The Southern Oscillation in the Early 1990s

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[1] Two aspects of the onset of a warm event in the Southern Oscillation are 1) the subtropical South Pacific High is weakened in early southern winter which weakens subtropical sea level pressure (SLP) gradients, and thus reduces the trade winds and upwelling along the equator, and 2) in the months that follow, and particularly during the mature phase of the warm event during southern summer, the negative SST anomalies in the equatorial cold-water tongue are displaced by positive anomalies and the water along the tropical Peruvian coast usually warms. To highlight the role of Southern Hemisphere subtropical processes in warm events, we focus on the first half of the 1990s. That period has been viewed as a five year warm event, but actually two warm events developed during this period according to the criteria in 1) and 2): in 1991 and 1994. This study emphasizes the importance of interactions between the subtropical and equatorial Pacific in warm events. *INDEX TERMS:* 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 1620 Global Change: Climate dynamics (3309); 3374 Meteorology and Atmospheric Dynamics: Tropical meteorology; 3309 Meteorology and Atmospheric Dynamics: Climatology (1620). **Citation:** van Loon, H., G. A. Meehl, and R. F. Milliff, The Southern Oscillation in the Early 1990s, *Geophys. Res. Lett.*, *30*(9), 1478, doi:10.1029/2002GL016307, 2003.

1. Introduction

[2] Forecasts of the development of warm and cold extremes of the Southern Oscillation are now operational, emphasizing equatorial Pacific SST [*Landsea and Knaff*, 2000; *Wang et al.*, 2002]. The onset of a warm extreme (often referred to as El Niño though the meaning of this term makes its application in this context ambiguous, see *Glantz*, 1996; we use "warm event") is a highly scrutinized part of the forecast [*Kerr*, 2002]. Processes in the subtropics play an important role in development of warm events [e.g. *van Loon and Shea*, 1985], and taking into account these processes should help to quantify the frequency and amplitude of warm events for the evaluation of possible climate change effects, as well as to contribute to predictability of onset and evolution of warm events.

[3] Various definitions have been used to denote its warm extremes (El Niño/Southern Oscillation [ENSO] events, or warm events). Most use some combination of the Southern Oscillation Index, sea level pressure from certain locations, equatorial Pacific SST, or tropical Pacific precipitation [e.g., *Rasmusson and Carpenter*, 1982; *van Loon and Madden*, 1981; *Harrison and Larkin*, 1996; *Trenberth*, 1997; *Wolter and Timlin*, 1998; *Smith and Sardeshmukh*, 2000]. There is general agreement about the onset year (yr_0) of most warm events, though there is disagreement whether some of the lower amplitude warmings qualify as warm events. There can be intervals as long as 11 years and as short as 2 years between warm events.

[4] The longest duration of a warm event has, since the mid-1800s, not been much more than about two years. But near the Date Line, in the Nino4 region (5 N to 5 S, 160 E to 150 W), the SSTs stayed above normal much of the time during the first half of the 1990s [*Trenberth and Hoar*, 1996, 1997]. There is one generally recognized warm event onset in 1991 [e.g., *Harrison and Larkin*, 1996]. The question has been raised if the period 1991–95 should be regarded as one five-year warm event beginning in 1991 [*Trenberth and Hoar*, 1996], a sequence of three warm events [*Goddard and Graham*, 1997], or some manifestation of decadal variability or climate change [*Folland et al.*, 2001; *Stocker et al.*, 2001]. Such a determination is important with regards to assessing the potential impact of global change scenarios on the frequency and amplitude of warm events [*Cubasch et al.*, 2001]. The goal of this paper, then, is to elucidate the role of Southern Hemisphere subtropical processes in warm events insomuch as they involve fundamental interactions between the southern subtropics and equatorial Pacific, and to use the early 1990s as a case study to evaluate these processes in defining warm events.

[5] Here we use the definition of a warm event in the Southern Oscillation given, for example, by *van Loon and Madden* [1981], *Rasmusson and Carpenter* [1982], *van Loon and Shea* [1987], *Kiladis and Diaz* [1989], and *Harrison and Larkin* [1996]. Equatorial Pacific SSTs most often begin rising in southern fall and culminate with maximum positive SST anomalies during the mature phase at the end of year zero and the beginning of $year_{+1}$. By then the equatorial upwelling in the cold-water tongue of the equatorial Pacific, and usually also the cold-water region along the Peruvian coast, are suppressed. A warm event is thus characterized by a positive SST anomaly on the equator and negative anomalies to the south and north.

[6] It has been shown that an important aspect of onset and development of a warm event is the weakening of the South Pacific subtropical high in southern late fall and

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winter [*van Loon*, 1984; *van Loon and Shea*, 1985, 1987; *Harrison and Larkin*, 1996; *Larkin and Harrison*, 2002; *Kidson and Renwick*, 2002]. Negative SLP anomalies spread across the South Pacific over the region 5 S to 45 S from southern fall to southern spring. This results in a slackening of the usual SLP gradient in the tropical south Pacific and a weakening of the trade winds with associated reduced upwelling [*van Loon and Shea*, 1985, 1987].

[7] Since many of the definitions of warm events noted above are somewhat subjective, and because we intend to combine subtropical SLP gradients along with equatorial Pacific SSTs for evaluating the early 1990s period, we define warm events as a combination of two criteria. Namely, Nino3.4 (120 W–170 W, 5 N–5 S) SSTs in the mature phase November–December–January–February (year_0 to year_{+1}) greater than 0.5 standard deviations of the time series from 1957–2001, and anomalous SLP gradients for 35 S minus 5 S, 155 W–105 W, with magnitudes greater than −0.5 standard deviations averaged for the months May–June–July of year_0 over that same time period. We prefer to avoid introducing new indices for warm events since, as we shall see below, indices can sometimes be misleading in a world where no two warm events are alike. However, these criteria for evaluating warm events in the early 1990s quantify the interaction between subtropical and equatorial processes important for warm events. The Southern Oscillation is a large-scale phenomenon and point-to-point indices can sometimes fail to capture the essential variability. An evaluation of warm and cold events for the entire second half of the 20th century is beyond the scope of this paper, but will be the subject of a subsequent study.

[8] Sea level pressure data are from the NCEP/NCAR reanalyses [*Kalnay et al.*, 1996], and SST data are from the Reynolds reconstructed SST dataset [*Reynolds and Smith*, 1994].

2. Sea Level Pressure

[9] Figure 1a shows the May–June–July (MJJ) composite of SLP anomalies for year zero of nine warm events from the mean for 1957–2001. This pattern is the same as that shown by *van Loon and Shea* [1985, 1987], *Kiladis and van Loon* [1988], *Harrison and Larkin* [1996] and others. Note the negative anomalies of over 3 mb near 35–40S stretching across the entire Pacific basin with positive anomalies of over 8 mb centered at 60S.

[10] The statistical significance of these SLP anomalies is difficult to quantify because of an unknown, non-uniformly distributed, spatial dependence in the SLP fields. If the SLP fields did not contain this spatial dependence, then significance would be quantifiable by means of a student t-statistic. Following *Madden et al.* [1999] we make note of this distinction by defining a t^* statistic.

[11] We let

$$t^* = \frac{SLP(x,y) - \overline{SLP}}{\sqrt{\frac{\sigma^2_{SLP}}{N_t}}}$$

where $\overline{SLP}$ is the MJJ average SLP map for non-warm event onset years (37 years) from 1957–2002 (considered a conservative estimate to eliminate contributions from large-amplitude anomalies associated with warm events), σ^2_{SLP} is the corresponding variance map, and N_t is the number of SLP grid points in the analysis. Since the spatial dependence remains unquantifiable, t^* serves as a qualitative measure indicative of large departures in SLP anomaly from typical year-to-year SLP deviations we can expect for the study period.

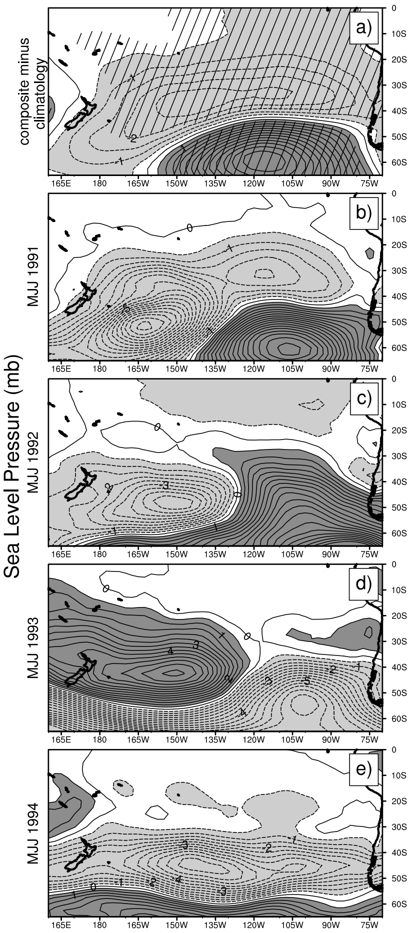

Figure 1. Warm event SLP anomaly differences for MJJ from 1957–2002, a) composite minus climatology, warm event years zero are 1957, 1963, 1965, 1972, 1976, 1982, 1986, 1991, 1997; hatching indicates areas that are probably statistically significant near the 5% significance level; b) MJJ 1991; c) MJJ 1992; d) MJJ 1993; e) MJJ 1994.

[12] The hatching in Figure 1a depicts areas where the quantity $t^*/1.96$ is greater than ±1.0 which would indicate significance of the SLP differences exceeding the 5% significance limit if we could ignore the differences between t^* and a true t statistic for the 9 warm event onset years. Largest amplitudes of $t^*/1.96$ exceed −4.0 near 30 S, and are greater than +3.0 near 60 S (not shown). Even given the inherent spatial dependence, the magnitudes and large regional scales of the hatched areas denoting $t^*/1.96$ exceeding the 5% significance limit in Figure 1a indicate that the SLP anomalies in the warm even onset years are statistically different from the background SLP variance in the region. The analysis was repeated for random collections of 9 years within the study period that are not warm or cold event onset years. In those cases (not shown), the t^*

analysis did not suggest significant departures with respect to the regional standard deviations.

[13] Therefore, the SLP differences in Figure 1a represent a statistically significant weakening of the gradients around the South Pacific subtropical high during the onset of warm events. *van Loon and Shea* [1987] and *Kiladis and van Loon* [1988] showed that positive SST anomalies in the South Pacific Convergence Zone (SPCZ), through cyclonic development, give rise to these negative SLP anomalies, and this was confirmed with a numerical experiment by *von Storch et al.* [1988]. *Kidson and Renwick* [2002] likewise have shown that the negative SLP anomalies are associated with frequent cyclones forming over the warmer water in the South Pacific Convergence Zone (SPCZ). Such processes arise from coupled interactions in previous seasons [*van Loon and Shea*, 1985, 1987; *Meehl*, 1987, 1997; *Kiladis and van Loon*, 1988; *von Storch et al.*, 1988; *Meehl and Arblaster*, 2002]. The pattern is consistent with a reduction of the trade winds with less equatorward transport of cold water and air in the subtropical gyre, and reduced upwelling in the equatorial Pacific and tropical coast of Peru.

[14] The SLP anomalies in Figure 1b for MJJ 1991, a warm event onset, is similar to the composite in Figure 1a, with the anomalous subtropical SLP gradient, defined earlier, greater than minus one standard deviation. In contrast, MJJ 1992 in Figure 1c shows negative SLP anomalies of as much as 5 mb in the southwest Pacific, and positive anomalies in the southeast Pacific reaching to 25 S with an anomalous SLP gradient of opposite sign to that in 1991, and a positive value of 0.75 standard deviations. The pattern is almost opposite in MJJ 1993 (Figure 1d) with positive SLP anomalies in the southwest Pacific and negative anomalies in the southeast Pacific, and an even greater positive anomalous SLP gradient value of almost one standard deviation. In both instances the SLP anomalies and the anomalous positive SLP gradient values are associated with stronger than normal trade winds over parts of the equatorial Pacific (not shown). In MJJ 1994, there were strong negative SLP anomalies stretching right across the South Pacific with negative values reaching north to the equator and maximum values over −6 mb near 40S, with a negative SLP gradient value of 0.5 standard deviations, denoting weakened trade winds. Positive anomalies lie south of about 55S. The MJJ 1994 pattern of SLP anomalies resembles the composite pattern in Figure 1a since both had anomalously negative SLP gradients, while 1992 and 1993 have anomalously positive subtropical SLP gradients.

3. Sea Surface Temperatures

[15] Figure 2a shows composite SST anomalies for the mature phase of warm events (November to February, yr_0 to yr_{+1}) for the same events as the SLP composite in Figure 1a. The familiar pattern of positive equatorial Pacific SST anomalies is evident, with positive anomalies extending from west of the Date Line to the South American coast and maximum values of over 2°C (standard deviations around 1°C, not shown; see also *Kiladis and Mo* [1998]).

[16] In the mature phase of the 1991 event, the SST anomalies in Figure 2b were similar to those in Figure 2a, with Nino3.4 SST anomalies for NDJF of 1.5 standard

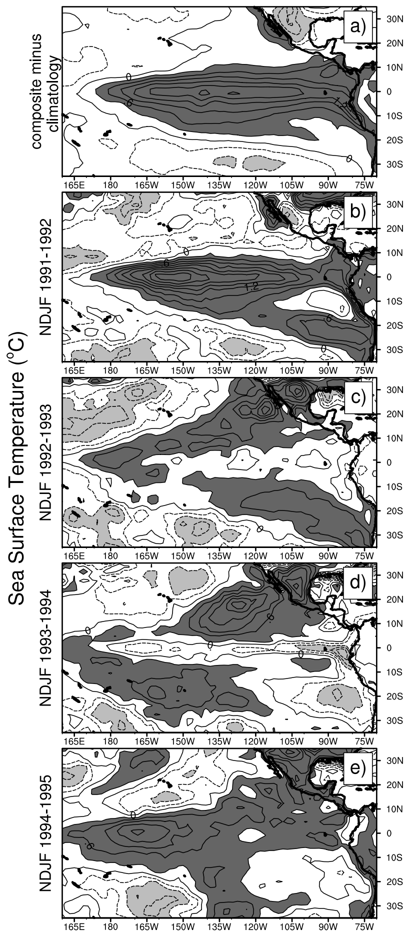

Figure 2. Warm event SST anomalies for the months November (yr_0) to February (yr_{+1}), a) composite minus climatology from 1957–2002, same warm events as in Figure 1; b) Nov.–Feb. 1991–92; c) Nov.–Feb. 1992–93; d) Nov.–Feb. 1993–94; e) Nov.–Feb. 1994–95.

deviations. However, in the next two southern summers (Nov–Feb 1992–93 and 1993–94, Figures 2c and 2d, respectively), there is no resemblance to the composite SST anomaly pattern in Figure 2a, with lower anomalies on the equator than to the north and south, and Nino3.4 SST anomalies of only 0.3 and 0.1 standard deviations, respectively. This pattern resembles that commonly associated with longer timescale variability in the tropical Pacific SSTs in observations and global coupled models [e.g., *Zhang et al.*, 1997; *Meehl et al.*, 1998; *Folland et al.*, 1999]. Such decadal variability (the Pacific Decadal Oscillation or Interdecadal Pacific Oscillation, see *Folland et al.*, 2001), has been shown to modulate ENSO and associated interannual teleconnections to various parts of the Pacific [*Power et al.*, 1999; *Folland et al.*, 2002], and is associated with significant climate impacts [*Mantua et al.*, 1997].

[17] Thus, neither year exhibits warm event characteristics of the mature phase. In contrast, in Nov.–Feb. 1994–95 [Figure 2e] there is an SST anomaly pattern closely resembling the warm event composite in Figure 2a, with positive anomalies in the equatorial Pacific extending from west of the Date Line to the coast of South America surrounded by negative anomalies to the north and south.

There are maximum values greater than +1.5°C (compared to standard deviations there of around 1°C), and a Nino3.4 SST anomaly of 0.9 standard deviations. This pattern of SST anomalies, combined with the MJJ subtropical SLP gradient weakening, qualifies as a warm event as defined above.

[18] It is interesting to note that the SLP anomalies were negative near Tahiti (Figures 1c and 1d) and those over northern Australia were positive (not shown) in MJJ of both 1992 and 1993. The Southern Oscillation Index (SOI, Tahiti minus Darwin) was thus negative and should indicate a warm event. But the index was misleading in this instance, because a warm event, as indicated by the SST anomalies along the equatorial eastern Pacific in the mature phase, and the SLP anomalies in the South Pacific in the onset months, were not evident in Figures 2c and 2d, and Figures 1c and 1d. The SOI, which is a point-to-point difference, was not indicative of a warm event. The Climate Prediction Center "OLR index" (a measure of convective activity between 160 E and 110 W), also does not show continuous warm event conditions in the early 1990s, especially not in 1993.

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