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Case analysis of a role of ENSO in regulating the generation of westerly wind bursts in the Western Equatorial Pacific

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[1] This study speculates that the low-frequency ENSO might have a regulating effect on the activity of episodic westerly wind bursts (WWB) in the western equatorial Pacific (WEP) based upon the analysis of two contrasting El Niño-Southern Oscillation (ENSO) phases during the 1996–2000 period and the onset of the 1982 El Niño. It suggests that ∇P_{WEP}, the equatorial sea level pressure (SLP) gradient between 130°E and 160°E, could be a key parameter in relating ENSO to the WWB generation. ∇P_{WEP} is a preconditioning index for the tropical cyclone formation associated with midlatitude atmospheric transient forcing (cold surges). The cyclonic circulation near the equatorial latitudes contributes to prolonged WWB events. ∇P_{WEP} can be regulated by ENSO because the location of the low SLP center, which determines the strength of the ∇P_{WEP}, correlates with the location of the warm water pool, and has an east-west migration in response to the ENSO phases. Through ∇P_{WEP}, the variability of WWB was linked to the ENSO phases. The study finds that ∇P_{WEP} was promoted prior to the 1997 El Niño when the warm pool was large and extended to near the dateline, whereas it was suppressed during the follow on La Niña in 1999/2000 when the warm pool was small and displaced far westward. However, only westerly winds generated in the east of New Guinea could have potential importance to ENSO as the islands over the Indonesian Seas represent a significant blocking to the propagation of a Kelvin wave. It appears that the active WWB in the winter of 1996 were generated on a favorable background conditioned by pre-El Niño state; they facilitated the development of the 1997 El Niño but may not be a sufficient onset mechanism. Further statistical analysis is required to test the veracity of the hypothesis derived from the case analysis of this study. *INDEX TERMS:* 4522 Oceanography: Physical: El Niño; 4215 Oceanography: General: Climate and interannual variability (3309); 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 3364 Meteorology and Atmospheric Dynamics: Synoptic-scale meteorology; 3374 Meteorology and Atmospheric Dynamics: Tropical meteorology; *KEYWORDS:* westerly wind bursts and El Niño-Southern Oscillation, air-sea coupling over the western Pacific warm pool, multiscale interactions in the ENSO phase transition, regulating effect of ENSO on synoptic-scale atmospheric events

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1. Introduction

[2] Short-period (5 to 20 day duration) westerly winds with amplitude greater than 5m/s and spatial scales of 500–4000 km in longitude and 400–1000 km in latitude are termed as westerly wind bursts (WWB) [*Luther et al.*, 1983; *Keen*, 1988; *Sui and Lau*, 1992; *Delcroix et al.*, 1993; *Verbickas*, 1998]. They are a prominent feature of wind variability in the western equatorial Pacific (WEP) between the months of November–April. On interannual timescales the activity of WWB varies greatly with the El Niño-Southern Oscillation (ENSO) as WWB become more frequent, more energetic, and extend farther eastward along the equator prior to and during El Niño and vice versa during La Niña [*Lukas et al.*, 1984; *Delcroix et al.*, 1993; *Verbickas*, 1998]. The variation of WWB with ENSO phase suggests that low-frequency ENSO might have a role in high-frequency atmospheric events. This study attempted to identify possible ENSO influence on the generation of WWB from observational data.

[3] The ENSO phenomenon arises from unstable interactions between the ocean and the atmosphere in the tropical Pacific [*Wyrtki*, 1975, 1977; *Philander*, 1981; *Rasmusson and Carpenter*, 1982; *Cane and Zebiak*, 1985; *Philander*, 1990]. The system oscillates between warm (El

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Niño) to neutral or cold (La Niña) conditions with a periodicity of roughly 2–7 years. WWB affect the onset of an El Niño [e.g., *McPhaden et al.*, 1998; *Kerr*, 1998] because they generate downwelling Kelvin waves that travel a great distance to suppress the thermocline in the eastern equatorial Pacific Ocean and trigger the El Niño-related warming of the sea surface [*Knox and Halpern*, 1982; *Harrison and Schopf*, 1984; *Lukas et al.*, 1984; *Lukas*, 1988; *Giese and Harrison*, 1991; *McPhaden and Hayes*, 1991; *Delcroix et al.*, 1993; *Kessler and McPhaden*, 1995; *Kindle and Phoebus*, 1995]. WWB also affect the structure of the western warm water pool [*McPhaden et al.*, 1988; *Lukas and Lindstrom*, 1991; *McPhaden and Hayes*, 1991; *Weller and Anderson*, 1996] and generate anomalous zonal currents that advect the warm pool water eastward, enhance the eastward movement of atmospheric convection, precipitation, and zonal wind convergence [*Deser and Wallace*, 1990; *Webster and Lukas*, 1992; *Liu et al.*, 1995; *Picaut et al.*, 1996; *Yan et al.*, 1997; *Delcroix*, 1998]. Easterly trade winds are relaxed in response, and in turn help to prolong westerly wind events (WWE, 30 to 90 day duration) and reinforce the El Niño warming [e.g., *McPhaden*, 1999]. We use WWB and WWE in this study to denote that the two types of westerly wind events not only have different durations but also are associated with different large-scale circulations. WWE result from the basin-scale adjustment of the wind field to the changes of SST during El Niño, while WWB, which is the subject of this study, originate from synoptic-to-intraseasonal atmospheric events and are observed mostly prior to an El Niño.

[4] Existing studies [e.g., *Luther et al.*, 1983; *Keen*, 1988; *Giese and Harrison*, 1991; *Sui and Lau*, 1992; *Kiladis et al.*, 1994; *Phoebus and Kindle*, 1994; *Hartten*, 1996; *Meehl et al.*, 1996; *Fasullo and Webster*, 2000] have provided detailed descriptions of spatial variation and temporal distribution of WWB. Major atmospheric conditions in leading to WWB are identified: twin or individual tropical cyclones [*Keen*, 1982; 1988], the passage of the convective phase of the intraseasonal Madden-Julian Oscillation (MJO) [*Lau et al.*, 1989; *Sui and Lau*, 1992; *Hendon et al.*, 1998], cold surges from extratropics [*Lim and Chang*, 1981; *Arkin and Webster*, 1985; *Love*, 1985a, 1985b; *Chu*, 1988], or the combination of all the three processes [*Yu and Rienecker*, 1998]. It has been observed that WWB were associated with the onset of every El Niño in the past 50 years [e.g., *McPhaden et al.*, 1998; *Kerr*, 1999], an indication that the connection between WWB and El Niño could be more than just coincidental. This connection, if physically based, could challenge leading ENSO theories that view the coupled ENSO system as a deterministic process [e.g., *Zebiak and Cane*, 1987; *Schopf and Suarez*, 1988; *Battisti and Hirst*, 1989; *Neelin et al.*, 1998; *Jin*, 1997] and could also challenge ENSO prediction skills that currently do not include the simulation of high-frequency atmospheric transients. Hence the answer to whether the generation of WWB could be influenced by low-frequency ENSO would be useful to both theoretical and modeling studies.

[5] This study sought to identify leading physical processes that linked the WWB activity with ENSO from observations. We chose to use an intercomparison approach in the analysis, as we believed that certain atmospheric structures associated with the generation of WWB would be different at different ENSO phases if the observed WWB–ENSO connection holds. Hence the differences that arise from contrasting two different phases could lead to the identification of likely ENSO effects. The two ENSO periods chosen in this study were the El Niño onset phase in the boreal winter/spring of 1996/97 and the follow-on La Niña mature phase in the winter/spring of 1999/2000. The availability of high-accuracy, high-resolution vector wind fields from scatterometers NSCAT and QuikScat during these two particular periods aided the analysis of synoptic-scale variability. However, the two chosen ENSO phases were not exactly opposite to each other. We have considered whether the onset phase of the 1998–2000 La Niña should be used, but this does not appear to be a better choice. First, the onset of La Niña conditions started in late spring of 1998 [e.g., *McPhaden*, 1999], which was in a season that WWB are climatologically not active. Second, onset mechanisms for El Niño and La Niña are not symmetric and so the comparison may not have to be performed on symmetric phases. Episodic wind events are regarded as a trigger in the former, while in the latter they are not a predominant factor as reflected equatorial waves play a more significant role [e.g., *Delcroix et al.*, 2000]. We have also considered whether two mature phases should be chosen. This choice again is not appropriate. Our focus is on the WWB activity during the winter/spring period with an emphasis on the onset phase of El Niño. The El Niño in the mature phase is associated primarily with prolonged WWE, whose genesis resides in the basin-scale atmosphere-ocean coupling [*Philander*, 1991] rather than those synoptic-to-intraseasonal atmospheric events that WWB are associated with [*Fasullo and Webster*, 2000]. Therefore it is more pertinent to have two contrasting ENSO phases with the same winter/spring period than to have two exact opposite ENSO phases.

[6] This study focused on the analysis of the 1996–2000 ENSO cycle with an extension to the onset of the 1982 El Niño. The 1997 and 1982 El Niños were the two strongest events of the century. Large anomalous atmospheric and oceanic conditions stood well above the background so that features essential to the event development can be easily identified. We present merely what we have found and could draw from the two cases. Generalizing features among past onsets of El Niño events demands the use of better quality and consistent data sets and is a subject being actively pursued in an ongoing study. Finally, we would like to point out that the case analysis approach only enables this study to speculate a mechanism indicated by data. Proving and establishing the thus-derived mechanism will require further statistical and dynamical analysis.

[7] The presentation is organized in the following way. Section 2 provides a brief description of the data. Section 3 presents the intercomparison analysis of the generation of WWB at the two chosen periods during the 1996–2000 ENSO cycle. Section 4 examines the onset of the 1982 El Niño with available NCEP data sets. Summary and discussions are included in section 5.

2. Data

[8] Fields of surface vector wind, sea level pressure (SLP), and sea surface temperature (SST) were analyzed.

A brief description of each data source is given in the following.

2.1. Scatterometer Winds

[9] High-quality, high-resolution surface wind fields are needed to map the evolution of synoptic-scale WWB events. For this purpose, vector wind observations from satellites were the focus of the analysis. Satellite wind fields were derived from two scatterometers, the NSCAT aboard the Advanced Earth Observing Satellite (ADEOS) by a mission of the National Space Development Agency of Japan and the SeaWinds on space mission QuikSCAT of the National Aeronautics and Space Administration (NASA). NSCAT winds were available from September 1996 to June 1997, while QuikScat winds were from September 1999 onwards. Data used in this study were processed by the Jet Propulsion Laboratory [*Tang and Liu*, 1996] and available twice daily with a resolution of 0.5° in both longitude and latitude.

[10] However, the record of satellite winds was too short to derive wind anomalies or to cover any other major El Niño events. To overcome these problems, we processed two other wind products, i.e., surface wind observations from TOGA Tropical Atmosphere Ocean (TAO) buoys in the equatorial Pacific and wind reanalysis from the National Centers for Environmental Prediction and National Center for Atmospheric Research (NCEP/NCAR). These two data-sets were used to aid and/or extend the analysis of satellite wind features.

2.2. TAO Measurements

[11] The TAO array, covering the equatorial belt (10°S–10°N) of the Pacific Ocean with surface and subsurface instrumentation fixed to approximately 70 moored buoys [*McPhaden et al.*, 1998], provides realtime measurements of oceanographic and surface meteorological variables. This study used daily-averaged TAO surface wind and SST observations to depict the anomalous conditions associated with the El Niño development and to supplement information for the scatterometer total vector field.

2.3. NCEP/NCAR Reanalysis Product

[12] The NCEP/NCAR reanalysis, produced from a state-of-the-art analysis-forecast system that assimilates data from 1948 to the present [*Kalnay et al.*, 1996], is available with a resolution of 2.5° in both longitude and latitude. The SLP data were used to help analyzing the changes of the pressure field associated with midlatitude atmospheric events. The winds were used to extend the analysis of the El Niño onset phase in the winter/spring of 1981/82. We were aware that although the NCEP fields have an overall realistic simulation of wind and SLP fields, they have biases in the tropical oceans [e.g., *Meisser et al.*, 2001]. However, pattern identification is the main target of the study and the NCEP product is used as a supplementary data set, we anticipate that the end results of the analysis would not be influenced by the inclusion of the NCEP data.

2.4. SST

[13] SST data from 1982 onward in weekly and monthly means and on 1° grid were analyzed. These data were produced from an optimal blending of SSTs from the Advanced Very High Resolution Radiometer (AVHRR) with in situ SST measurements [*Reynolds and Smith*, 1994].

3. What Caused the Variation of WWB During the 1996–2000 ENSO Cycle?

3.1. WWB in the Winter/Spring of 1996/97 and 1999/2000

[14] The state of the tropical Pacific in the boreal winter/spring of 1996/97 was sharply different from that of 1999/2000 due to their association with different ENSO phases. This is shown in Figures 1a–1b, the time-longitude plots of zonal wind component (colored) with the superposition of the 28–29°C isotherms (solid lines) for the months between October and May for the two respective ENSO periods. Both zonal winds and SSTs were the averages in the 5°S–5°N equatorial band. In the winter/spring of 1996/97 prior to the El Niño event, WWB were frequent (occurring every 30–50 days), intense (amplitude greater than 8 m/s), and long lasting (duration longer than 7 days). The two episodes, one in December 1996 and the other in February–March 1997, played a significant role in the onset of the El Niño warming in the eastern Pacific. For instance, the appearance of an anomalous warming in the eastern basin coincided with the arrival of the downwelling Kelvin waves generated by the WWB of December 1996 [e.g., *Yu and Rienecker*, 1998; *McPhaden*, 1999; *Delcroix et al.*, 2000; *Wang and Weisberg*, 2000]. The surface warming in the eastern basin reduced zonal SST gradient, weakened easterly trade winds, promoted and prolonged westerly wind anomalies that further enhanced the eastern warming, and fueled the El Niño development. Figures 2a–2b plot the evolution of the equatorial SST and zonal wind anomalies for an extended period that covers from September 1996 to December 1997. The positive feedback between westerly winds and SST in making a warm event [*Philander*, 1990] is clearly indicated by data.

[15] The genesis of the WWB occurring before February 1997 appears to be different from that of the WWB occurring after February 1997 in association with the switching of the east-west SST anomaly gradient around that time. The WWB occurred before the month presided over positive SST anomalies in the WEP and were a localized episodic event. Those WWB were often referred as a trigger of the event [*McPhaden*, 1999] because of their important role in triggering the eastern basin warming and setting off the El Niño development. WWB events after February 1997 were a basin-scale phenomenon with a center located to the west of large positive SST anomalies of the central and eastern basin, a manifestation that they were coupled with the basin-scale adjustment of the wind field to the changes of SST in the east [*Philander*, 1990].

[16] The basic evolution of the 1997 El Niño [e.g., *Wang and Weisberg*, 2000] largely followed the five canonical phase patterns constructed by *Rasmusson and Carpenter* [1982], among which November–January are referred as the onset phase and March–May as the peak phase. By such definition, the WWB occurring before February 1997 were associated with the onset phase of the event. Although it has been known that unstable interactions between the atmosphere and ocean are the cause of the amplification of both SST and westerly wind anomalies during an warm event

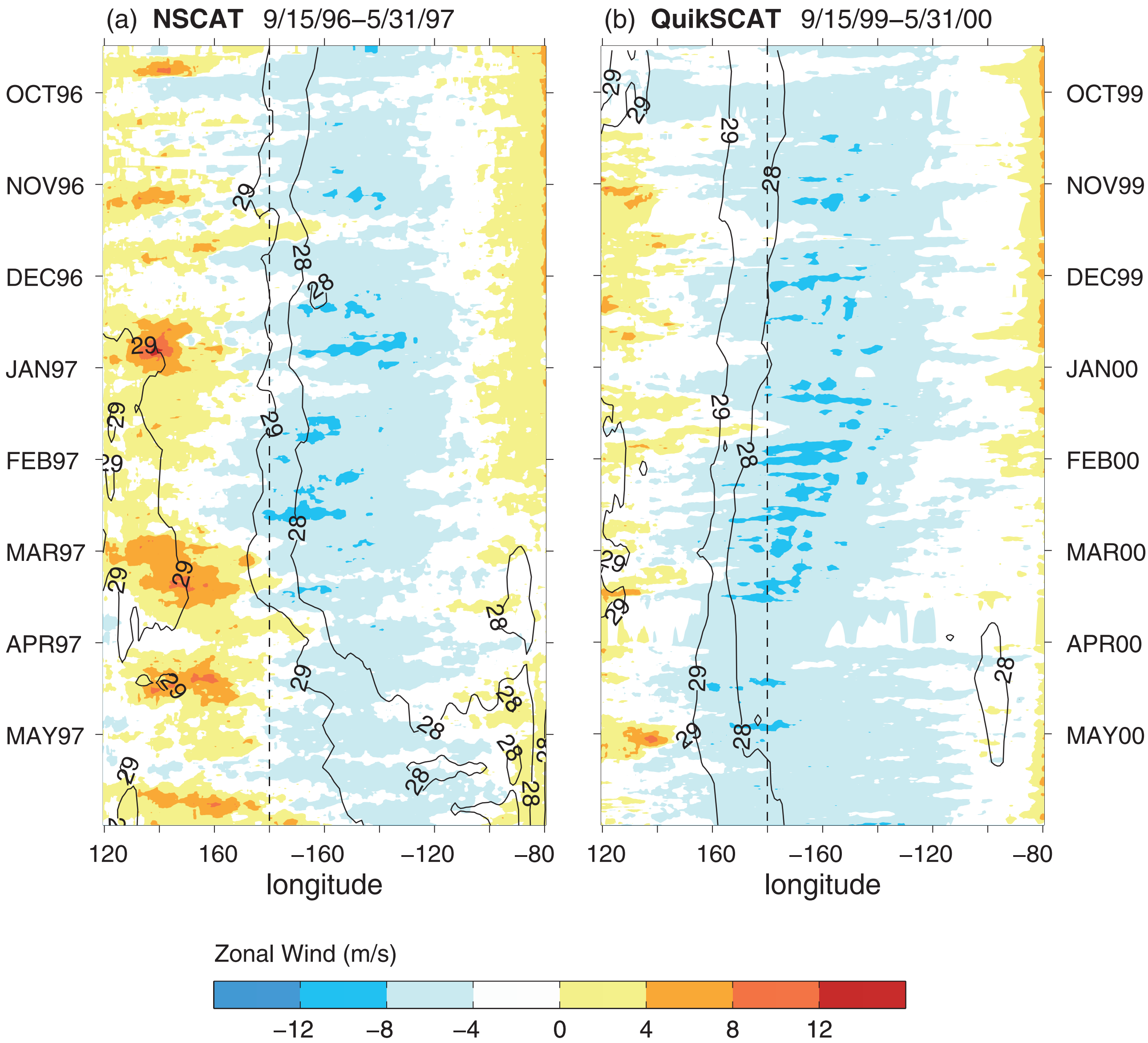


Figure 1. Time-longitude plots of the equatorial zonal winds averaged between 5°S–5°N (colored) for the period from October to May in (a) 1996/1997 from NSCAT and (b) 1999/2000 from QuikSCAT. Superimposed solid contours are the 28–29°C isotherms averaged between 5°S–5°N.

[*Hirst*, 1986], the cause of the active WWB during the onset phase, particularly whether those WWB are associated with evolving pre-El Niño state or merely a stochastic atmospheric forcing, is not yet well known. This study aimed at exploring these issues from observations and hence focused mainly on the atmosphere-ocean conditions with which the WWB in the onset phase were associated.

[17] The shift of the WWB activity with ENSO phase was dramatic. In the winter/spring of 1999/2000 when a La Niña prevailed (Figure 1b), there were only short-lived, weak pulses of westerly winds but no significant WWB over the entire period. Compared with Figure 1a, the 28°C isotherm, which is a commonly used index of the warm water pool, was persistently westward displaced. It can be seen that the enhanced WWB activity in the winter (defined as November–January in this study) of 1996/97 was associated with a warm pool extending to the east of the dateline, while the weak activity in the winter of 1999/2000 was related to a warm pool that was displaced about 20° westward to the west of the dateline. Hence there appears to exist a relationship between the location of the warm pool and the activity of WWB. Such a relationship was possible because the WWB in the onset phase occurred over positive SST anomalies in the WEP (Figures 2a–2b). In the following, physical processes that could support this relationship are analyzed using observational data.

3.2. Cause of WWB Variability With ENSO Phases

3.2.1. Generation Mechanisms for WWB

[18] To examine the WWB variability and its possible connection to the warm water pool, we first need to know how WWB were generated during an onset phase. The analysis by *Yu and Rienecker* [1998, hereafter referred to as YR98] indicated that the generation of the strong WWB

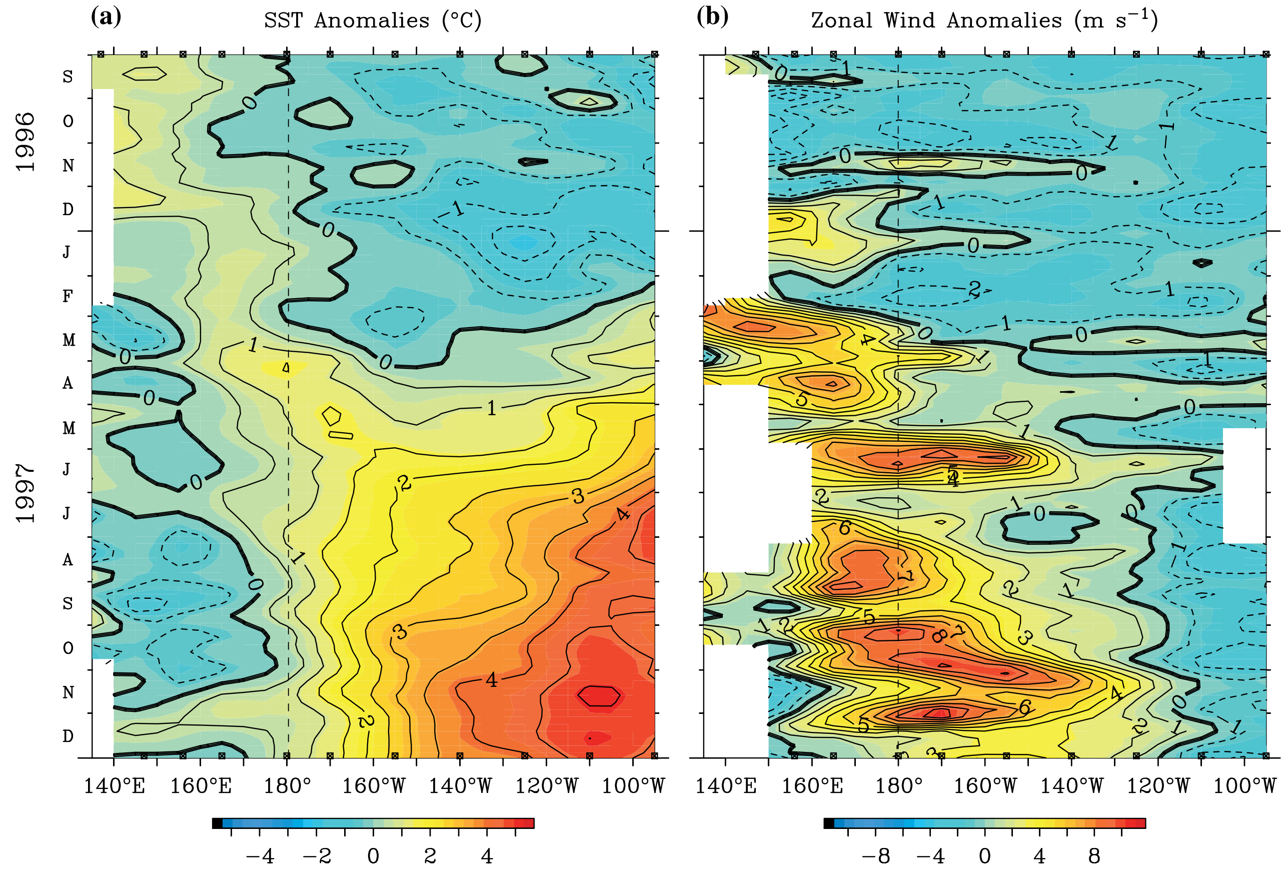


Figure 2. Time-longitude plots of the equatorial (a) SST anomalies and (b) zonal winds anomalies observed from TOGA TAO for the period between October and May in 1996/1997.

episode in December 1996 involved a series of processes: the penetration of northerly surges from East Asia to near equatorial latitudes, a formation of individual/paired tropical cyclones after the surge forcing, and the enhancement of westerly winds on the equator by the cyclonic circulation. It also indicated that though the development was in phase with the convective passage of the intraseasonal Madden-Julian Oscillation (MJO) that propagated from the Indian Ocean (see Figure 2 in YR98), the large amplitude of the westerly winds was produced by the aforementioned processes in the western Pacific sector not by the MJO alone.

[19] Northerly surges are a notable feature of the East Asian winter monsoon circulation [e.g., *Ramage*, 1971; *Murakami*, 1979; *Chang et al.*, 1979; *Chang and Lau*, 1980; *Lim and Chang*, 1981]. The outbreak of cold air from the East Asian continent results from the build-up of the Siberian high by excessive radiative cooling and persistent cold air advection along with the blocking effect of the Tibetan plateau to the southwest [*Lau and Li*, 1984]. The surges are most active during the peak monsoon months of November to February, occurring at an interval of several days to weeks [*Zhang et al.*, 1997]. To see how the surges were involved in making a tropical cyclone and resulting in a WWB event, the case in December 1996 is described here.

[20] Figures 3a–3g show a sequence of vector wind fields observed from the scatterometer NSCAT. The event began on 18 December, featuring a northerly surge from the East Asian continent over north of 30°N. The northerly winds reached the equatorial region between 120 and 140°E over the open seas northeast of New Guinea (an island stretches southeast to northwest with the northern end crosses the equator at about 130°E). East of the area, a weak westerly flow appeared on the equator (Figure 3a). As the surge continued for the next few days, the westerly flow fed into a cyclonic circulation centered north of the flow and the westerly flow itself was in turn amplified greatly by the subsequent development of the cyclone (Figures 3b–3d). Meanwhile, a cross equatorial flow was induced and this led to the development of another cyclone in the Southern Hemisphere (Figures 3e–3f). The two cyclones, straddling the equator, were combined to reinforce and extend eastward the westerly winds on the equator. As YR98 showed, this sequence of development was also observed during the second WWB episode occurring in February–March 1997. All the evidences seem to suggest that northerly surges were the forcing (or trigger) for the formation of individual/pared cyclones in the near equatorial region and the resulting cyclonic circulations were the source for a sustained and large-amplitude WWB event. This inference is consistent with previous analyses [*Love*, 1985a; *Chu*, 1988; *Chu and Frederick*, 1990]. It is worthy noting that the surges from the Southern Hemisphere could exert the same influence on the formation of tropical cyclones and associated near-equatorial WWB in the WEP [e.g., *Love*, 1985b]. Those southerly surges, however, occur mostly during the austral

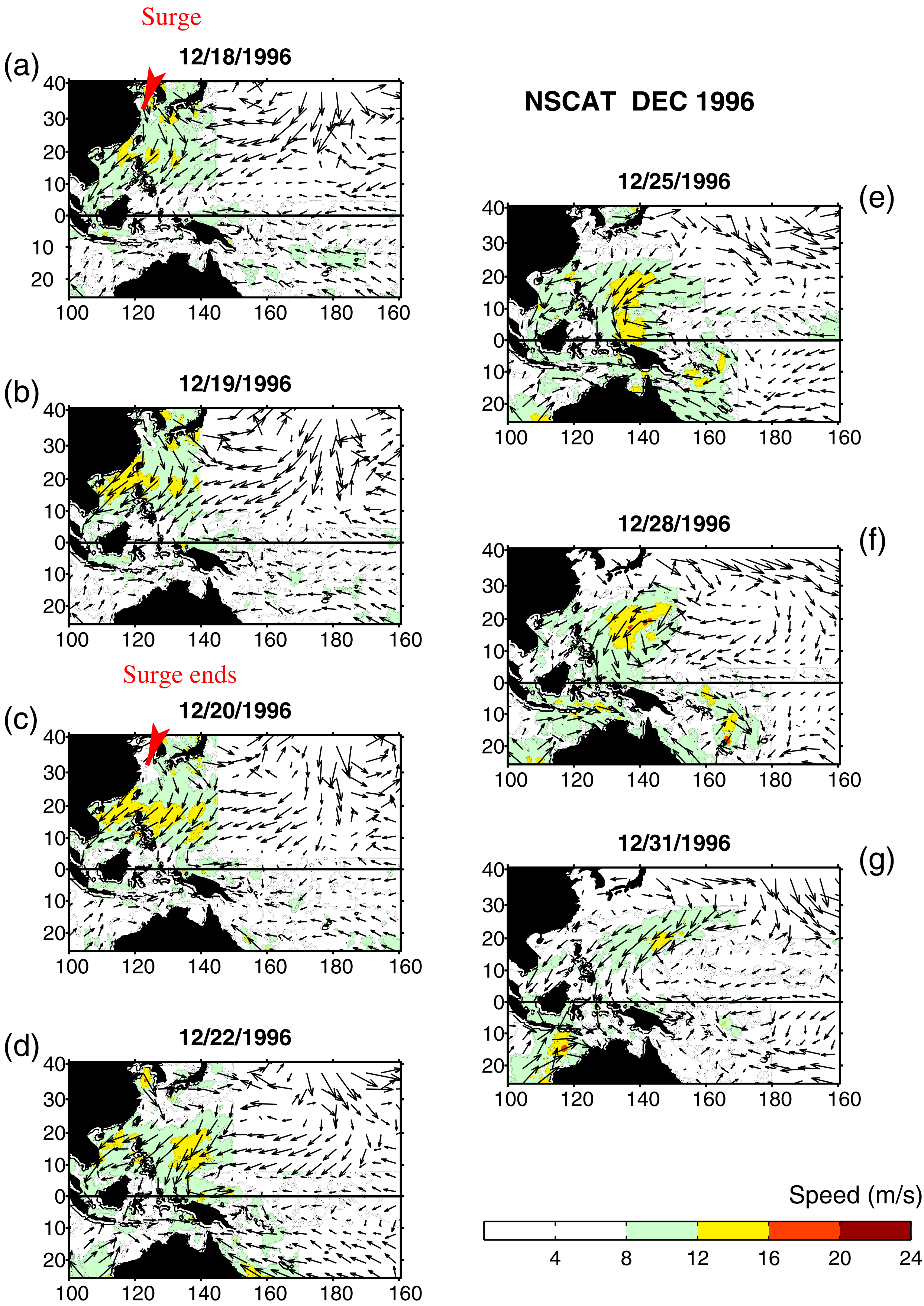


Figure 3. Evolution of daily surface wind fields derived from NSCAT in December 1996. Wind speed greater than 8m/s in the west of 160°E is colored.

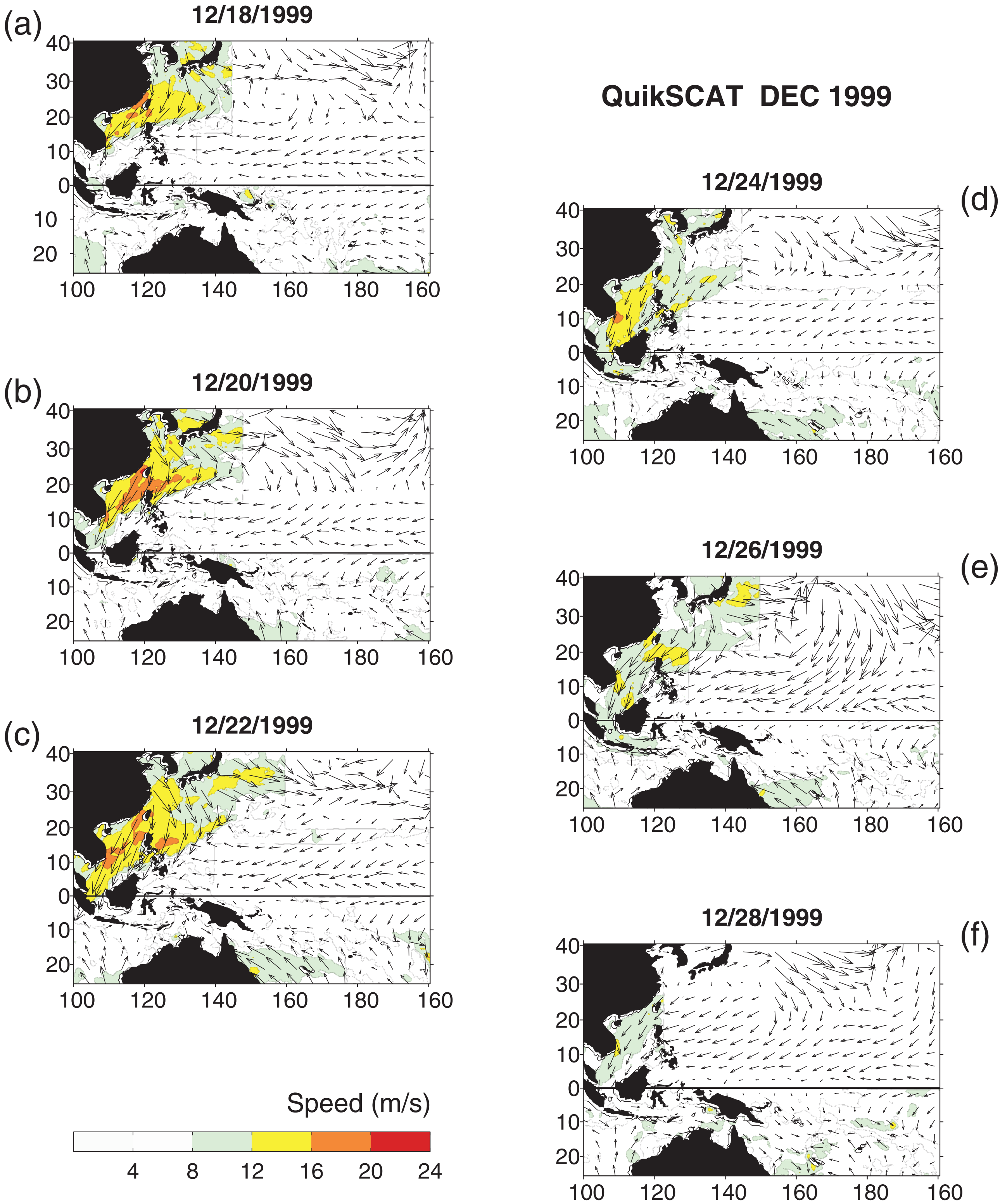


Figure 4. Same as Figure 3 but derived from QuikSCAT in December 1999.

winter/spring season (May–November) and do not contribute to the onset phase of El Niño.

[21] Now that the role of northerly surges has been indicated by data in the winter/spring of 1996/97, was the lack of a WWB event in the winter/spring of 1999/2000 due to the lack of northerly surges? This appears not to be the case. We found that northerly surges still frequented the equatorial area but there was no subsequent development of tropical cyclones. For instance, a strong northerly surge occurred during 18–24 December 1999, as is evident from the QuikSCAT vector wind observations in Figures 4a–4f. However, the surge passed through the South China Sea and impacted the equatorial region between 100–120°E, about 20° westward compared with the surge location in Decem-

ber 1996. This westward shift of surge pathway apparently shifted also the equatorial region influenced by surge forcing. To the east of New Guinea, the surge influence was limited: only weak westerly winds presented briefly on 22 December (Figure 4c) and there were no cyclone development and no large-amplitude equatorial westerly winds.

3.2.2. Role of Northerly Surges

[22] The impact of northerly surges on SLP, air temperature, and northerly wind speed can be felt all the way to near-equatorial latitudes and leads to tropical convective activity and cyclone formation [e.g., *Krishnamurti et al.*, 1973; *Murakami*, 1979; *Chang and Lau*, 1980; *Chu*, 1988; *Kiladis et al.*, 1994; *Compo et al.*, 1999]. Two processes are responsible for the rapid widespread influence of northerly surges [*Chang et al.*, 1983]: fast gravity waves with typical speed of 40 m s^{-1} and the passage of the front at 10 m s^{-1}.

[23] The role of northerly surges in initiating tropical cyclone formation has been well documented [*Chang et al.*, 1979; *Chang and Lau*, 1980; *Love*, 1985a; *Chu*, 1988; *Chu and Frederick*, 1990]. At near-equatorial latitudes the Coriolis term is negligible and the advection term is about one order smaller than the pressure gradient and so there holds the relationship:

$$\frac{\partial U}{\partial t} \approx -\frac{1}{\rho}\frac{\partial P}{\partial x}$$

(where U denotes the zonal flow, P denotes the surface pressure, and ρ denotes the air density). This indicates that a change in the equatorial zonal pressure gradient can lead to the acceleration of an equatorial jet. As northerly surges are pressure surges [e.g., *Compo et al.*, 1999], they can raise the SLPs at near-equatorial latitudes, generate a west-to-east pressure gradient in the region, and force a down-pressure-gradient westerly equatorial flow. The flow then leads to a cyclonic circulation north of the equator governed by the conservation of potential vorticity [*McBride and Zehr*, 1981]. Once the cyclone is formed, the cyclonic circulation contributes to the westerly flow on the equator and leads to the amplification of WWB.

[24] The different responses of equatorial pressures to northerly surges in December 1996 and 1999 are shown in Figures 5a–5b, respectively. The top seven panels in each figure are the SLP deviations at seven locations along the 130°E meridian as a function of time and the eighth panel is the SLP deviation on the equator at 160°E. The ninth (bottom) panel is the change of ($P_{130}-P_{160}$), the SLP differences between 130°E and 160°E on the equator. ($P_{130}-P_{160}$) is used to characterize the west-to-east SLP gradient in the WEP and is denoted by ∇P_{WEP} hereafter. ∇P_{WEP} is eastward (westward) if ($P_{130}-P_{160}$) is positive (negative). In December 1996 (Figure 5a), an increase of SLP was observed at all latitudes along 130°E after the surge occurred. However, P_{160}, the equatorial SLP located to the east of the surge influence region (Figure 3), experienced a drop one day after. The drop at this location appears to coincide with the initiation of a cyclone formation north of the equator. The combination of the surge-induced increase in P_{130} and a cyclone-induced drop in P_{160} led to a rapid rising of ($P_{130}-P_{160}$) till its peak at 23 December, when the surge ceased and a tropical cyclone had been well established (Figure 3). The positive response of ($P_{130}-P_{160}$) to the surge forcing did not occur in the winter/spring of 1999/2000 (Figure 5b). At that time after a northerly surge on 17 December, both P_{130} and P_{160} increased. The increment in P_{160} was larger than that in P_{130} and so a positive ($P_{130} - P_{160}$) was not established. Consequently, there were no ∇P_{WEP}, no down-pressure-gradient zonal jet to precondition the cyclone development, and no WWB (Figure 4).

[25] On the equator, the Coriolis force vanishes. If ($P_{130}-P_{160}$) is increased by 1 mb, the near-equatorial zonal wind would be accelerated by 2 m s^{-1} within a day, given that Δx is the distance between the two locations and the air density ρ is 1.225 kg/m^3. Thus the 2 mb increase in ($P_{130}-P_{160}$) by the northerly surge (Figure 5a) had significant impact on the near-equatorial wind. It can be seen that, within the three days between 18 December and 20 December 1996 (Figure 1a), the amplitude of the zonal wind was boosted by more than 4 m/s by the surge forcing.

3.2.3. Cause of Different ∇P_{WEP} Behaviors

[26] Here we attempted to find indications from existing data about why ∇P_{WEP} behaved differently during the two different winter periods. For this purpose, the daily evolution of the SLP fields in Decembers 1996 and 1999 is shown in Figures 6 and 7, respectively.

[27] The positions of 130°E and 160°E on the equator are marked by dark grey squares for an easy comparison with Figure 5. In December 1996 at the beginning of the northerly surge, there was a broad near equatorial low (enclosed by the 1008 mb isobar) to the east of New Guinea and ($P_{130}-P_{160}$) was positive (Figure 6a). During the surge, the equatorial SLPs between 110–130°E were perturbed by the surge forcing. This can be seen by following the movement of the 1010 mb isobar (the dashed line): the isobar was displaced from near 10°N on 18 December (Figure 6a) to near the equator on 20 December (Figure 6c). The equatorward movement appears to have increased SLP on the equator and enhanced ($P_{130}-P_{160}$) (Figure 5). Subsequently, there were a cyclone development to the northeast of New Guinea and a formation of the cross-equatorial cyclonic pair (Figures 6d–6g). The development was similar to the one indicated by the surface wind fields (Figures 3d–3g).

[28] By comparison, the low SLP center (SLP < 1008 mb) right before the surge in December 1999 (Figure 7a) was small and westward. It lay over the region between 120° and 170°E at near-equatorial latitudes so that both P_{130} and P_{160} were covered by the low SLP center and no ∇P_{WEP} existed (see the bottom panel in Figure 5b). When the surge took place, the surge affected the equatorial region between 90–110°E, which was about 20° westward compared with the region in December 1996 (Figure 6a). The impact of the surge seems to compress the low SLP center to the southwest of New Guinea and further keep it to the south of the equator. Consequently, both P_{130} and P_{160} were increased and the increment on P_{160} was even larger as the low SLP center being moved away from the location (see also Figure 5b). This may be a reason that positive ∇P_{WEP} was not established during the surge event.

[29] The two cases seem to suggest that the response of ∇P_{WEP} was related to the presurge condition of ∇P_{WEP}. If ∇P_{WEP} was preconditioned in the background prior to the surge event, the surge forcing promoted its magnitude and

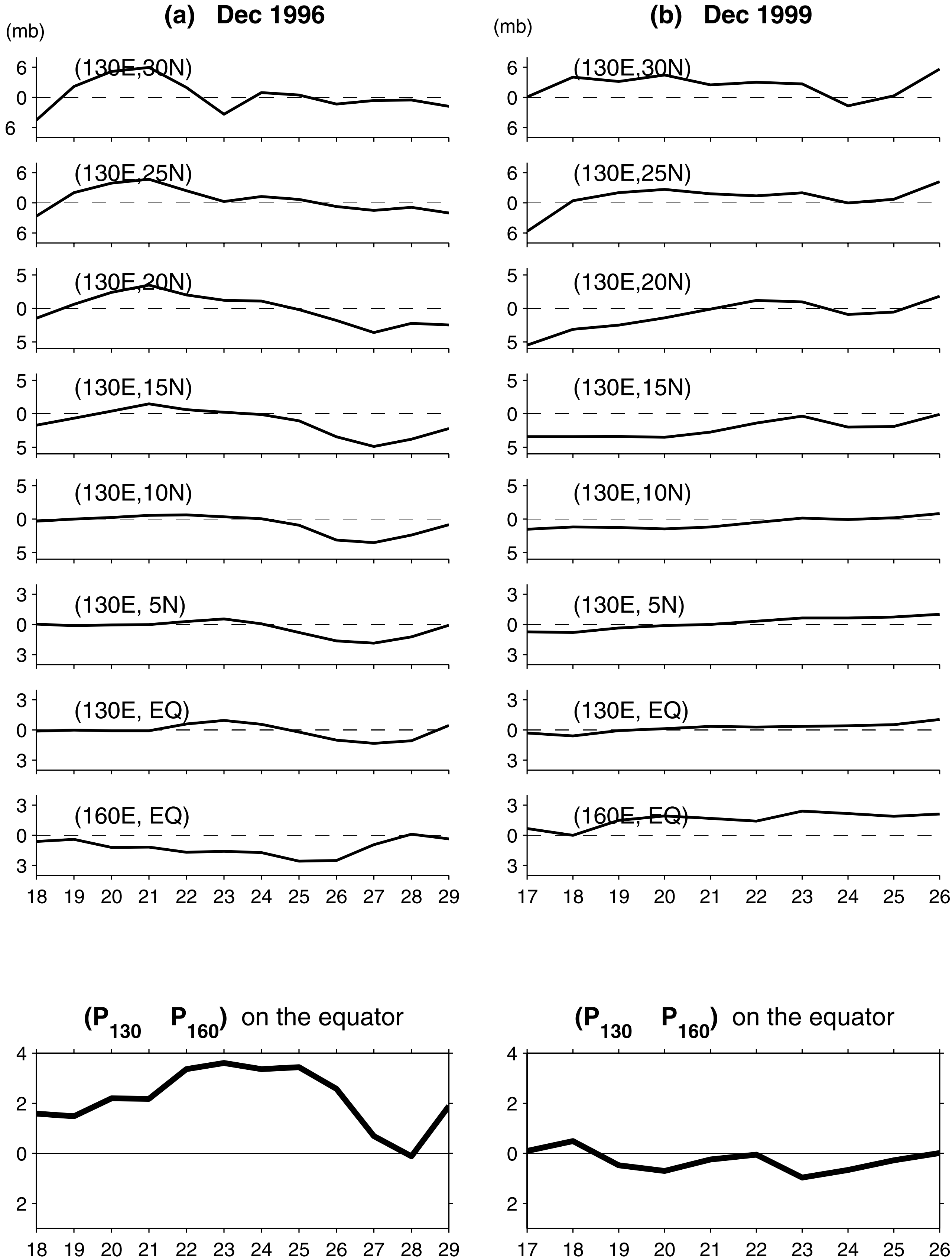


Figure 5. Deviations of the SLP at seven locations along the 130°E meridian (top seven panels) and at the equatorial position [EQ, 160°E] (the eighth panel), and the changes of ($P_{130}-P_{160}$) on the equator (the bottom panel) as a function of time for the period of (a) 18–29 December 1996 and (b) 17–28 December 1999.

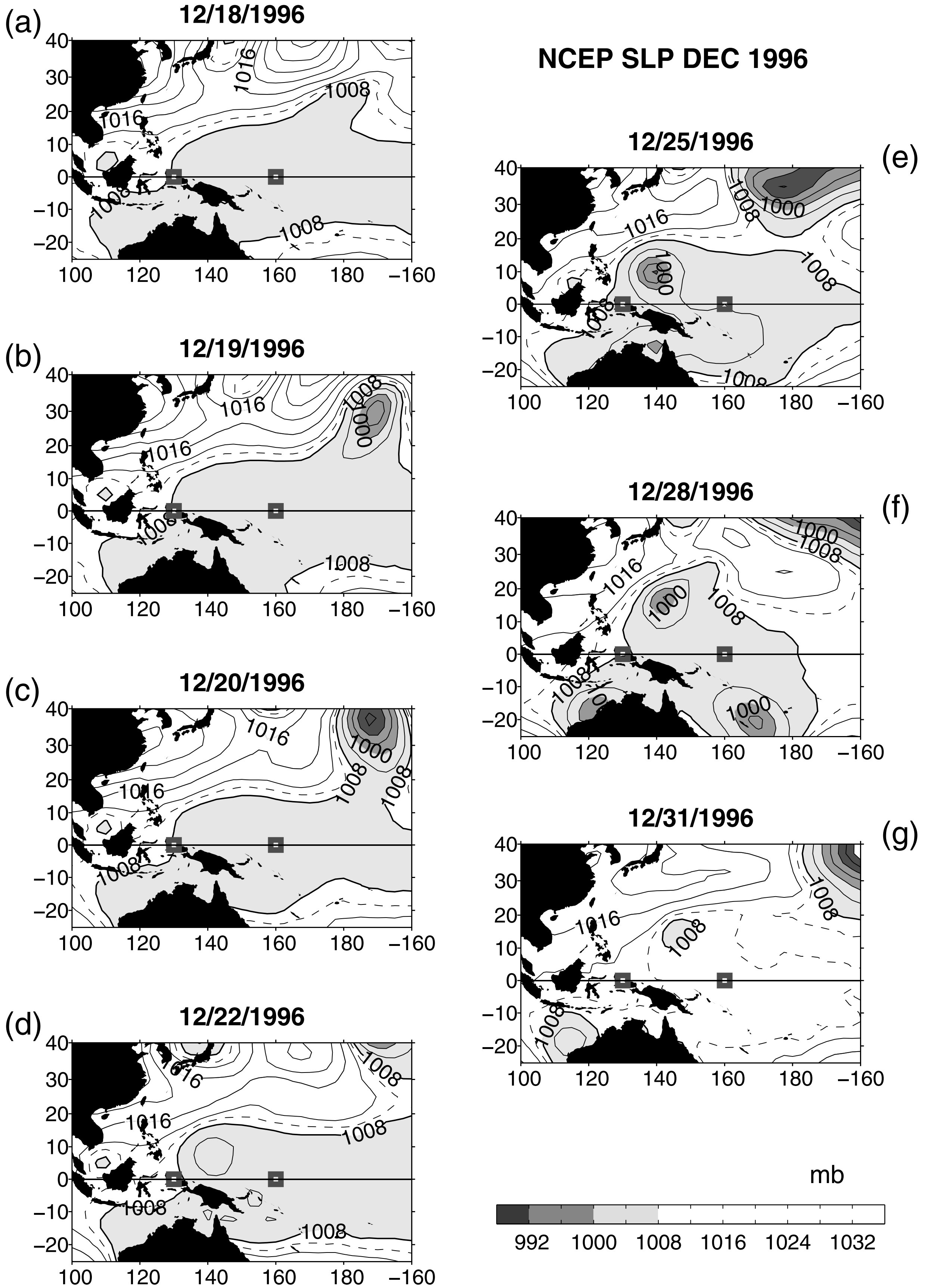


Figure 6. Evolution of daily SLP fields associated with a surge event in December 1996. The contour of the 1010 mb isobar is dashed and the 1008 mb isobar is highlighted. The two gray squares on the equator indicate the two positions used to calculate ∇P_{WEP}.

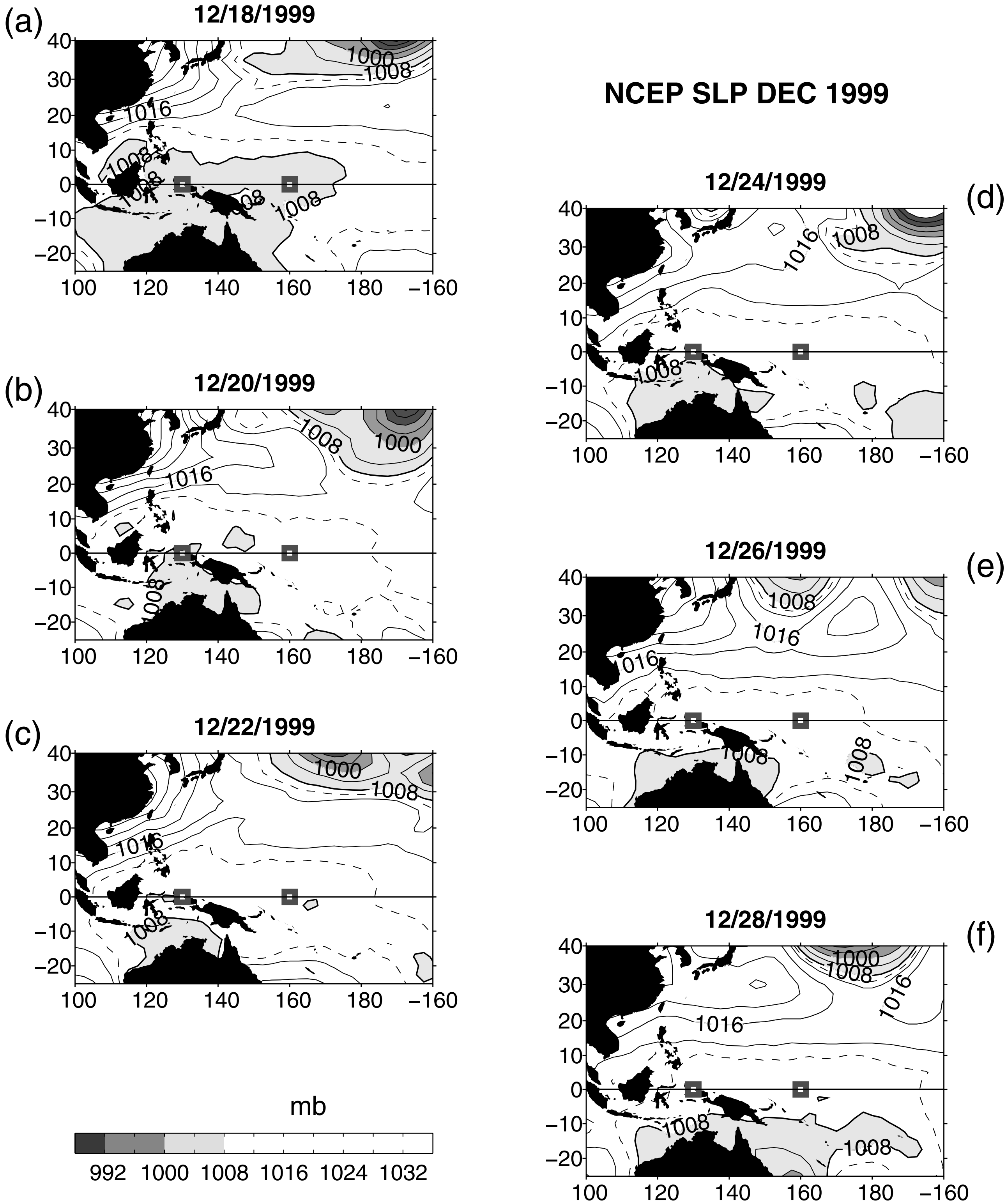


Figure 7. Same as Figure 6 except for the event in December 1999.

triggered the formation of cyclones and WWB event. Otherwise, if ∇P_{WEP} was absent in the background and the surge forcing could not excite it, the development of subsequent atmospheric event was halted. It appears that the preconditioning of the background ∇P_{WEP} might have played a role in regulating the generation processes of WWB. Yet, proving and establishing the mechanism is beyond the scope of this study.

3.2.4. Role of The Warm Pool in Preconditioning ∇P_{WEP}

[30] In the tropics, the location of low SLPs is closely related to the location of high SSTs [*Ramage*, 1971]. Such a

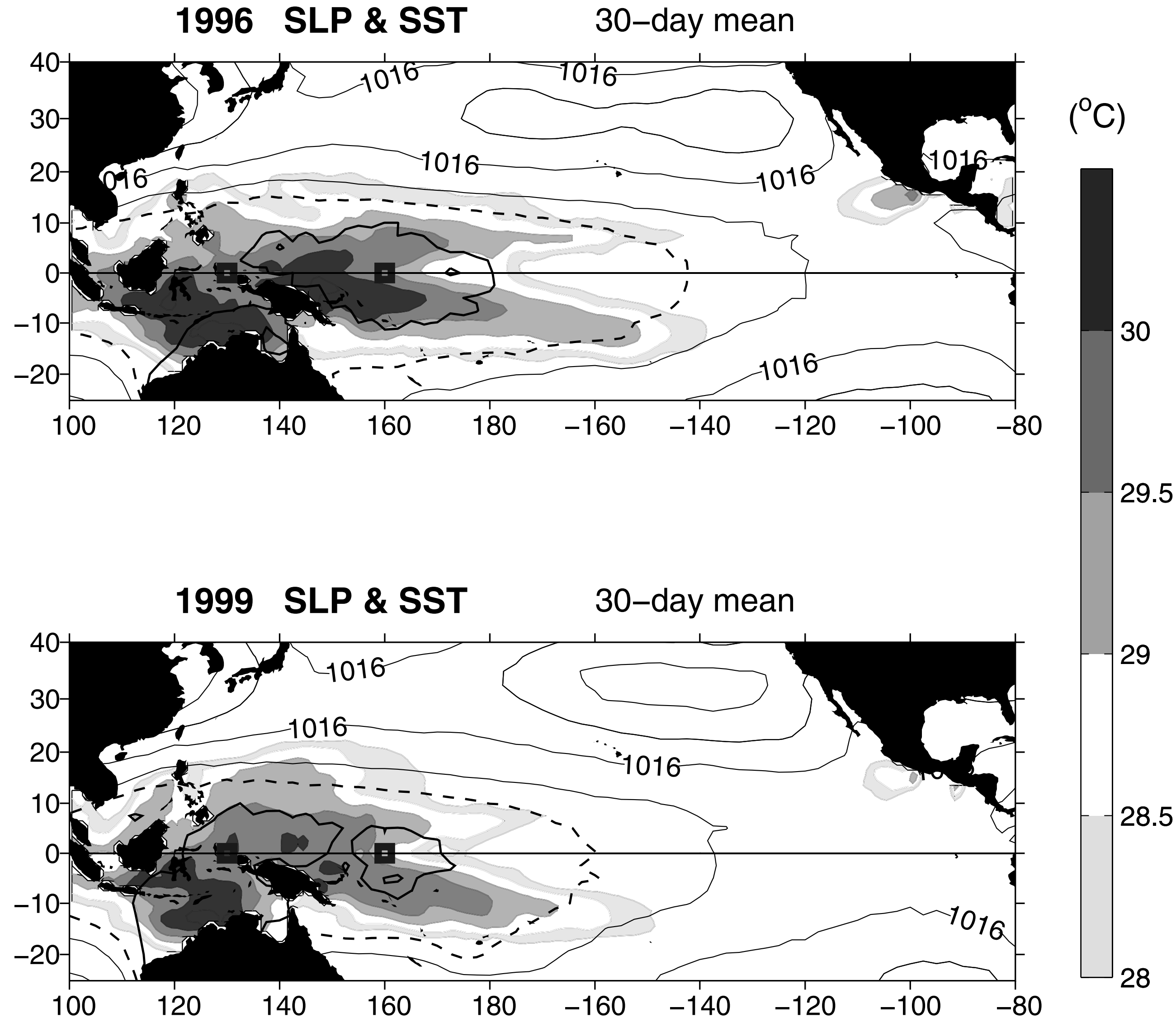


Figure 8. The warm water pool of the western Pacific averaged over the 30-day period prior to surge events in December (a) 1996 and (b) 1999. Only SSTs greater than 28°C are shaded. The SLP fields averaged in the same period are superimposed (contours). The 1008 mb isobar is highlighted and the 1010 mb isobar is dashed. The two gray squares on the equator indicate the two positions used to calculate ∇P_{WEP}.

relationship implies that the precondition of ∇P_{WEP} might be associated with the changes of the warm water pool in the WEP. To examine how the warm pool was involved in the process, Figures 8a–8b superimpose the mean SLP field (contours) over the mean SST field (colored isotherms of 28°C and higher are plotted) in the tropical Pacific basin. The fields were the averages over the 30-day period prior to the surges in Decembers 1996 and 1999, respectively. It is evident that there existed a seeming correlation between low SLPs and high SSTs in the western Pacific. Before the surge of December 1996, the warm water pool was large and its eastern edge on the equator was located east of the dateline. Correspondingly, the overlying low SLP center was large and displaced primarily in the east of New Guinea. However, before the surge of December 1999 the warm pool was small and displaced about 20° westward and so the associated low SLP center was reduced westward and confined largely in the west of New Guinea. The changes in the background appear to be the cause of the change in ∇P_{WEP}: ∇P_{WEP} existed prior to the surge in December 1996 but vanished in 1999. This result substantiates the indications of Figures 5a–5b that the background ∇P_{WEP} was preconditioned differently before the two surge events.

[31] *Wyrtki* [1975, 1985] was the first to point out that an accumulation of warm water in the west is a necessary precondition for an El Niño to occur. The composite phase patterns of *Rasmusson and Carpenter* [1982] showed that positive SST anomalies in the vicinity of the dateline are a major characteristic in the onset phase of a warm episode. The evolution of the SST field during the onset of the 1997 El Niño was consistent with the canonical pattern: the surface water was about 0.5°C warmer near the dateline associated

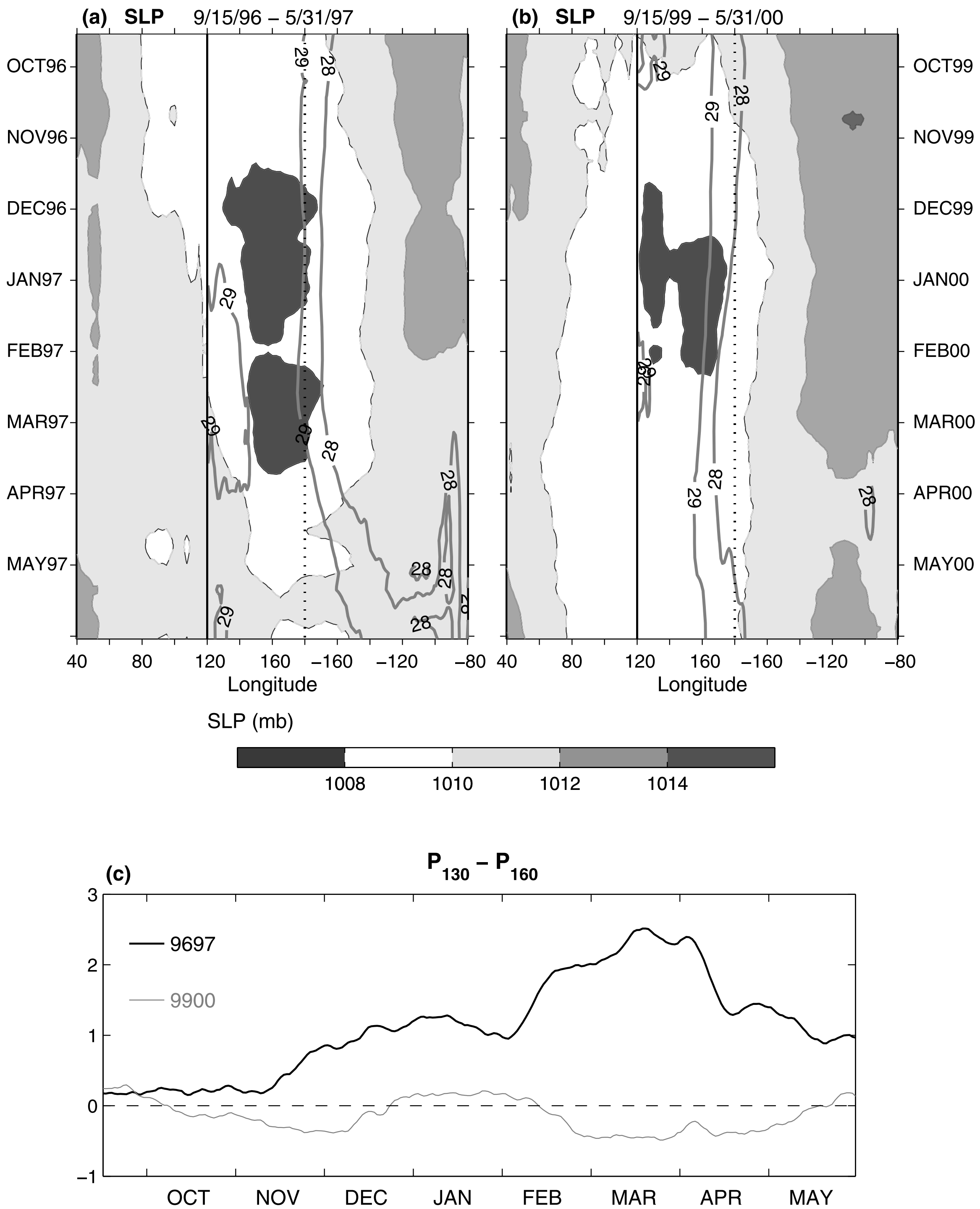


Figure 9. Time-longitude plots of the equatorial SLPs averaged between 5°S–5°N (shaded) in (a) in 1996/1997 and (b) 1999/2000 for the entire Indo-Pacific sector. A 30-day running mean was applied. The 1010 mb isobar is dashed. The superimposed thick gray contours in the Pacific sector are the 28-isotherms averaged between 5°S–5°N applied with a 5-point running mean. (c) The change of ∇P_{WEP} during the two corresponding periods.

with the expansion of a large warm pool (Figures 1a and 2a). A direct consequence of the build-up of the warm water pool appears to be the build-up of overlying low SLPs, which then enhanced ∇P_{WEP} (Figure 8a). Apparently, our analysis suggests that the onset phase of the 1997 El Niño fostered a background favorable for WWB events to develop.

[32] By contrast, the winter/spring of 1999/2000 of a La Niña mature phase tended to suppress high-frequency atmospheric transients because the westward displacement of the warm pool and low SLPs tended to restrain ∇P_{WEP}. Figures 9a–9b plot the evolution of the SLPs (shaded background) and the 28–29°C isotherms (thick gray lines)

averaged between 5°S and 5°N during the two respective ENSO phases. The plots were made for the entire equatorial Indo-Pacific sector from 40°E to 80°W to provide a full view of the structure associated with the east-west migration. The plots also applied a 30-day running mean to the daily mean SLP field and a 5-point running mean to the weekly SST field to filter out high-frequency motions so that the evolving background structure can be emphasized. The westward progression of the low SLP center associated with the mature phase of La Niña in the winter/spring of 1999/2000 contrasted sharply to the eastward displacement associated with the pre-El Niño in the winter/spring of 1996/1997. The plot of the variation of ∇P_{WEP} during the two target periods (Figure 9c) appears to support our analysis that ∇P_{WEP} was sensitive to the location of the low SLP center. Unlike the ∇P_{WEP} in 1996/97 that was persistently positive and strengthened with time, ∇P_{WEP} in the winter/spring of 1999/2000 was mostly negative.

[33] It is worthy noting that the eastward migration of the warm pool/low SLPs to the central Pacific in late spring (e.g., April–May) of 1997 (Figure 9a) did not further amplify ∇P_{WEP} (Figure 9c), although westerly wind events were growing stronger and frequent at that time. This might be due to the fact that the genesis of those winds resided largely in the basin-scale coupling between the atmosphere and ocean. As is shown in Figure 2, the westerly wind anomalies after March 1997 were centered on the western edge of the large positive SST anomalies of the eastern basin. ∇P_{WEP}, though it could be a likely preconditioning index for tropical cyclone formation related to extratropical atmospheric forcing, may not be representative in terms of describing the wind events resulting from basin-scale atmospheric adjustment to the changes in SST. Our central focus is on the WWB genesis during November–January. The regulating effect of the warm pool migration on ∇P_{WEP} during this period was suggested by the two cases analyzed here (Figure 9c). However, we note that this mechanism could not be proved or established in this study due to the limitation of the case analysis approach. Further statistical and dynamical study is required to test the veracity of the mechanism.

3.2.5. Location of New Guinea

[34] The island of New Guinea, consisting of Papua New Guinea on the eastern half and Irian Jaya, Indonesia on the western half, stretches southeast to northwest and joins the equator at about 130°E. Its location is emphasized in this study because it is a dividing line between the open ocean in the east and the island-studded Indonesian Seas in the west. A map detailing its geometric relationship to the Indonesian Archipelago and regional topographic features is shown in Figure 10a.

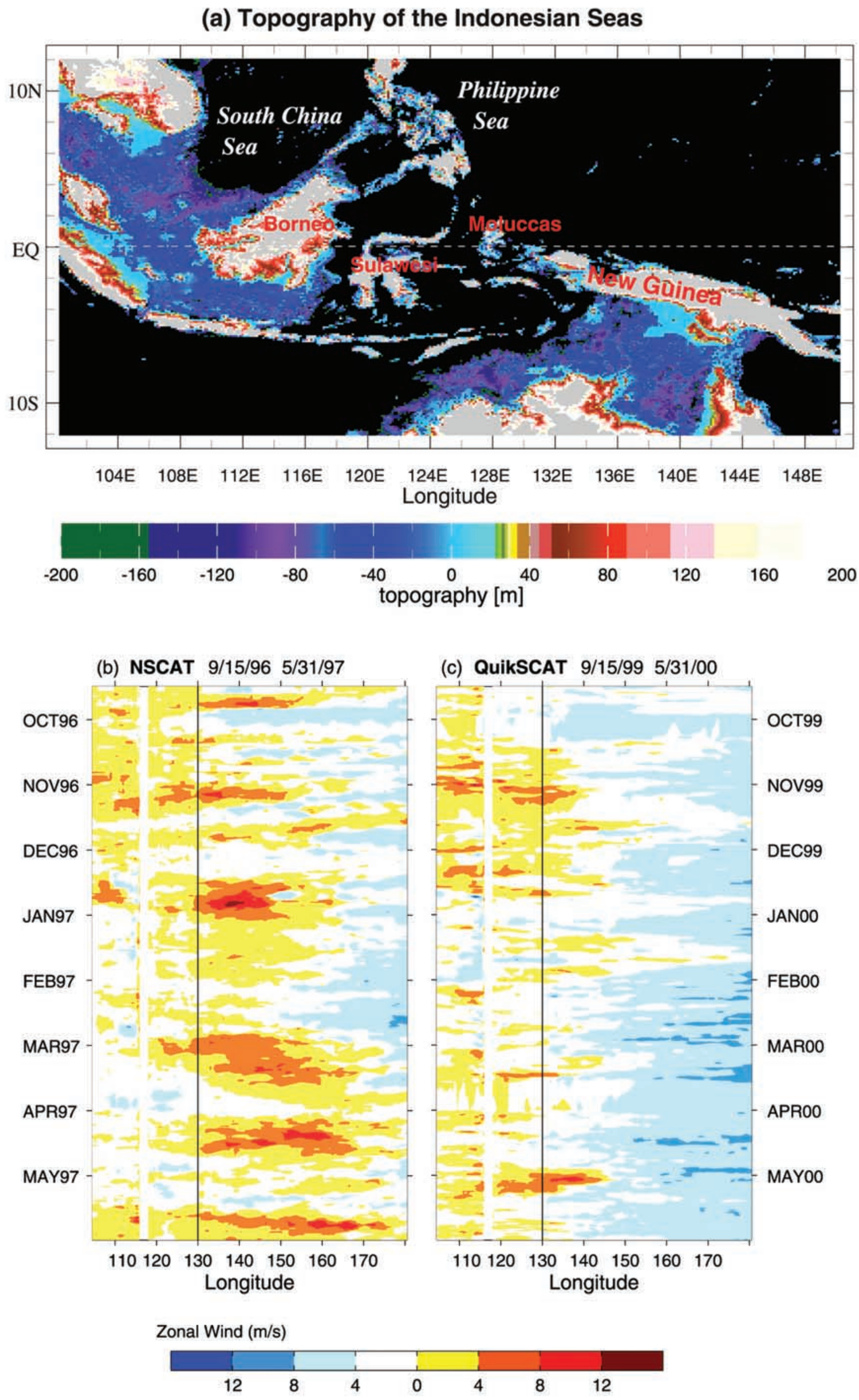
[35] Our analysis has indicated that the westward migration of the warm pool and associated low SLP center during the winter of 1999/2000 did not induce a response of ∇P_{WEP} between 130° and 160°E and hence did not lead to a WWB event over the open seas east of New Guinea. However, it should be noted that the westward shift of the background state was actually a westward shift of the action center; an SLP gradient existed in the background between 110° and 130°E (see Figure 9b) and the surge pathway was westward into the South China Sea (Figure 4). Westerly winds were still visible but they were mostly located west of 130°E, quite unlike in the winter/spring of 1996/97 that they were mostly located east of 130°E (Figures 10b–10c). Furthermore, the westerly winds generated in the far west were short-lived and had a short zonal fetch. In sum, the westerly winds generated during the two different ENSO periods differed greatly in the location, duration, and spatial extension.

[36] During the winter/spring of 1999/2000, no large-amplitude downwelling Kelvin waves were detected by TOPEX/Poseidon altimeter or TAO moored buoy measurements [*Climate Prediction Center*, 2001]. We speculated, but a detailed examination is beyond the scope of this paper, that the weak oceanic response might be due to two reasons. One is that the ocean response is a function of temporal and spatial variations of the wind [*Kessler et al.*, 1995; *Hendon et al.*, 1998]. A forcing with a short zonal patch and short duration has only a weak effect on ocean. The other is that downwelling Kelvin waves generated by westerly winds occurring in the west of New Guinea could be disturbed considerably by near equatorial islands. The Indonesian Archipelago have several near and cross-equator islands that have a north-south extent of at least 5° in latitude, much larger than the equatorial radius of deformation of about 250 km. *Cane and du Penhoat* [1981] obtained analytical solutions to compute the effect of the extent and position of the island on the propagation of low-frequency equatorial waves. Our computation based on their solutions [cf., *Cane and du Penhoat*, 1981, equations (2.17)–(2.18)] indicated that the transmission rate of incident Kelvin wave past the islands of Borneo (centered on 113°E), Sulawesi (120°E), Moluccas (128°E), and New Guinea is 0.54, 0.54, 0.75, and 0.85, respectively. Cumulatively, these chain islands represent significant blocking effect on the propagation of a Kelvin wave. This implies that the westerly winds appeared in the west of New Guinea may have limited influence on the ENSO system compared to those generated in the east of New Guinea. Our analysis supports the analysis of *Clarke and Van Gorder* [2001] that highlighted the usefulness of the wind variability within 130°–160°E for ENSO prediction.

4. WWB and the Onset of the 1982 El Niño

[37] The onset of the 1982 El Niño [*Halpern et al.*, 1983; *Rasmusson and Wallace*, 1983; *Lukas et al.*, 1984; *Keen*, 1988] featured also active WWB events, very similar to the onset of the 1997 El Niño. This can be seen from Figures 11a–11b, variations of the equatorial zonal winds and isotherms of 28°C and 29°C averaged in the 5°S–5°N band for both onset phases from the NCEP/NCAR reanalysis. Time-dependent wind fields for the onset phase of 1982 El Niño can only be obtained from atmospheric reanalysis products, however, those wind are known to have a weak bias in the tropical oceans [e.g., *Meisser et al.*, 2001]. The weaker winds might be caused by the lack of in situ/satellite observations in early 1980s that crippled the data-assimilation ability in simulating the strength of syn-

Figure 10. (opposite) (a) Topography of the Indonesian Seas. Water depth deeper than 200 m is darkened. (b–c) Time-longitude plots of the zonal winds averaged over the equatorial band (2°S–2°N) in 1996/1997 and 1999/2000, respectively.


(a) Topography of the Indonesian Seas
10N
EQ
10S
South China Sea
Philippine Sea
Borneo
Moluccas
Sulawesi
New Guinea
104E 108E 112E 116E 120E 124E 128E 132E 136E 140E 144E 148E
Longitude
-200 -160 -120 -80 -40 0 40 80 120 160 200
topography [m]
(b) NSCAT 9/15/96 5/31/97
OCT96
NOV96
DEC96
JAN97
FEB97
MAR97
APR97
MAY97
110 120 130 140 150 160 170
Longitude
(c) QuikSCAT 9/15/99 5/31/00
OCT99
NOV99
DEC99
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APR00
MAY00
110 120 130 140 150 160 170
Longitude
Zonal Wind (m/s)
12 8 4 0 4 8 12

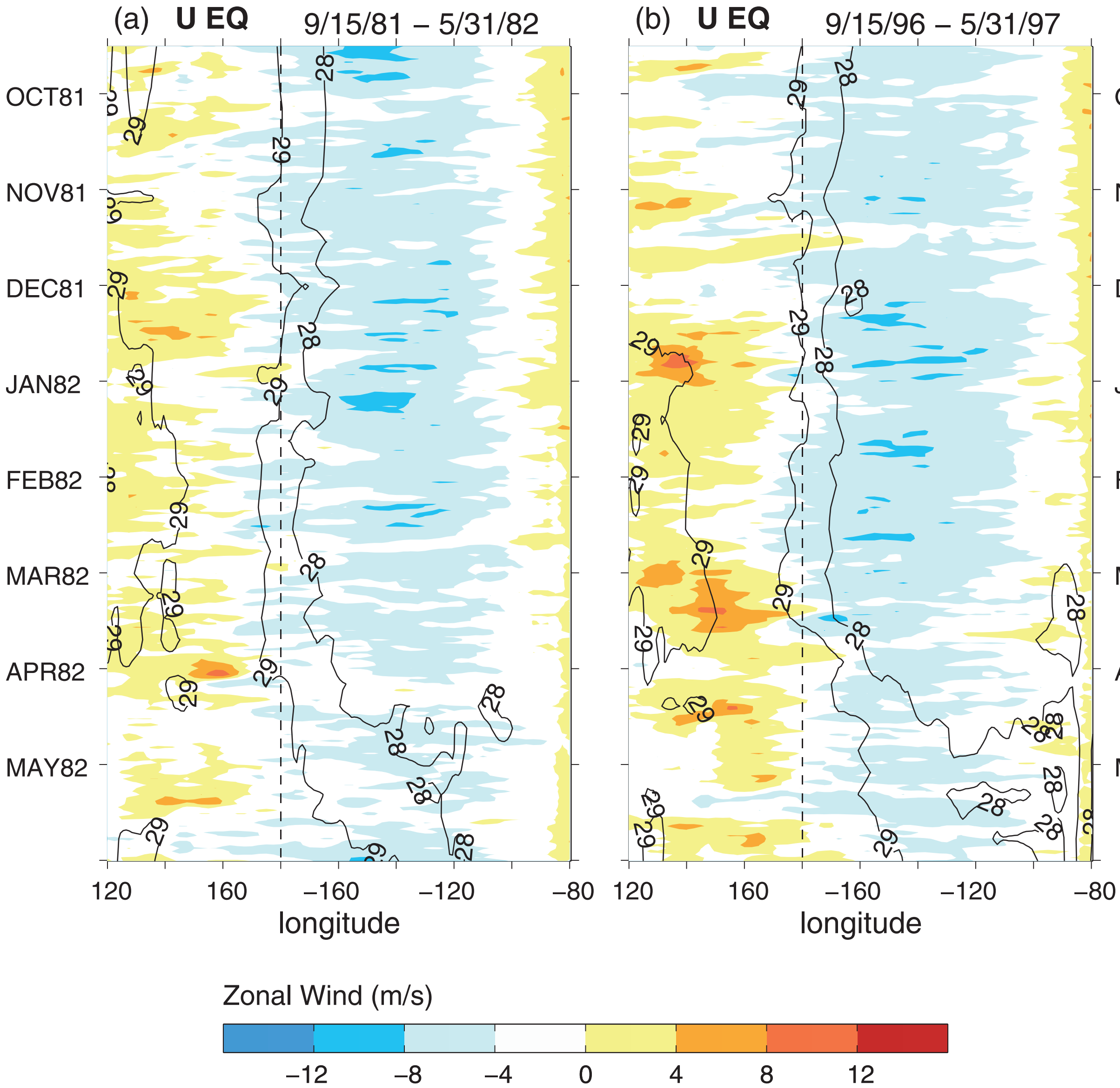

Figure 11. Same as Figure 1 but for (a) 1981/1982 and (b) 1996/1997 using winds from the NCEP/NCAR reanalyses daily products.

optic-scale events and by the use of a coarse spatial resolution (2.5°grid) that may not be able to capture synoptic-scale variability resolution. It is obvious that the amplitude of the NCEP reanalyses winds of 1981/1982 is weaker than that of 1996/1997, and the latter is weaker than satellite counterparts (Figure 1a). Nevertheless, the following features are clearly indicated by data.

[38] 1. The pattern of the westerly wind variability during the winter/spring of 1981/1982 was similar to that in the same period of 1996/1997. Strong WWB events occurred in December 1981, February 1982, and late March/early April of 1982, and the last event was followed by the eastward migration of the warm water pool.

[39] 2. During both onset phases, the eastern edge of the warm pool (e.g., the 28°C isotherm) was displaced to the east of the dateline. The generation of the WWB in December 1981 (Figures 12a–12g) was very similar to that in December 1996. The sequence of processes involved the enhancement of background ∇P_{WEP} by northerly surges, acceleration of an equatorial westerly flow to the east of the northerly winds, formation of individual/pared tropical cyclones, and amplification of the westerly winds near equatorial latitudes by the cyclonic circulation. Note in Figure 11a that the equatorial low was displaced east of New Guinea before the surge, a favorable condition that allowed the existence of ∇P_{WEP} and the subsequent cyclone development (Figure 6a).

[40] The similarity between the two onset El Niño phases further fuels the speculation on the possible role of ENSO in regulating the WWB activity. Yet detailed statistical analysis based upon all past onsets of El Niño events is needed before such a mechanism can be fully established.

5. Discussions and Summary

[41] This study suggests that the ENSO might have a regulating effect on the generation of WWB based upon the

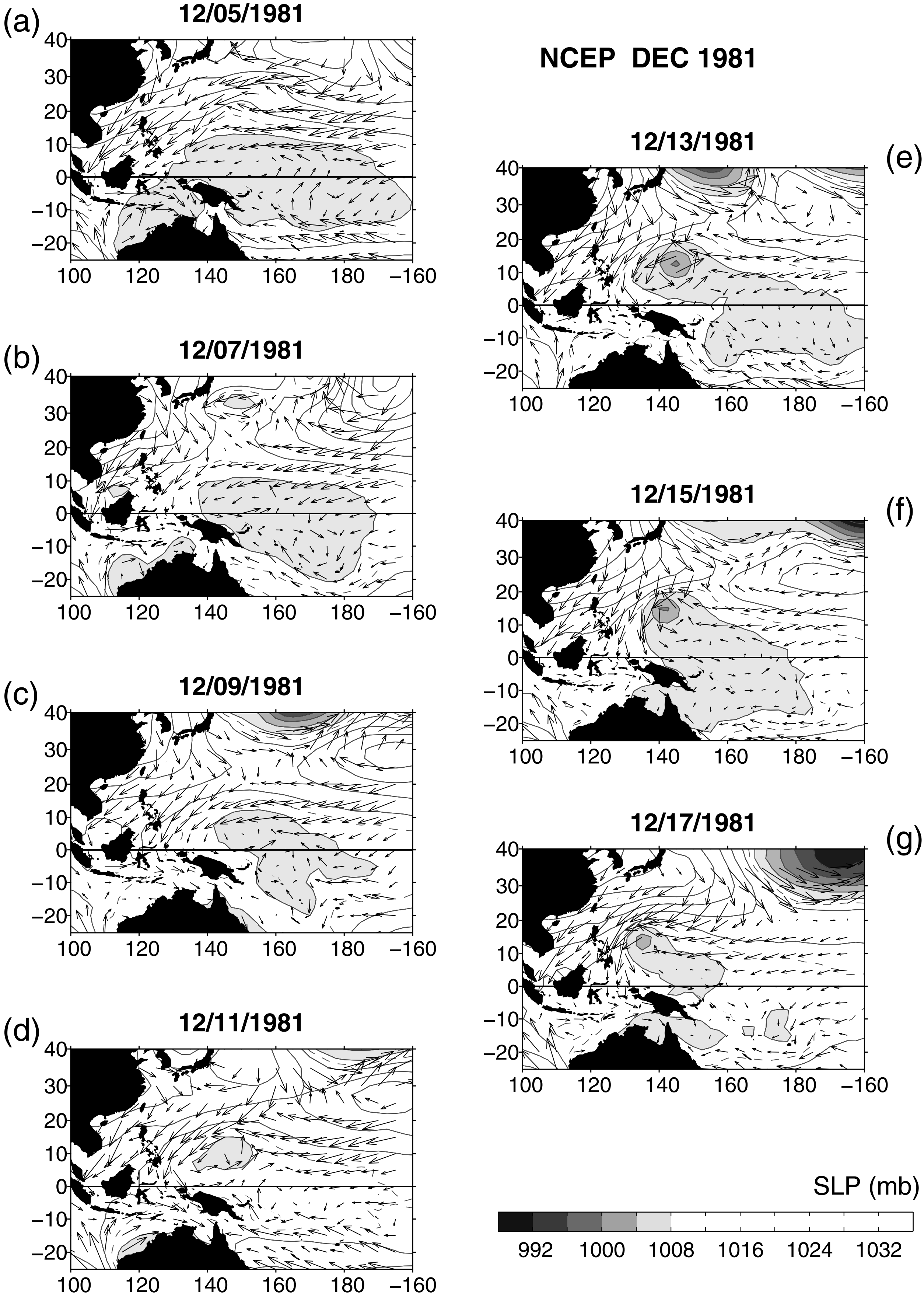

Figure 12. Evolution of daily surface wind fields (vector field) and daily SLP fields (shaded) in December 1981.

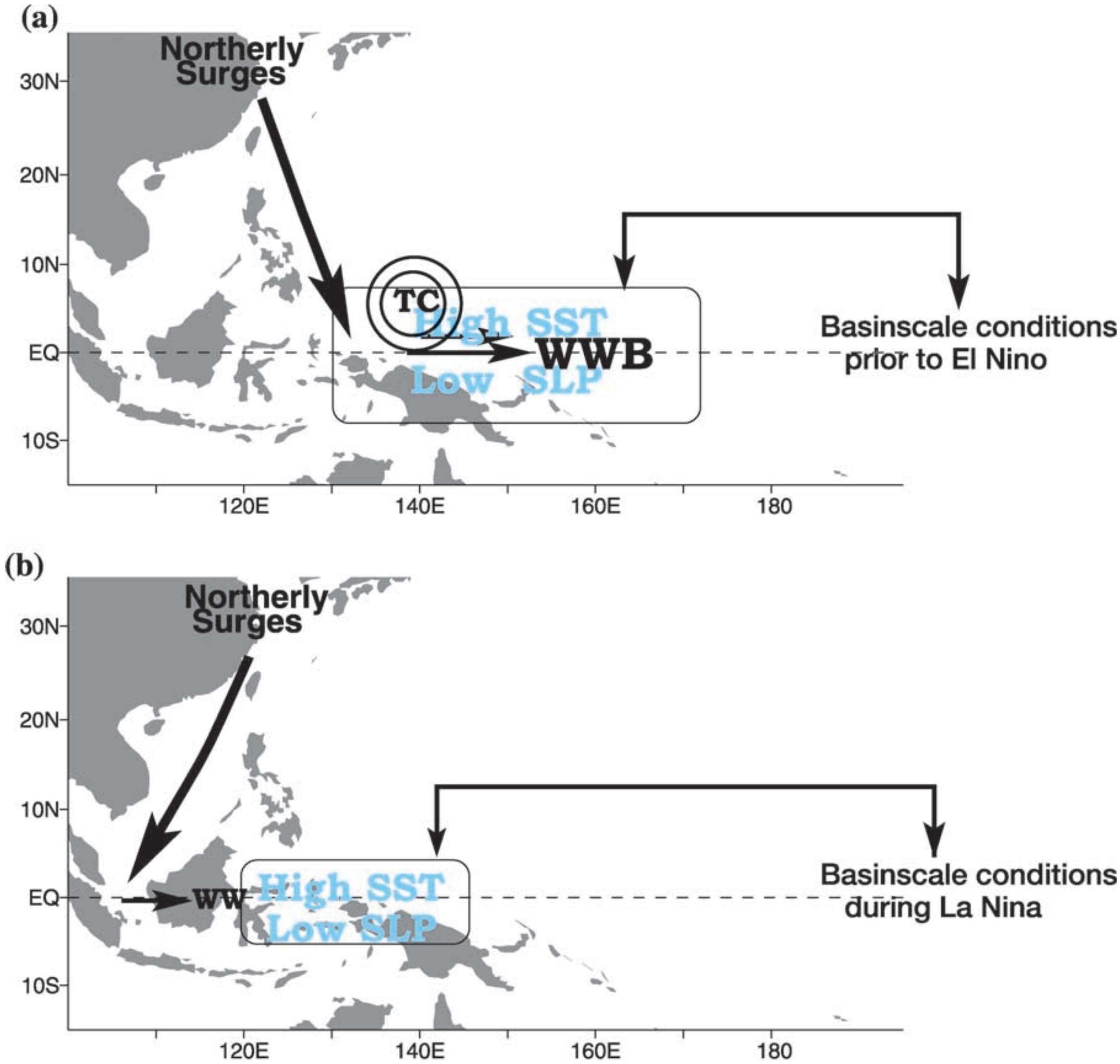

Figure 13. Schematic diagram depicting (a) the generation of WWB in 1996/1997 and (b) the dislocation of the processes in 1999/2000.

evidence from the case analysis. Two contrasting ENSO phases, i.e., the onset El Niño phase in the winter/spring of 1996/1997 and the La Niña mature phase in the winter/spring of 1999/2000 and also the onset of the 1982 El Niño were analyzed.

[42] The study indicates that WWB and ENSO, though having distinctly different timescales, appear to be structurally related. A key parameter could be ∇P_{WEP}, the west-to-east pressure gradient in the western equatorial Pacific (WEP) between 130°E and 160°E. ∇P_{WEP} is a preconditioning index for the formation of tropical cyclones in response to midlatitude atmospheric transient forcing. The cyclonic circulation on the equator is a leading contributor to prolonged WWB events. ∇P_{WEP} can be regulated by ENSO because the location of the low SLP center, which determines the strength of the ∇P_{WEP}, correlates with the location of the warm water pool and has an east-west migration in response to the ENSO phases. The case analysis shows that ∇P_{WEP} was promoted prior to El Nino when the warm pool was larger and extended to near the dateline, while it was suppressed during La Niña when the warm pool was small and displaced in the far west.

[43] A schematic diagram can be drawn to depict the chain action of the cyclone/WWB formation and the feedback with the evolving ENSO background for the two cases analyzed (Figures 13a–13b). During the onset phase of the 1997 El Niño, the warm pool was large and displaced east of New Guinea. In response, a large low SLP center was developed in the east of New Guinea that supported a positive ∇P_{WEP}. When a northerly surge occurred, the associated high pressures raised the SLPs in the near-equatorial region, which then enhanced existing ∇P_{WEP}, accelerated an equatorial westerly flow, induced the formation of individual/paired tropical cyclones and produced a large-amplitude WWB event. Those WWB were shown to have important local and remote implications for the El Niño development [*Lukas*, 1988]. In contrast, during the La Niña mature phase in the winter/spring of 1999/2000, the warm pool was small and shifted westward. Consequently, low SLPs were displaced westward that did not support a ∇P_{WEP} in the east of New

Guinea. The surge pathway was also shifted westward and could not alleviate the situation so that no cyclone formation followed the surge and no large-amplitude WWB occurred to the east of New Guinea. It appears that there were positive feedback in both situations: the El Niño background favored WWB which in turn tended to maintain El Niño, whereas the La Niña background unfavored WWB which in turn tended to sustain La Niña. However, this study serves only to point out the likely regulating mechanism of ENSO on the activity of high-frequency WWB. Proving and establishing such mechanisms are beyond the scope of the study.

[44] Our analysis further indicated that the westward shift of the system associated with the La Niña phase was actually a westward shift of the genesis of the WWB generation. Westerly winds with short duration and small spatial scales were visible in the west of New Guinea over the island-studded Indonesian Seas during the winter/spring of 1999/2000. However, no large-amplitude downwelling Kelvin waves were observed during the period. We speculated that the weak effect of those westerly winds on the open seas to the east of New Guinea might be due to both the weak amplitude and short duration of the winds and the blocking effect of near and cross equatorial islands in the Indonesian Seas on the propagation of equatorial waves.

[45] That only westerly winds generated in the east of New Guinea have potential importance to ENSO system further highlights the seemingly relationship between WWB and El Niño. This relationship has three noteworthy aspects. First, northerly surges, a forcing for a WWB formation, have a peak season from November to February [*Zhang et al.*, 1997] that spans the onset phase of El Niño. The frequency of the surges does not seem to have substantial changes with ENSO phases [*Zhang et al.*, 1997] although the surge pathway does [*Compo et al.*, 1999]. This implies that northerly surges can occur every winter no matter the phase of ENSO and are not a unique feature for the onset of a warm event. Second, our case analysis indicates that the ENSO phases appear to have a regulating effect on the equatorial responses induced by northerly surges. If so, the activity of WWB can vary with the ENSO phases even though the surge forcing presents every winter. Third, WWB have shown to influence the development of El Niño because the downwelling Kelvin waves excited by WWB during the onset period can play a detrimental role in triggering ENSO SST anomalies in the eastern Pacific [*Lukas et al.*, 1984]. However, WWB might not be a sufficient onset mechanism by themselves because they apparently appeared as a passive response to the warm pool condition as indicated in our case analysis. Hence the ENSO–WWB relationship is likely an interactive relationship with each influencing the other's development. However, they may not be the genesis for each other because WWB were induced by extratropical-tropical atmospheric interactions indicated from our analysis whereas ENSO resides in coupled ocean-atmosphere interactions in the tropical Pacific.

[46] It should be noted that there might exist a connection between northerly surges and the movement of the MJO convection in the region between the eastern Indian Ocean and the western Pacific [*Weickmann and Khalsa*, 1991; *Hsu et al.*, 1990; *Sui and Lau*, 1992], which could affect the timing and frequency of WWB events. These issues will be pursued in future studies. In addition, it has been known that the time-space evolution of El Niño changes with decade. The warm SST anomalies before 1976 propagate westward from the South American coast to central Pacific [*Rasmusson and Carpenter*, 1982] while the propagation of the warm SST anomalies reverses in recent decades: being eastward from the western-central to eastern Pacific [*Wallace et al.*, 1998; *Fedorov and Philander*, 2001]. The importance of WWB to the onset of El Niño at different decades and the regulation of WWB by the changing ENSO background are the topics being continually pursued.

[47] **Acknowledgments.** L. Yu is grateful to Gene Rasmusson for his insights on the WWB-ENSO relationship and his constant encouragement and generous support during the course of the study. Roger Lukas is sincerely thanked for many thought-provoking discussions on a wide range of topics including the western Pacific air-sea interactions, tropical-extratropical atmospheric interactions, and scale interactions and Eli Tziperman for constructive comments that improved the presentation of the manuscript. L. Yu thanks the support by the NASA's NSCAT and QuikSCAT science programs and the WHOI Assistant Scientist Awards. The scatterometer winds from NSCAT and QuikScat are obtained from the Seaflux project at JPL led by one of the coauthors (W.T.Liu). Reynolds SST data and the NCAR/NCAR reanalysis data are provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, from their website at http://www.cdc.noaa.gov/. The equatorial zonal wind and SST data for Figure 2 are taken from the website at http://www.pmel.noaa.gov/tao/. This is contribution 10485 from the Woods Hole Oceanographic Institution.

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