Strong Ocean–Atmosphere Interactions during a Short-Term Hot Event over the Western Pacific Warm Pool in Response to El Niño

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(Manuscript received 24 August 2015, in final form 23 February 2016)

ABSTRACT

A short-term hot event with a very high sea surface temperature (SST $\geq 30^\circ$C) occurred in the western Pacific warm pool during November 2006. The interactions between this ocean hot event, atmospheric convection, and large-scale dynamics are studied using satellite observations, buoy measurements, air–sea fluxes analysis, and global reanalysis. It is shown that SST variation and deep convection over the western Pacific behave like a remote response to the El Niño warm SST anomaly in the central Pacific that induces westward-moving atmospheric convection and equatorial waves. The large-scale subsidence associated with propagating convection not only promotes high SSTs in the western Pacific through establishing cloud-free conditions and increasing heat content in a thin ocean mixed layer, but also produces convective instability through capping substantial water vapor in the lower troposphere. Under the precondition of convective instability and the steering of tropical easterlies, some convective systems propagate coherently from the central to western Pacific and intensify. In particular, new cloud clusters are dynamically attracted to the warmest oceans with maximum atmospheric instability. The enhanced convective activity then transfers oceanic energy into the atmosphere, strengthens upper-ocean mixing, and returns the positive SST anomalies to more typical values. In such a coupled system, synoptic-scale convective activities at an interval of 5–8 days are selectively amplified and thus are filtered to an intraseasonal (20–30-day) oscillation, depending on the phase of the hot event over the western Pacific. The observed evidence has implications for the predictability of short-term climate, and it offers critical information for validating the coupled ocean–atmosphere dynamics in climate models.

1. Introduction

The tropical western Pacific is the warmest open ocean in the world, with an annual-mean sea surface temperature (SST) greater than 29°C. As the western Pacific warm pool underlies intense atmospheric convection, it provides a major source of energy for driving the general circulation of the atmosphere (e.g., Webster 1994; Anderson et al. 1996). It has long been recognized that the coupled ocean–atmosphere dynamics over the warm pool plays an essential role in the seasonal monsoon and interannual variability, such as the El Niño–Southern Oscillation (ENSO). In the recent two decades, it is increasingly apparent that the air–sea interactions are also pronounced in variability on short time scales ranging from diurnal to intraseasonal (Webster and Lukas 1992; Chen et al. 1996; Sui et al. 1997; Bernie et al. 2005). The understanding of the multiscale processes that couple SST variation and atmospheric convection over the warm pool can be crucial to improve our prediction of climate variability.

One important feature of the air–sea interactions over the warm pool is that the relationship between high SST and deep convection is strongly dependent on large-scale circulation. On average, the frequency and intensity of deep convection increases dramatically at
SSTs of 26°–28°C (e.g., Graham and Barnett 1987). At SSTs above 29.5°C, however, deep convection tends to reduce with increasing SST (Waliser and Graham 1993). Such an intriguing feature, in fact, signals that very high SST mostly occurs under conditions of suppressed convection, which are associated with large-scale subsidence. Because the subsidence, as part of the atmospheric circulation, is greatly governed by organized deep convection over nearby or remote regions, the formation of high SST can be viewed as a response to remote forcing (Waliser 1996; Lau et al. 1997; Tompkins 2001). On the other hand, given changes in large-scale rising motion and moisture convergence, deep convection tends to develop and/or moves to the warmest oceans, and it contributes to a feedback cooling of the ocean surface through enhanced cloudiness and surface heat exchange (Lau et al. 1997; Fasullo and Webster 1999; Li et al. 2000). 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intraseasonal, seasonal, and interannual variability (Waliser 1996; Qin et al. 2007). On the intraseasonal time scale, the evolution of hot spots particularly during TOGA COARE has been linked to the organized convection of the MJO (Lau et al. 1997; Johnson et al. 2001; Duvel and Vialard 2007). However, it is still unclear how hot spots interact with organized convection under other conditions, such as ENSO. During El Niño, the preferred occurrence of organized convection shifts to the central Pacific. It seems that the eastward-shifted convection can serve as a remote forcing on the western Pacific. Qin et al. (2008) reported a short-term hot event in the western Pacific during November 2006 when El Niño occurred (referred to as HE0611). During that period, the MJO is mainly located in the Indian Ocean and does not reach the Maritime Continent and Pacific Ocean (www.bom.gov.au/climate/mjo/; Wheeler and Hendon 2004). The analysis of heat fluxes shows that the very high SST in the western Pacific is induced by strong solar radiation and suppressed latent heat loss. Qin and Kawamura (2010) further suggested that the HE0611 formation is related to cloud-free conditions with large-scale subsidence over the western Pacific, which are caused by deep convection over the central Pacific. These studies, while focusing on an oceanic response to atmospheric forcing, arouse our curiosity on the short-time scale subsidence over the western Pacific, which are caused by deep convection over the central Pacific. Specifically, how does an ocean hot event interact with the evolving convective systems over the western Pacific on short time scales? How can this local air–sea coupling be modulated by the influence of the El Niño warm SST anomaly and deep convection over the central Pacific? To what extent can the above processes regulate the oscillations of the ocean–atmosphere system over the western Pacific?

The widely accepted but poorly quantified importance of the air–sea interactions in the very high-SST phenomena has triggered renewed interest. Using various datasets available recently, we report a kind of short-term interaction between high SST, deep convection, and large-scale dynamics during an observed hot event, which can be distinct from those relating to the MJO previously reported. The goal is to provide observational evidence to infer the role of the El Niño–related positive SST anomalies in forcing short-term responses of the ocean–atmosphere system and climate variability over the western Pacific warm pool. Section 2 describes the datasets and methods used in this study. Section 3 presents the joint behaviors of SST and atmospheric convection during the life cycle of the hot event. Section 4 examines the large-scale dynamics and atmospheric stability that support the growth/propagation of convective systems over the warm oceans. Section 5 estimates the generation of convective instability and intraseasonal oscillation to clarify the vital linkage between the hot event and deep convection. Section 6 investigates the oceanic response to convective activities and discusses the role of SST regulation in regional climate. Then the conclusions are presented.

2. Data and methods

To describe the oceanic and atmospheric conditions, we use a variety of data sources, including satellite observations, buoy measurements, air–sea flux analysis, and global reanalysis products. First, to monitor a large-scale thermal condition of the ocean surface, we use the Merged Satellite and In Situ Data Global Daily SST (MGSST) provided by the Japan Meteorological Agency. Following Qin et al. (2007), the hot event is defined as the very high SST above 30°C that can occupy an area greater than $3 \times 10^6$ km$^2$ and last for longer than 10 days. To identify atmospheric convection, we use the cloud products from the Japanese Geostationary Multi-transport Satellite (MTSAT). The high-level cloudiness is detected using the brightness temperature from the 11-μm channel of MTSAT. If the brightness temperature at any pixel is less than 260 K, that pixel is categorized as cloudy (Qin and Kawamura 2010). The hourly cloud-motion winds are also applied to present the upper-level circulation associated with deep convection. To estimate the rainfall amount of moist convection, we use the latest product of the Tropical Rainfall Measurement Mission (TRMM 3B42, version 7), which provides the 3-hourly mean rain rate with a resolution of 0.25° longitude/latitude (Huffman et al. 2007).

Because conventional soundings are sparse over oceans, observing the atmospheric processes associated with a hot event is quite difficult. In this study, we use the sounding data from the Atmospheric Infrared Sounder (AIRS) along with the Advanced Microwave Sounding Unit (AMSU) onboard the Aqua satellite, which is one of the most advanced temperature- and humidity-sounding systems deployed in space (Parkinson 2003). The AIRS sounding product provides three-dimensional maps of the air temperature, water vapor, and cloud properties on a daily time scale. The data quality over oceans has been verified by many previous studies (e.g., Tian et al. 2006). The standard product (Level 3, version 005) provided by the NASA Jet Propulsion Laboratory is used here. It includes the profiles of temperature and water vapor mass mixing rate $q$ at 12 pressure levels from 1000 to 100 hPa, with a spatial resolution of 1° longitude/latitude. The ascending and descending satellite paths are averaged to derive the daily mean. The AIRS product
has been applied to describe the variations of temperature and water vapor over the tropical Pacific (Qin and Kawamura 2009a; Takahashi et al. 2013), and its usage is further extended here.

To measure convective instability in the atmosphere, we estimate the convective available potential energy (CAPE; Moncrieff and Miller 1976). The CAPE presents the energy released by an air parcel when it is lifted, following an undilute adiabat, from the level of free convection (LFC) to the level of neutral buoyancy (LNB). Following Emanuel (1994), CAPE can be estimated as an integration of the difference between the equivalent potential temperature \( \theta_e \) of an air parcel from the source level and the saturation equivalent potential temperature \( \theta^* \) of ambient air aloft. It should be noted that a large value of CAPE does not necessarily lead to convection occurrence, because an air parcel usually needs to overcome a stable layer between the surface and LFC. The needed energy to lift the air parcel to its LFC for convection initiation is the convective inhibition (CIN). The CIN thus presents a convection barrier that prevents convection even though high CAPE exists. Here, we calculate CAPE and CIN using the AIRS profiles of air temperature and water vapor. We also estimate the vertical profiles of potential temperature \( \theta_e \), \( \theta^* \), and \( \theta^*_e \) so that we can use their configuration to reveal the processes responsible for CAPE generation.

To study the large-scale atmospheric conditions, we use the Japanese 55-year Reanalysis Project (JRA-55; Kobayashi et al. 2015). JRA-55 is one of the new-generation global reanalyses, and it has an improved data quality even on a diurnal scale (Chen et al. 2014a). The pressure-level analysis data used here have a spatial resolution of 1.25° longitude/latitude and a vertical interval of 25 hPa below 700 hPa and 50 hPa above. We mainly use the variables of geopotential height, horizontal winds, vertical velocity, and humidity to depict the activities of the subtropical high, synoptic-scale waves, tropical easterlies (including trade winds), low-level convergence, large-scale rising motion/subsidence, and atmospheric circulation.

To monitor the sea surface conditions and upper-oceanic processes, we use the measurements from the Tropical Atmosphere Ocean (TAO/TRITON) buoy array. The variables include the near-surface winds, temperature, humidity, radiation, and the vertical profiles of seawater temperature (McPhaden et al. 2009). To analyze the heat budget at the ocean surface, we also employ the objectively analyzed air–sea fluxes (OAFlux) products offered by the Woods Hole Oceanographic Institution (Yu and Weller 2007). The variables include the daily mean radiation fluxes, the latent and sensible heat fluxes, and the derived net heat gain.

3. Characteristics of the ocean thermal condition and atmospheric convection

a. Variations of SST and rainfall during the life cycle of HE0611

The HE0611 event occurred over the tropical western Pacific during 7–24 November 2006 (Qin et al. 2008). Here, we revisit the evolution of HE0611, with emphasis on the joint behaviors of SST and atmospheric convection. Figure 1 shows the spatial distribution of the averaged SST at the mature stage of the hot event (15–22 November). It is shown that the hot event consists of two major areas of very high SST (viz., HE0611-West

![Figure 1](image-url)
and -East). The maximum SST over HE0611-West (-East) appears in the oceans to the north of New Guinea (near the date line). HE0611-East has a larger area and higher magnitude of SST than HE0611-West. Such an eastward-shifted region of high SST corresponds to an occurrence of an El Niño event in the winter of 2006.

To show the temporal evolution of the hot event, we examine the measurements from the buoys moored in the highest-SST areas of HE0611-West and -East. Figure 2a shows that the SST over HE0611-West features one period of increase (7–23 November) and two periods of decrease (3–5 and 24–25 November). The SST variation has a range from 20.8°C to 31.0°C relative to the climatological November mean (29.5°C). In contrast, the SST variation over HE0611-East has a very small range of 0.2°C through November 2006 (Fig. 2b). The SST stays at a high value of 30.4°C, which is about 1.8°C above the climatological November mean (28.6°C).

The different behaviors of the SST between HE0611-West and -East in Figs. 2a and 2b are associated with ocean thermal properties and atmospheric conditions. Over HE0611-East, the persistent high SST is mainly sustained by an anomalous deep layer of hot seawater in the upper ocean with a 28°C depth of ~110 m (not shown), in relation to El Niño (Qin and Kawamura 2010). Corresponding to the El Niño SST anomaly, active rainfalls appear near the date line (Fig. 2d). The accumulated rainfall amount in November 2006 is ~30% more than the climate mean. Strong rainfall events usually last for ~2 days and are repeated at an interval of 5–8 days. It seems clear that the frequent activities of organized deep convection occur over the tropical central Pacific as a response to the El Niño positive SST anomalies.

Over HE0611-West, the layer of hot seawater, with a 28°C depth of ~60 m, is much shallower than that over HE0611-East (not shown). Figures 2a and 2c show that the SST warming trend during 7–23 November occurs with suppressed rainfalls, while the SST cooling trend on 3–5 and 24–25 November is coincident with active rainfalls. Thus, the variations of both SST and rainfalls exhibit a short-term oscillation with a period of 3–4 weeks. This feature over HE0611-West differs from that over HE0611-East, where the persistent positive SST anomaly produces high-frequency rainfall events. It offers an interesting case for studying the physics underlying short-term air–sea interactions. Previous studies had attributed the formation of high SST in HE0611-West to the forcing of deep convection over HE0611-East (Qin et al. 2008; Qin and Kawamura 2010). In the next sections, we pay more attention to the feedback and response of high SST to the evolving convective systems, especially at the mature and decay stages of the hot event, so that a full cycle of the air–sea coupling is clarified.

b. Evolution of the convective systems during the life cycle of HE0611

To examine the convection evolution, we calculate the longitude–time variations of observed rainfall during November 2006. Figure 3a shows that convective systems repeatedly occur over the zone of 170°E–160°W; most of them exhibit a feature of westward movement. Two organized convective systems reach near the date line on 5–10 and 14–15 November, but they dissipate when moving to the west of 160°E. The decayed convection clearly explains that HE0611-West is dominated by a dry spell during 11–22 November, as marked by the dashed circle in Fig. 3a. Figure 3a also shows that a convective system forms at 160°–170°W on 19 November. Compared to the previous two systems, this
migrating system has a longer duration, and it exhibits more coherent movement from HE0611-East to -West (dashed line in Fig. 3a). The convective system propagates by a distance of 43°–53° longitude in 6 days, indicating a phase speed of 9–11 m s⁻¹. It produces the most intense rainfall near 147°E during 24–25 November, which displays a strong convection growth over the highest-SST ocean near 2°N. Consequently, the convective system produces the most intense rainfall in these hours, as shown in Figs. 2c and 3.

A comparison of the growing and mature phases of the cloud cluster shows that it tends to move southwestward (i.e., toward the warmest ocean near New Guinea) (Figs. 5a–c vs Figs. 5g–i). Such a movement of the cloud cluster is clearly displaced from its parent convective system that continues moving westward through the period, as shown by those cloud clusters at 8°–10°N. Therefore, the high SST over HE0611-West appears not only to promote convective growth, but also to attract the propagation of a new cloud cluster.

The cloud cluster gradually dissipates during 1200–2000 UTC 25 November (Figs. 5j–l) and is replaced by the new ones later (Fig. 3a). Along with the passage of this cloud cluster, the SST over HE0611-West undergoes a large decrease (0.8°C) during 24–25 November (Fig. 2a). The MGD SST data also show that the hot event comes to an end on 24 November, when the area size of the very high SST over HE0611-West declines to ~2.5 × 10⁶ km² and becomes disconnected from the high-SST area of HE0611-East (not shown). As a summary, we see that the hot event and migrating convection exhibit strong joint behaviors, particularly over HE0611-West.

### 4. Atmospheric conditions that support convection propagation and growth

To reveal the governing processes of convection, we focus on the convective system that exhibits coherent
propagation and growth during 22–24 November when SST maximizes. Two questions are addressed here: Is the convection passively migrating over the warm oceans, or is it dynamically attracted to the high-SST areas? Why does it grow strong and differ from the previous systems that dissipate over HE0611-West?

First, we examine the activities of the subtropical high and associated tropical easterlies. Figure 6a shows that the subtropical high, as outlined by 5880-gpm contours, is mainly established to the west of 155°E on 18–20 November. It shifts eastward on 21 November, with its eastern flank extending to the date line. On 22–24 November, the body of the subtropical high is considerably enlarged, and the maximum geopotential height increases from 5890 to 5910 gpm. It exhibits a west–east-oriented pattern, with an axis at ~18°N stretching for more than 50° longitude. Such a belt-shaped subtropical high becomes favorable for the formation of wave disturbances. At the southeast of the subtropical high, a weak trough like an inverted “V” migrates westward on 22–24 November, as marked by the triangles in Fig. 6a. Such a wavelike disturbance is regarded as an easterly wave in the tropical easterlies (e.g., Reed et al. 1977; Gu et al. 2004). The easterly wave travels 8°–9° longitude per day, with a phase speed of 10–11 m s$^{-1}$. A comparison of Figs. 4, 5, and 6a suggests that the easterly wave is collocated with the westward-moving cloud clusters, which is similar to the features over the Atlantic Ocean (Payne and McGarry 1977; Machado et al. 1993; Dieng et al. 2014). An analysis of the geopotential height anomaly (not shown) further shows that the easterly wave is gradually deepening on 22–24 November when the cloud clusters grow strong. It seems clear that convective diabatic heating is in a position to enhance the wave development (Reed et al. 1977; Gu et al. 2004; Hsieh and Cook 2007), which manifests an active feedback of the hot event onto large-scale dynamics. It also relates to a transition of a westward-moving mixed Rossby–gravity wave (section 5c). The easterly disturbance, however, is absent on 11 and 16 November when migrating convection arrives at the east of HE0611-West (not shown). Such a difference seems to explain that the migrating convection is more organized during 22–24 November than those in the middle of November (Fig. 3a).

Figure 6b shows the evolution of zonal wind averaged over the migratory path of convective systems. At 500 hPa, the tropical easterlies strengthen during three years.
FIG. 5. The 4-hourly variations of the high-level clouds and cloud-motion vectors from 0000 UTC 24 Nov to 2000 UTC 25 Nov 2006. The valid time is marked in the bottom right of each figure. With cloud tracking by eye inspection, the clouds in red in a moving zone highlight the cloud cluster that undergoes a rapid growth over the high-SST ocean of HE0611-West. The open circle denotes the buoy moored at 2°N, 147°E.
periods (8–9, 14–15, and 21–24 November) and correspond to three convective episodes. The wind speed increases from 7–8 to 10–14 m s$^{-1}$, which is consistent with the strengthened subtropical high (Fig. 6a) that enhances the pressure gradient by 50%–60% (not shown). The easterly winds at 900 hPa also strengthen during these three periods and display the surges of trade winds (Fig. 6b). It is noted that the strong trade winds usually have a lag of several hours to the enhancement of midlevel winds. Such a lag suggests that the strengthened subtropical high may play a role in enhancing the trade winds. The mechanisms for the variations of the subtropical high, while interesting in the aspect of large-scale dynamics, are beyond the scope of this study and can be viewed as an external force.

During 21–24 November, the easterly winds below 400 hPa have a magnitude of 10–13 m s$^{-1}$. The wind speed agrees fairly well with the phase speeds of the convective system and easterly wave, which indicates a strong steering effect of the midlevel easterlies. We also note that the easterlies below 600 hPa are strongest during 21–25 November, and the strong trade winds can last for a long period of ~5 days.

To reveal the lower-tropospheric conditions, we examine the spatial distributions of the trade winds and horizontal convergence at 900 hPa. Figure 7a shows that, on 20 November, the trade winds are still in a suppressed phase. The relatively strong winds of 10 m s$^{-1}$ appear to the north of HE0611-West (8°–18°N, 135°–160°E), likely because of the westward-located subtropical high (Fig. 6a). The low-level convergence mainly occurs at the Maritime Continent, whereas the trade winds converge with the flows from the Southern Hemisphere. A cyclonic circulation is dominant near the date line, where the convective system develops.

Figure 7b shows that, on 22 November, the trade winds strengthen at a large scale over the western Pacific (along the zone of 6°–18°N). A large increase of wind speed (~6 m s$^{-1}$) occurs at 10°–12°N, 155°–165°E; the maximum wind speed reaches a magnitude of 16 m s$^{-1}$. Such a surge of trade winds is caused by the sudden
eastward shift and enhancement of the subtropical high on a synoptic time scale that strengthens the pressure gradient at its southern flank and increases zonal wind through the midlower troposphere (Fig. 6). Correspondingly, the low-level convergence is strengthened in a belt-shaped zone at ∼6°N between the strong trade winds and cyclonic circulation.

Figure 7c shows that, on 24 November, the cyclonic circulation moves into HE0611-West, with the strongest trade winds occurring at its northern flank. Strong low-level convergence is seen in the vicinity of cyclonic circulation. The convergence greatly contributes to the spatial patterns of moisture convergence (∼−2 × 10⁻⁴ g kg⁻¹ s⁻¹) and large-scale ascent (∼−0.1 Pa s⁻¹); thus, it is favorable for convective growth over HE0611-West. Similar west-moving disturbances also appear near the surface in both JRA-55 and satellite observations (not shown). Such a change in large-scale rising motion and moisture convergence also allows high SST to induce deep convection (Graham and Barnett 1987; Lau et al. 1997; Sui et al. 1997). Therefore, a combination of the steering midlevel easterlies, easterly wave, and low-level convergence/ascent in the enhanced trade winds seems to explain well the coherent migration of the convective system at the mature stage of the hot event.

While most of the cloud clusters migrate westward in the tropical easterlies, a cloud cluster moves southwestward over HE0611-West, and it grows explosively within several hours (Fig. 5). The convection develops at the center of cyclonic circulation near the equator (Fig. 7c), whereas the midlevel easterlies mainly appear to the north of 6°N (not shown). Despite a lack of steering flows, the cloud cluster seems to be attracted to the warmest ocean near New Guinea, which may relate to a large amount of moist energy and atmospheric instability over there. To address this issue, we estimate the evolution of CAPE and CIN over HE0611-West.

Figures 8a and 8b show that, during 19–22 November, the CAPE significantly increases over the warmest ocean near New Guinea. The CAPE value reaches a maximum of up to 2200 J kg⁻¹ during 23–24 November prior to convective growth (Fig. 8c). The variation of CIN is opposite to that of CAPE, with a minimum during 22–24 November (not shown). The large CAPE and decayed CIN indicate a strong convective instability. With the low-level ascent by an enhanced convergence (Fig. 7c), the air parcel may be lifted to overcome small CIN, triggering new convection. It is also noted that the convection movement induced by convective instability may have a speed of approximately $0.3\sqrt{\text{CAPE}}$ relative to midlevel winds (Moncrieff and Miller 1976; Chen et al. 2014b). A CAPE value of 1800–2200 J kg⁻¹ with weak ambient wind gives a propagation speed of 13–14 m s⁻¹.

As the large CAPE is located to the southwest of a newborn cloud cluster, it is conducive to a southward movement of convection (Figs. 5 and 8c). Such an effect of convective instability may explain the observed cloud cluster that moves southwestward and is clearly displaced from its northern counterparts. The CAPE value decreases during 25–26 November after the passage of the cloud cluster (Fig. 8d), indicating a decay of convective instability.

To depict an impact of instability on convection intensity, we further compare the spatial pattern of the CAPE precondition with the activities of the cloud cluster. Figure 9a shows that high CAPE is established mainly over the warmest oceans and reaches the maximum at a rectangular region of 2°S–4°N, 142°–152°E where SST is highest. Figure 9b shows that the cloud cluster arrives at the northeast of the high-CAPE area at 0000 UTC 24 November. Figures 9c and 9d show that the cloud cluster grows explosively after 1600 UTC 24 November, when it moves to the area of the maximum CAPE. Therefore, this cloud cluster, in terms of both movement and intensity, is dynamically attracted to the warmest oceans with enhanced convective instability. It is distinct from other cloud clusters at the north that are steered passively by the tropical easterlies and embedded disturbances.

5. Generation of convective instability and intraseasonal oscillation

a. Moist static energy and physical processes associated with CAPE genesis

Given the importance of the atmospheric instability in convection evolution, we investigate how convective instability forms over HE0611-West. First, we examine the vertical structure of moist static energy relating to CAPE. As described in section 2, CAPE relates to a difference between the air parcel’s $\theta_e$ from the source level and the ambient air’s $\theta_v$. Thus, the configuration of $\theta_e$ and $\theta_v$ profiles can be used to illustrate the vertical structure of CAPE. For the air parcel from 1000 hPa, CAPE corresponds to the hatched area between straight and bold lines in Fig. 10.

Figure 10a shows that, on 7 November, the hatched area between the straight and bold lines is still small, which denotes a small CAPE at the beginning of the hot event. Figure 10b shows that the hatched area increases by 3–4 times on 22 November and produces a large CAPE of up to 1800 J kg⁻¹ at the mature stage of the hot event (Fig. 8b). On 26 November, the hatched area shrinks considerably, particularly at middle and lower levels (Fig. 10c), which corresponds to a depletion of CAPE after convection consumption (Fig. 8d). We also
FIG. 8. Spatial pattern of 2-day mean CAPE at 1000 hPa during (a) 19–20, (b) 21–22, (c) 23–24, and (d) 25–26 Nov. Only values above 1000 J kg\(^{-1}\) are plotted. See section 2 for the definition and calculation of CAPE. The open circle denotes the buoy moored at 2°N, 147°E.
FIG. 9. Enlarged views of the CAPE value and cloud cluster over HE0611-West. (a) CAPE above 1000 J kg$^{-1}$ during 23–24 Nov. The rectangle (2°S–4°N, 142°–152°E) denotes the region of the maximum SST, as in Fig. 1. (b)–(d) The life cycle of the migrating cloud cluster over the high-SST region, as indicated by the high-level clouds in gray. The solid line denotes the area with a CAPE value above 1800 J kg$^{-1}$ from (a). The open circle denotes the buoy moored at 2°N, 147°E.
note that the hatched area in Fig. 10b is rather wide at most of the levels, and thus it helps produce strong updrafts (Blanchard 1998). The hatched area is much narrower in Figs. 10a and 10c and is unfavorable for sustaining convective updrafts, because the dilution in the realistic atmosphere acts to reduce the difference between the lifted air parcel and the environment (Emanuel 1994).

Figure 10a also shows that the LFC (LNB), as marked by the bottom (upper) intercept of straight and bold lines, is located at 850 hPa (200 hPa) on 7 November. The LFC becomes low at 900 hPa on 22 November (Fig. 10b). Thus, it becomes easier for the surface air parcel, if disturbed, to reach such a low LFC, initiating new convection. Meanwhile, the LNB becomes high at 150 hPa, which potentially allows for convective updrafts reaching a high level. Therefore, deep convection likely takes place if there is low-level ascent to overcome the small CIN and to release the large CAPE to drive strong updrafts. The demanding large-scale ascent becomes available after 22 November, as the low-level convergence is gradually enhanced by the strong trade winds and westward-moving disturbances (Figs. 7b,c). The favorable conditions of convective instability and low-level ascent thus explain the explosive growth of organized convection over the high-SST ocean of HE0611-West.

Second, we examine the physical processes responsible for CAPE genesis. The time change of CAPE can be induced by the change of either boundary layer $\theta_e$ or midupper-level $\theta^*_e$ (e.g., Emanuel 1994; Chen et al. 2014b). As we know, at a given pressure, the $\theta_e$ value is a function of potential temperature $\theta$ and water vapor $q$, while the $\theta^*_e$ value relates to potential temperature $\theta$. Thus, the evolutions of $\theta_e$, $\theta$, and $q$ profiles allow us to clarify the forcings of the boundary layer and free atmosphere and to quantify the roles of thermal and moist processes in CAPE genesis. Figure 11a shows that the $\theta_e$ value increases steadily at low levels below 850 hPa from 7 to 24 November, while it decreases at middle levels near 600 hPa. The largest increase occurs at 1000 hPa, from \(\theta_e\); 347 K during 7–9 November to more than 355 K during 22–24 November. Such an increasing (decaying) moist static energy in the lower (middle) troposphere leads to an establishment of convective instability, producing the high CAPE value as shown in Figs. 8 and 10b.

Figure 11b shows that the $\theta$ value increases by \(\sim 2 \text{ K}\) at the midupper levels. Such a warming aloft acts as a stabilization effect, and it increases $\theta^*_e$, reducing CAPE. A comparison of Figs. 11a and 11b shows that the increase of low-level $\theta_e$ is about 4 times that of the midupper-level $\theta$. Thus, the destabilization effect by boundary layer warming/moistening is much larger than the free-atmospheric stabilization, producing the large increase of CAPE (Fig. 10b). These results agree with other observational studies (e.g., McBride and Frank 1999; Chen et al. 2014b), suggesting that the CAPE genesis is mainly attributable to the low-level processes that regulate the properties of the boundary layer air mass. The boundary layer warming/moistening that determines the generation of convective instability is also observed during another hot event (Johnson et al. 2001).

Figure 11b shows that the boundary layer $\theta$ increases by \(\sim 2 \text{ K}\) from 7 to 24 November. Thus, the increase of boundary layer $\theta_e$ (\(\sim 8 \text{ K}\) in Fig. 11a) is partly due to the warming effect (\(\sim 2 \text{ K}\)), while the remaining major portion (\(\sim 6 \text{ K}\)) is attributable to the moistening process, as
shown by an increase of low-level water vapor (Fig. 11c). It can be concluded that a large amount of moisture content over the warm oceans is the primary cause of the large CAPE at the mature stage of the hot event. Figure 11c also shows that the moisture increase mainly occurs below 800 hPa, in a robust contrast to a moisture decrease at midupper levels. Such a vertical structure indicates that the moisture supply, evaporated from the high-SST ocean, is mainly capped in the lower troposphere. As a combined result of low-level moistening and midlevel drying, the $\theta_e$ difference between 1000 and 600 hPa increases from 14 to 25 K (Fig. 11a). The vertical gradient of $\theta_e$ clearly suggests that the troposphere becomes convectively unstable with an increasing CAPE, as shown in Figs. 10a and 10b.

b. Moisture regulation related to CAPE genesis

We then proceed to examine the atmospheric circulations, with emphasis on the regulations of moisture content that are crucial for a buildup of convective instability. Using cloud-motion winds, Qin and Kawamura (2010) found that the upper outflows from remote convection over HE0611-East are convergent to HE0611-West. They proposed that the outflows may establish a large-scale subsidence over HE0611-West, and the consequent cloud-free conditions lead to high SST. However, the subsidence cannot be directly observed; its regulation of the moisture profile is yet to be clarified. To validate this hypothesis, we examine the variations of vertical motion derived from JRA-55. Figure 3b shows that the longitude–time variations of rising motion are in a good agreement with those of the observed convective systems in Fig. 3a. In particular, the rising motion associated with the migrating system during 19–25 November is presented. Such a similarity suggests that JRA-55 can capture well the convective activities and their related circulation. It is also shown that the large-scale subsidence is dominant over HE0611-West during 11–22 November (dashed circle in Fig. 3b) and corresponds to cloud-free conditions that are conducive for SST warming (Fig. 3 in Qin and Kawamura 2010).

To further understand the impacts of organized convection, Fig. 12a shows a longitude–vertical section of atmospheric circulation and humidity on 21 November, when the convective system is still located at HE0611-East and is approaching HE0611-West. Large-scale updrafts occur at 172°–183°E and extend through the troposphere as a result of deep convection. The related sinking motion appears in the midupper levels at both the west and east sides of updrafts (i.e., at 155°–165°E). This is because the outflow from deep convection becomes horizontally convergent over adjacent areas (not shown). Figure 12a clearly shows that the sinking motion is accompanied by a low relative humidity of 20%–40% because of a drying effect of the subsidence. Although the sinking motion is strongest at 155°–165°E, the midupper dry layer stretches farther west because of a strong advection by tropical easterlies. Such an extended dry area corresponds to cloud-free conditions and strong solar radiation over HE0611-West, resulting in a rapid oceanic warming (Qin et al. 2008; Qin and Kawamura 2010). A similar pattern of midupper subsidence is seen in previous convective events (Fig. 12b). Meenu et al. (2012) also suggested that large-scale atmospheric circulation in the upwind direction greatly regulates cloudiness over the warm oceans through low-level convergence, midlevel vertical winds, and upper-level divergence. It seems clear that the SST response to radiative forcing is pronounced
to the west of organized convection, where the dry layer of subsidence is extended in the presence of strong easterlies (Figs. 12a). The feature is distinct from that during the MJO with the strongest SST warming to the east, whereas the west region is affected by westerly wind bursts that act to cool sea surface (e.g., Lau and Sui 1997).

Figure 12a also shows a shallow layer of high relative humidity in the lower troposphere to the west of 170°E. In this snapshot near noon, the weak rising motion can be seen at ~137°, ~143°, 153°, and 166°E. These updrafts usually have a low vertical extent below 700 hPa under a capping of midupper subsidence. They appear to induce the localized increase of water vapor, as shown by the bulging contours of relative humidity. It is suggested that the shallow convection and convective mixing play a role in moistening the atmospheric boundary layer. This is likely due to the warm ocean with a large diurnal SST as in HE0611-West that tends to generate shallow cumulus clouds and weak precipitation with a daytime peak (Johnson et al. 2001).

We further examine the impacts of midupper subsidence and boundary layer processes on the temporal variations of water vapor over HE0611-West. Figure 13a shows that the omega and specific humidity at 500 hPa generally have variations with an opposite phase. The subsidence (i.e., a small and positive value of omega) is persistent during 14–22 November and leads to low humidity. The lowest humidity of ~1 g kg⁻¹ is seen during 14–17 and 20–22 November when remote convection occurs near the date line and large-scale subsidence dominates the eastern area of HE0611-West. With an enhancing upward motion (negative omega), the humidity increases considerably on 23 November and reaches a maximum of 3.4 g kg⁻¹ by 24 November.

Figure 13b shows the variations of low-level divergence and humidity averaged over HE0611-West. Large increases of humidity occur during 14–17 and 20–22 November, as marked by the dashed arrows. It is clear that the humidity exhibits a large diurnal variation when its daily mean increases. Because the divergence prevails in these days, the moisture confluence in trade winds is relatively weak. These features suggest that the strong diurnal mixing in the boundary layer plays a crucial role in moistening the lower troposphere. Because the diurnal mixing tends to be strong in fair weather, it is closely linked to the midupper dry layer during these days.
Therefore, the large-scale subsidence induced by remote deep convection provides favorable conditions not only for the formation of the hot event, but also for the accumulation of lower-tropospheric moisture, resulting in the dramatic increase of CAPE over the warm oceans.

Figure 13b shows that the humidity stays at a high value during 23–24 November and has a small diurnal variation, indicating the decayed boundary layer mixing. The high humidity is mainly due to a strong low-level convergence (Figs. 13b and 7c). The feature leads to the variations of moisture profiles in Fig. 11c: the midlevel drying maximizes on 22 November, while the low-level moistening continues until 24 November. This change corresponds to a shift from fair to convective weather.

c. Atmospheric oscillations over the tropical Pacific during HE0611

It has been recognized that equatorial waves control a large fraction of tropical atmospheric variability (e.g., Kiladis et al. 2009). The atmosphere–ocean system may express a response to the multiscale waves crossing the target region during HE0611. Using the analysis technique of Wheeler and Kiladis (1999), we calculate a global space–time spectrum of the tropical rainfall from September 2006 to January 2007. Figure 14 shows that the signals of the $n = 1$ and 2 westward inertio-gravity (WIG), mixed Rossby–gravity (MRG), $n = 1$ equatorial Rossby (ER), $n = 0$ eastward inertio-gravity (EIG), and Kelvin waves, as well as the MJO, are clearly discerned. The power peaks of the westward-propagating WIG, MRG, and ER waves appear at the periods of 2, 5–8, and 20–30 days, respectively. The eastward-propagating signals of EIG, Kelvin, and the MJO have peaks at 2–4, 4–6, and 30–60 days. Thus, we identify the existence of MRG, ER waves, and MJO activity at a time scale roughly comparable to the life-span of the hot event.

To decompose multiscale atmospheric variability and its regional features, we perform bandpass filter analysis for the oscillations at prominent periods. Figure 15 shows the Hovmöller diagram of the filtered 200-hPa velocity potential and 500-hPa vertical motion. Moderate-to-strong MJO activity occurs from late September to mid-October, but it weakens considerably from late October to mid-December (Figs. 15a,d), as noted in the monitoring report of the NOAA Climate Prediction Center (www.cpc.ncep.noaa.gov/products/precip/CWlink/MJO). The phase propagation even turns westward over the western Pacific in the field of 500-hPa omega (Fig. 15d). In the 20–30-day spectrum, the westward-propagating signal is observed over the western Pacific (Figs. 15b,e), which may correspond to the ER wave activity in Fig. 14b. At the shorter spectrum of 5–8 days, the synoptic-scale disturbances tend to propagate westward (eastward) to the west (east) of the date line. Thus, it is suggested that the target region is mainly affected by the westward-propagating oscillations from the central to western Pacific, while the eastward-propagating MJO is less active during HE0611.

Figure 16 gives an enlarged view of the low-frequency and synoptic-scale component over the Pacific during November 2006. It is shown that the 5–8-day disturbance of vertical motion is pronounced near the date line and
corresponds to the frequent convective systems over the
central Pacific. The out-of-phase oscillation occurs at
\(-160^\circ\text{E}\) as a result of the circulation adjustment by
organized convection, as in Fig. 12. Figure 16b further shows
that the 5–8-day oscillation of atmospheric humidity
propagates coherently from 170°W to 140°E, because the
air mass can be drifted westward by ambient winds.

To further illustrate synoptic-scale disturbances, we
examine the daily evolution of high-pass filtered hori-
zontal winds and vertical motion during 14–25 November.
Figure 17 shows a westward propagation of the convectively coupled equatorial waves with off-equatorial updraft to the north of southerly wind. The features are consistent with classic theory for the MRG wave over the equatorial central Pacific (Liebmann and Hendon 1990; Yang et al. 2007a,b; Kiladis et al. 2009). The phase speed of the MRG wave is estimated as 8–10 m s\(^{-1}\) in Fig. 17, which is in the low range of 8–25 m s\(^{-1}\) observed by previous studies. It explains a longer period of 5–8 days in this study than the typical 3–5 days. As a convective core (C in Figs. 17a–f) moves westward to the western Pacific during 14–19 November, a new one emerges at 170\(^\circ\)–160\(^\circ\)W and continues the westward propagation (C\(^{0}\) in Figs. 17e–l). Unlike the preceding wave with decayed convection, the new wave maintains its updraft strength over the western Pacific. The latitude of the convective area gradually shifts from \(\sim 5^\circ\)N on 18 November to \(\sim 10^\circ\)N on 25 November. As the wave turns to the northwest, it develops tropical-depression (TD-type) structures that coincide with an “easterly wave” over the western Pacific, as also shown in Figs. 6a and 14a and in previous studies (Takayabu and Nitta 1993; Kiladis et al. 2009). Figures 16 and 17 show that most of the MRG waves are initiated and grow at 170\(^\circ\)E–160\(^\circ\)W, where strong latent heat from the persistent warm SSTs may support a growth of the convective coupled waves. Although they usually tend to decay over the western Pacific, some of them can be sustained or transformed to easterly waves (the right panel of Fig. 17). It seems clear that such a selective process depends on the favorable preconditions of convective instability and large-scale dynamics over the western Pacific, as discussed in sections 3 and 4.

Figures 16a and 16b show that the 20–30-day oscillation signal relating to the ER wave is initiated at 175\(^\circ\)E–170\(^\circ\)W with the highest SST. It propagates westward and greatly strengthens at 140\(^\circ\)–170\(^\circ\)E, with a phase of suppressed convection and dry conditions during 11–21 November. Such a phase lock between vertical motion and humidity indicates that the drying effect of subsidence leads to a prolonged cloud-free period for increasing SST and atmospheric instability over the western Pacific. During that period, the synoptic-scale disturbances tend to dissipate as they move from the central to western Pacific. The 20–30-day oscillation shifts toward a convective phase during 22–25 November when the westward-moving synoptic-scale disturbance grows over the western Pacific. These regional features in Figs. 15–17 thus highlight that the intraseasonal oscillation is strongly enhanced over the western Pacific as a response to the local air–sea coupling and the activities of MRG and ER waves propagating from the central Pacific. As such, processes may work in other periods, as shown in Fig. 15; further climate statistics on more cases may give us insight into the role of multiscale waves in both the short-term climate variability and the causes of the hot spots.

6. Oceanic response to convective weather and its roles in regional climate

a. Air–sea fluxes and oceanic processes over HE0611-West

In this section, we examine sea surface heat fluxes and upper-oceanic processes to reveal how the ocean
responds to the changing weather conditions over HE0611-West. Figure 18 shows the hourly variation of sea surface variables from buoys moored at the highest-SST area. The SST exhibits large diurnal variation during 15–23 November (Fig. 18a), as in Qin et al. (2008). It resembles the diurnal cycle observed in TOGA-COARE, in which the warming phase has a faster trend and higher amplitude than the cooling phase (Anderson et al. 1996; Lau et al. 1997). The residual warming from the diurnal cycle leads to an increase of...
daily mean SST. As shown in Figs. 18b and 18c, the fair weather with strong solar heating and light winds drives the diurnal cycle, except that a relatively strong wind in the daytime of 20 November may damp the diurnal SST response (Price et al. 1986; Large et al. 1994). Figure 19 shows the air–sea heat flux diagnosed from OAFlux over HE0611-West. During 15–23 November, the shortwave radiation, which is offset by sensible/latent heat fluxes and longwave cooling, leads to a daily net heat flux of 100–130 W m$^2$ in the ocean surface. It is larger than that of 50–70 W m$^2$ during the dry phase of MJO (Weller and Anderson 1996; Lau and Sui 1997), mainly because of relatively large downward shortwave radiation. The net heat flux thus rapidly increases SST and leads to a maturation of the hot event (Qin et al. 2008; Qin and Kawamura 2009b). Such strong solar radiation occurs in a prolonged cloud-free period, which is closely related to the midupper subsidence over HE0611-West induced by deep convection from HE0611-East (Figs. 12–14). The very high SST over HE0611-West thus expresses a response to the forcing of remote convection, as also noted in previous studies (Kawamura et al. 2008; Qin and Kawamura 2010).

Figure 18 shows that a dramatic change of surface variables occurs on 24 November when convective system arrives at HE0611-West. The SST begins to decrease, and its diurnal cycle almost vanishes, indicating a collapse of previous diurnal heating pattern (Fig. 18a). This is accompanied by suppressed shortwave radiation (Fig. 18b) because of a cloud-shielding effect caused by the increased cloudiness of the convective system (Fig. 5). The convective activities also lead to an increase of surface winds and relative humidity, as well as a temperature drop (Figs. 18c–e). An analysis of air–sea fluxes shows that both latent and sensible heat fluxes strengthen significantly during 24–25 November (Figs. 19a,b), suggesting a strong transfer of oceanic energy into the atmosphere. The strong latent heat loss and decayed incoming shortwave radiation contribute to a negative heat gain in the ocean surface (Fig. 19e), which is analogous to that observed during a wind burst (Weller and Anderson 1996). Such a decline of net heat flux acts to reduce the SST and dissipate the hot event (Maloney and Sobel 2007). It seems that shortwave radiation and evaporative flux variability are crucial to regulating the temperature
of the warm pool during the hot event (Fasullo and Webster 1999; Li et al. 2000; Saji et al. 2006).

Besides the air–sea fluxes, we further examine the oceanic processes using the thermal profiles derived by buoy measurements. Figure 20a shows that the diurnal variation of water temperature occurs at the ocean surface before 24 November. The diurnally mixed layer is quite shallow, as solar heating tends to warm and stabilize the upper ocean (Price et al. 1986; Large et al. 1994). The ocean warm layer disappears since 24 November (Fig. 20a), and there is a small delay between the maximum wind and the minimum SST (Figs. 18a,c). This suggests a rapid surface cooling due to the vanishing of the ocean warm layer as soon as surface winds increase (Duvel and Vialard 2007). Although the seawater temperature near the surface decreases largely by 26 November, the temperature at 25-m depth increases slightly (Fig. 20b). This change corresponds to an effect of mechanical mixing by which the strong surface winds act to deepen the mixed layer (Large et al. 1994; Korty et al. 2008). In an idealized experiment, Sobel and Gildor (2003) also noted that the maximum response to hot spot evolution in the ocean mixed layer occurs at 10–20 m deep. As a contrast to the surface layer, Fig. 20a shows that seawater temperature below 50 m exhibits a semidiurnal variation through the whole period of interest, probably because of oceanic internal waves and/or tides. It seems to have minimal impact on the diurnal cycle of seawater temperature near the surface.

b. Discussion on the impacts of SST regulation on regional climate

It is well known that the tropical SST behaves like a self-regulating system: high-SST tends to destabilize the atmosphere and triggers convection; convection in turn removes high SST and restores atmospheric stability; high SST then rebuilds in the following fair weather, and a new cycle starts (e.g., Stephens et al. 2004). Since remote organized convection helps sustain an extended warming period of the upper ocean, the SST regulation is pronounced on intraseasonal time scales (Waliser 1996; Lau et al. 1997; Johnson et al. 2001; Sobel and Gildor 2003). In this study, we see that the organized convection works effectively at all stages of the SST regulation. Remote convection strengthens both hot event formation and convective instability; it takes 2–3 weeks to establish high SST by a large warming of the upper ocean under the prolonged cloud-free and light-wind conditions (Figs. 16, 18–19). Then it acts as the initial disturbance for local convective growth that cools the ocean surface, in which it takes days of convection migration to remove the shallow layer of warm water (section 6a). Thus, the regulating cycle (20–30 days) of SST and deep convection can be shorter than that of MJO.

Fig. 20. Vertical profiles of the seawater temperature observed at the TAO/TRITON buoy site (2°N, 147°E).
(a) Temporal variations during 15–27 Nov, with contour for ≥29°C at an interval of 0.5°C. (b) Daily mean on 7, 22, and 26 Nov. The location of the buoy is shown in Fig. 1.
with a period of 30–60 days reported by previous studies. A key tuning of the oscillation phase is the convective system that propagates and grows over the western Pacific during 22–25 November with a precondition of ocean–atmosphere energy (Figs. 2, 8–10, 16). It is noted that the evolution of organized deep convection, as in a hurricane, is mainly governed by the initial convection intensity, the thermodynamic state of the atmosphere, and the heat content in the upper-ocean layer (Emanuel 1999; Shay et al. 2000). We see that the latter two factors depend strongly on the phase of the hot event over the western Pacific.

In this hot event of interest, the westward response of local SST and convection to remote forcing differs from that in the eastward-propagating MJO (e.g., Lau et al. 1997). During El Niño, the anomalously active convection over the central Pacific seems to trigger the westward-moving ER and MRG waves (Figs. 14–17), which behave like a downscaling process from the seasonal El Niño event. In particular, the associated convective systems serve as high-frequency disturbances that migrate from HE0611-East to -West in the tropical easterlies (Figs. 2d, 16–17). While most systems dissipate over HE0611-West, some of them, such as that during 22–25 November, are selectively amplified by the feedback of the hot event on convection growth through a precondition of atmospheric instability (sections 3–5), which greatly enhances low-frequency variability of rainfall and SST (Figs. 2a,c). Such oscillations are repeated four times from September 2006 to January 2007 (Fig. 15); thus, the westward SST–convection responses revealed here may be typical. Since convective systems mainly regulate adjacent circulation (Figs. 12–13), they may have more impacts on the eastern part of HE0611-West than on the western part. As shown in Figs. 3 and 16, all of the convective systems are dissipating at the west of 140°E. As a result, the rainfall anomaly of November 2006 is +30% near the date line; it reduces to −12% over 140°–160°E and to −46% over 120°–140°E. Such a zonal gradient of the rainfall anomaly highlights the importance of the air–sea interactions in modulating regional climate.

Previous studies noted that over the thin mixed layer region, the intraseasonal SST response to convective perturbation is pronounced and recurrent (Weller and Anderson 1996; Bernie et al. 2005; Duvel and Vialard 2007; Li et al. 2013). In this study, we also see that air–sea fluxes and ocean processes over HE0611-West are significantly influenced by the changing weather of deep convection (from remote to local). The upper ocean with a thin mixed layer can be switched between heat gain and heat loss, resulting in a large fluctuation of SST. The hot event, while forming as a response to remote convection, comes to an end itself through attracting the migrating convection and promoting local convection. The present results of study are supportive of the previous hypotheses that treated the very high-SST phenomena as an unstable state (Waliser 1996; Sobel and Gildor 2003). Because the upper ocean can strongly interact with atmospheric convection over HE0611-West, it leads to an extended area of very high SST and a long-lived convective system that develops beyond the El Niño warm ocean (i.e., HE0611-East), suggesting an amplified short-term regional climate.

7. Concluding remarks

The large-scale phenomena of very high SSTs are typical fluctuations of SST on time scales of weeks to months over the warm pool. In previous theories, these hot spots/events are recognized as part of the dynamic and thermodynamic adjustment to atmospheric variability associated with the eastward-propagating MJO. This study presents a new kind of hot event evolution with a strong response to and feedback onto the westward-propagating organized convection and equatorial waves that are induced by El Niño while an active MJO is absent from the Pacific Ocean. The findings are summarized as follows:

1) Organized deep convection frequently develops over the central Pacific, where the El Niño positive SST anomalies prevail during November 2006. These remote convective systems help to sustain cloud-free conditions over the western Pacific, as their associated subsidence can extensively dry the midupper troposphere. Strong solar radiation in fair weather and light-wind conditions leads to a net heat flux that increases the heat content of a thin ocean mixed layer. It leads to the formation of a hot event in the western Pacific and corresponds to a charge of oceanic energy.

2) The atmospheric subsidence and large diurnal SST also enhance shallow convective mixing while capping moisture in the lower troposphere. The accumulated water vapor can overcome the stabilization effect of subsidence warming; it plays a vital role in generating a large CAPE and a decayed CIN. Convective instability thus becomes evident over the warm ocean, indicating a charge of atmospheric moist energy.

3) Given the enhanced convective instability and large-scale rising motion, the hot event at the mature stage turns to support a long-lived convective system that can propagate coherently from the central to western Pacific. Some cloud clusters are clearly displaced from their parent system when they are dynamically
attracted to the warmest oceans with maximum CAPE aloft. The local convection is shutting off strong shortwave radiation while enhancing upward air–sea fluxes and downward mechanical mixing. It shifts the ocean mixed layer from heat gain to heat loss, which removes the high SST and dissipates the hot event (i.e., a discharge of system energy).

4) With a charge phase of 2–3 weeks and a discharge phase of several days, the ocean–atmosphere system experiences pronounced oscillation at a period of 3–4 weeks, which is shorter than the time scale of hot spots associated with the MJO. It is shown that, in the regime of tropical easterlies, the westward-propagating organized convection is effective to trigger a westward SST response, in contrast to the MJO in which the SST warming mainly occurs to the east of organized convection. The relevant 5–8-day disturbances that resemble the westward-moving mixed Rossby–gravity waves are most evident over the central Pacific. Although most of these synoptic-scale waves tend to decay over the western Pacific, some of them can be sustained by the feedback of the hot event on convection growth with enhanced atmospheric instability. Such a selective process acts to enhance the westward-propagating equatorial Rossby wave and engenders a pronounced intraseasonal oscillation of rainfall. Based on our results, we conclude that the El Niño warm SST anomaly in the central Pacific can force distinct short-term SST and convection responses over the western Pacific.

The results of this study emphasize strong air–sea interactions during a hot event, which may have important implications regarding the mechanism of climate variability over the warm pool. It is known that the anomalous Walker circulation during El Niño, as a climate mean state, tends to establish fair weather over the western Pacific. In this study, we found that it does not necessarily suppress convection through the period. Intermittently, organized deep convection can develop over the western Pacific because of the response and feedback of the upper-ocean hot events. Such a strong coupling between SST fluctuations and intraseasonal oscillations also suggests an inherent predictability of the climate system (Lau and Sui 1997; Woolnough et al. 2000; Johnson et al. 2001; Duvel and Vialard 2007). In particular, the pronounced 20–30-day oscillations associated with a hot event may bridge the gap between weather forecasts and monthly climate prediction. It has been shown that the coupled forecasts, in which the SST is determined by the interactive air–sea coupling processes, can increase significantly the predictability of intraseasonal oscillations and seasonal rainfall anomalies compared to the uncoupled forecasts (Fu et al. 2008; Zhu and Shukla 2013; Shukla and Zhu 2014). In view of the fact that the hot event involves air–sea coupling behaviors on a broad range of time scales, it requires a proper description in the climate models (Sui et al. 1997; Bernie et al. 2005; Li et al. 2013). Thus, the hot event evolution and its climate impacts provide a benchmark for evaluating model performance on multiscale air–sea interactions.

In this study, the short-term hot event in the western Pacific is strongly related to El Niño. One may expect that such air–sea coupling processes occur in other years as there are warm SSTs lasting for months in the central Pacific. Similar short-term variations of SSTs in the western Pacific took place in the recent winters of 2002 and 2009 (not shown). Instead, La Niña conditions appear to inhibit the formation of hot spots in the warm pool (Waliser 1996). Since the occurrences of hot events are quite common in the Indo-Pacific warm pool (Qin et al. 2007), further observational analyses are needed to clarify whether the present findings are generalized in a variety of ocean–atmosphere conditions. More efforts to explore the recent remote sensing data for monitoring oceanic conditions, atmospheric energy, and convective activities are also warranted to provide additional insights into the climate dynamics.

Acknowledgments. The authors are grateful to Dr. H. Kawamura, Dr. W. Wang, and three anonymous reviewers for their helpful comments. They thank JMA for providing the Merged Satellite and In Situ Data Global Daily Sea Surface Temperatures (MGDSST), cloud-motion winds, and global reanalysis JRA-55. Thanks also go to the GSFC and TSDIS of NASA for providing TRMM rainfall data, the NASA Jet Propulsion Laboratory for providing the satellite infrared sounding product AIRS, the Woods Hole Oceanographic Institution for providing the air–sea fluxes data, and the NOAA Pacific Marine Environmental Laboratory for providing the TAO/TRITON buoy data. This study was supported by the National Natural Science Foundation of China (Grant 41306015 and 41575068) and the Strategic Priority Research Program of the Chinese Academy of Sciences (Grant XDA10010304).

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