS SLP anomalies.

Composite fields for ENSO warm (cold) episodes are constructed by adding (subtracting) the regression pattern for a given season and epoch to (from) the climatological mean pattern for that season and epoch. The resulting maps represent conditions corresponding to a typical (one standard deviation) perturbation of the CTI. This technique was previously employed by Pazan and Meyers (1982) and DW to document the total fields during warm and cold episodes of the ENSO cycle.

3. Seasonality of ENSO-related anomalies

Figure 3 shows the time series of the CTI for the full period of record 1950–92. Values for the climatological-mean cold season in the equatorial cold tongue region (July–November) are indicated by dark shading and those for the warm season (January–May) by light shading. The major excursions of the ENSO cycle are clearly evident in this figure. During the early epoch 1950–78 most of the peaks in the cycle coincided with the cold season. The compositing procedure used by RC exploits this coincidence to label significant events in the life history of warm episodes of the ENSO cycle in terms of specific calendar months. Such phase locking between the ENSO cycle and the annual march was largely absent during the later epoch.

Regression maps of the fields of SST and surface wind upon the CTI are shown separately for the warm and cold seasons in the equatorial Pacific for the earlier epoch in Fig. 4 and the later epoch in Fig. 5. These maps may be interpreted as documenting typical anomalies during warm episodes of the ENSO cycle. They are comparable in many respects to patterns for different periods of record presented by RC, DW, and Nigam and Shen (1993).

Certain aspects of the seasonality of the ENSO cycle were evident during both epochs. For example, the ENSO-related SST anomalies were more concentrated in the eastern Pacific (apart from the coastal El Niño signature) and the zonal wind anomalies extended farther westward along and to the north of the equator during the cold season, whereas the meridional wind anomalies in the northern Tropics were stronger and more concentrated in the central Pacific during the warm season. These seasonal differences are consistent with the distinctions between the “transition” and “mature” phases of the RC composite and with later results of Harrison (1987). The most notable distinction between the two epochs is the greater strength of the warm season anomalies during the later one.

Figures 6 and 7 show the corresponding regression maps for SLP. An abbreviated version of the surface wind anomalies is included in order to provide a broadscale context for the wind field in the previous figures. In both epochs the SLP anomaly nodal line extended much farther east in the subtropical North Pacific during the warm season, giving rise to a
Fig. 10. January–May warm and cold episode composites for (a,b) MSU rainfall, (c,d) SST, and (e) and (f) surface wind for 1979–92. Thresholds for light and medium shading correspond to 10 and 30 cm mon⁻¹. SST contour interval 1°C, the 27°C isotherm is thickened, and values above 27°C are not plotted.
FIG. 11. As in Fig. 10 but for July–November. Thresholds for light, medium, and heavy shading correspond to 10, 30, and 50 cm mo⁻¹, respectively.
stronger anomalous meridional gradient in the central North Pacific, consistent with the stronger meridional wind anomalies along 6°N. The SLP results are generally consistent with published results of Barnett (1983, 1984a,b, 1985), Graham et al. (1987), and Wright et al. (1988), based on shorter periods of record.

Figure 8 shows the corresponding ENSO-related rainfall anomalies, as estimated from MSU data for the later epoch only. The warm season is characterized by strong rainfall anomalies throughout the equatorial dry zone, whereas the cold season is characterized by a narrow, hook-shaped band of anomalies along the periphery of the dry zone on the equatorward flanks of the ITCZ and South Pacific convergence zone (SPCZ). The rainfall anomalies appear to be consistent with the divergence patterns inferred from the superimposed surface wind anomalies.

Corresponding results for layer-averaged tropospheric temperature, as inferred from channel 2 of the MSU, are shown in Fig. 9, superimposed upon the same rainfall patterns. The characteristic dipole pattern noted previously in upper-tropospheric wind data by Horel and Wallace (1981), Arkin (1982), Rasmussen and Wallace (1983), and Yulaeva and Wallace (1994) is evident in both seasons, but it is much more prominent during the warm season.

4. Seasonal-mean composite fields for ENSO warm and cold episodes

Warm and cold season composites of the total fields (climatology plus anomalies) of SST, rainfall, and surface wind were constructed for warm and cold episodes of the ENSO cycle, using the procedure described in section 2c. Fields for the later epoch are presented in Figs. 10 and 11, and those for the earlier epoch are included for comparison in the appendix. The fields for each variable are presented separately in order to facilitate comparison of any desired pair of fields through the use of transparencies.

The influence of ENSO upon the equatorial Pacific cold tongue/dry zone is most pronounced in the warm season, when the equatorial Pacific cold tongue and dry zone virtually disappear during warm episodes of the ENSO cycle. The corresponding contrasts in the cold season are more subtle: the cold tongue and dry zone are present regardless of the polarity of the ENSO cycle; during warm episodes the ITCZ and SPCZ encroach into the outer periphery of the dry zone, leaving the core intact. Results for this limited period of record give the impression that within the core of the dry zone, significant rain falls only during warm seasons that fall within warm episodes. Meridional profiles of zonal averages of the MSU rainfall total fields for the western (160°E−170°W) and eastern (160°−100°W) Pacific,
shown in Fig. 12, illustrate more clearly the general filling in of the dry zone at these times. The composite wind fields for the earlier epoch, presented in the appendix, suggest that rainfall exhibited a similar behavior with respect to the annual march and the ENSO cycle.

5. Discussion

Seasonal breakdowns other than the climatological-mean warm and cold seasons were considered in this study. Of particular interest are the results for December–February (DJF), which corresponds to the mature phase of the RC composite. They closely resemble the warm season (January–May) results shown above, but the rainfall anomalies and the associated meridional surface wind anomalies inferred from the regression analysis were even stronger (by ~20%) than those for the warm season. The dipole pattern in the MSU-2 tropospheric temperature anomalies was also slightly stronger, but no less equatorially symmetric. It is notable that only the data for the cold season exhibited the hook-shaped rainfall anomalies along the periphery of the equatorial dry zone shown in Fig. 8b. The existence of such a signature suggests that in this season SST in the core of the dry zone is too cold to support deep convection, even during ENSO warm episodes. Why the equatorial rainfall anomalies during warm episodes peak in January–February, a month or two before the peak of the warm season, is not clear.

The results presented above demonstrate that many of the features in the RC composite of ENSO warm
episodes observed during the 1950–78 epoch and Barnett's (1983) complex EOF analysis of the ENSO cycle during the same period of record were also evident in data for the more recent 1979–92 epoch, which lacked the distinctive two-year rhythm that characterized parts of the earlier epoch. Hence, while the two-year rhythm may have contributed to the distinctive sequencing of the warm episodes in the earlier studies relative to the annual march, it evidently cannot be the decisive factor in determining the seasonality of the structure of the anomalies.

Direct evidence concerning the rainfall signature is limited to the later epoch, for which MSU imagery was available, supplemented by GOES Precipitation Index rainfall estimates from 1986 onward (Arkin and Meisner 1987). However, in view of the strong consistency between the surface wind and SLP analyses for the two epochs, it seems quite likely that the rainfall results are representative of the entire 1950–92 period of record.

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APPENDIX

Total Fields for 1950–78

For purposes of comparison with the analyses presented in section 4, total fields of SST and surface vec-
tor wind for warm and cold episodes of the ENSO cycle in January–May and July–November 1950–78 are presented in Figs. A1 and A2, respectively. The differences in SST and surface vector wind between the early epoch analyses (Figs. 10 and 11) and those presented here reflect differences in both the background climate (Fig. 1) and the regression fields (Figs. 4 and 5) between the two epochs.

REFERENCES