Evolution of the Madden–Julian Oscillation in Two Types of El Niño

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ABSTRACT

Evolution characteristics of the Madden–Julian oscillation (MJO) during the eastern Pacific (EP) and central Pacific (CP) types of El Niño have been investigated. MJO activities are strengthened over the western Pacific during the predeveloping and developing phases of EP El Niño, but suppressed during the mature and decaying phases. In contrast, MJO activities do not show a clear relationship with CP El Niño before their occurrence over the western Pacific, but they increase over the central Pacific during the mature and decaying phases of CP El Niño. Lag correlation analyses further confirm that MJO activities over the western Pacific in boreal spring and early summer are closely related to EP El Niño up to 2–11 months later, but not for CP El Niño. EP El Niño tends to weaken the MJO and lead to a much shorter range of its eastward propagation. Anomalous descending motions over the Maritime Continent and western Pacific related to El Niño can suppress convection and moisture flux convergence there and weaken MJO activities over these regions during the mature phase of both types of El Niño. MJO activities over the western Pacific are much weaker in EP El Niño due to the stronger anomalous descending motions. Furthermore, the MJO propagates more continuously and farther eastward during CP El Niño because of robust moisture convergence over the central Pacific, which provides adequate moisture for the development of MJO convection.

1. Introduction

The Madden–Julian oscillation (MJO; Madden and Julian 1971, 1972), a dominant component of the tropical atmosphere on the intraseasonal time scale, has important influences on the weather and climate around the world (Zhang 2005, 2013; Li et al. 2014). El Niño was first discovered to be characterized as an anomalous rising of the sea surface temperature (SST) over the equatorial eastern Pacific caused by large-scale atmosphere–ocean interaction (Bjerknes 1966; Rasmusson and Carpenter 1982). It is one of the pivotal factors for the interannual variation of the climate system. It has evident and wide influences on weather and climate globally (Zhou et al. 2010; Yuan and Yang 2012; Yuan et al. 2012). The relationship between El Niño and weather and climate has been a hot topic since the 1980s.

Lau and Chen (1986) first showed that the MJO could excite the El Niño event through energy transfer. Many following studies suggested that the enhanced MJO activities over the western Pacific in boreal spring and early summer favor the occurrence and development of El Niño events as a stochastic forcing (Hendon et al. 2007; Marshall et al. 2009). Joint effects of stronger MJO activities and anomalous oceanic Kelvin waves over the equatorial central-western Pacific could lead to the occurrence of El Niño (Li and Liao 1998; McPhaden et al. 2006). The MJO influences ENSO mainly through the interannual variability of its intensity (Li and Liao 1998; Li et al. 2003; McPhaden et al. 2006; Zavala-Garay et al. 2005). The MJO energy is a small fraction of the total

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energy in the atmosphere, but it might have a pivotal effect on promoting the development of El Niño (Zavala-Garay et al. 2005; Kapur et al. 2012). Some numerical model results show that the MJO signal, especially its amplitude, might closely relate to model prediction skills of El Niño (McPhaden 1999; Kapur and Zhang 2012). For example, the simulated El Niño of 1997/98 became stronger by 50% when MJO forcing was included (Kessler and Kleeman 2000). However, other studies did not find a clear simultaneous relationship between the MJO and El Niño (Hendon et al. 1999; Slingo et al. 1999). Indeed, anomalous MJO activities over the western Pacific lead El Niño events by several months (Li and Liao 1998; Zhang and Gottschalck 2002; Li et al. 2003; Tang and Yu 2008). Hendon et al. (2007) indicated that the MJO over the western Pacific in boreal spring and early summer is closely related to El Niño. The occurrence of El Niño may not be a result of any individual MJO event, but rather the cumulative effect of a sequence of several MJO events (Zhang and Gottschalck 2002; Zavala-Garay et al. 2005). Furthermore, the relationship between the MJO and El Niño has a decadal variation (Zhang and Gottschalck 2002; Tang and Yu 2008). So the relationship between them may have seasonal, regional, and decadal dependence.

The variability of the MJO also can be influenced by El Niño (Li and Smith 1995; Roundy and Kravitz 2009; Guschchina and Dewitte 2012; Kapur and Zhang 2012; Chen et al. 2015). The horizontal distribution of SST may be important to certain aspects of the MJO, such as its initiation, intensity, propagation, and prediction (Pegion and Kirtman 2008a,b; Ray et al. 2009; Ray and Zhang 2010; Kim et al. 2010; Wang et al. 2015). The eastward propagation of MJO is enhanced during an El Niño developing summer, while it is weakened during the decaying summer (Lin and Li 2008). In an El Niño winter, the MJO over the Pacific (especially over the western Pacific) is weakened, discontinuous in its eastward propagation, and inclines to a “barotropic” vertical structure (Li and Smith 1995; Chen et al. 2015). Certain changes have been found in MJO characteristics during El Niño based on observations and numerical simulations. They include enhanced (weakened) low-level zonal winds over the central (western) Pacific, slower eastward propagation, and an eastward extension of MJO convection (McPhaden 1999; Tam and Lau 2005). The feedback effect of SST to the MJO in a coupled model is also very important in the relationship between the MJO and ENSO (Kapur and Zhang 2012). The interaction between the MJO and El Niño and their relationship needs further exploration.

Recently, a new type of El Niño, in which warm SST anomalies are in the equatorial central Pacific near the date line, has drawn more research attention (Ashok et al. 2007; Kao and Yu 2009, Kug et al. 2009; Ren and Jin 2011). In many recent studies, El Niño events are classified into two types: eastern Pacific (EP) El Niño and central Pacific (CP) El Niño. These two types of El Niño have different impacts, such as tropical cyclones over the Atlantic and Pacific (Kim et al. 2009; Chen and Tam 2010) and precipitation over East Asia, the United States, and Australia (Wang and Hendon 2007; Weng et al. 2007; Yuan and Yang 2012; Yuan et al. 2012). Their impacts are also different in their different phases (Yuan and Yang 2012; Yuan et al. 2012).

However, most previous studies on the relationship between the MJO and El Niño did not distinguish these two types of El Niño events. Recent studies that do consider the different relationships between the MJO and the two types of El Niño have yielded controversial results. Feng et al. (2015) showed that the MJO could promote the development of CP El Niño but not EP El Niño. The activities of the MJO are enhanced during the predeveloping, mature, and decaying phases of CP El Niño, whereas there are no significantly intensified MJO activities prior to EP El Niño. However, Guschchina and Dewitte (2012) showed that enhanced MJO activities appeared in boreal spring and summer prior to EP El Niño, but also appeared in the mature and decaying phases of CP El Niño. They suggested that the MJO might contribute to the triggering of EP El Niño, and to the persistence of positive SST anomalies of CP El Niño. Yuan et al. (2015) showed that enhanced MJO activities over the western Pacific and its eastward propagation are important to the onset for both types of El Niño. Therefore, the relationship between MJO and these two types of El Niño events needs further investigation.

This study documents the evolution features of the MJO during the life cycle of these two types of El Niño and proposes possible mechanisms. Data and methods are described in section 2. Characteristics of SST and OLR related to these two types of El Niño are briefly shown in section 3. Different relationships between the MJO and two types of El Niño are discussed in section 4, and their possible reasons are explored in section 5. A summary and discussion are given in section 6.

2. Data and methods

Daily mean atmospheric data of wind and specific humidity are from the National Centers for Environmental Prediction (NCEP)–National Center for Atmospheric Research (NCAR) reanalysis (Kalnay et al. 1996), with a horizontal resolution of 2.5° × 2.5°. Daily mean outgoing longwave radiation (OLR) data with a 2.5° × 2.5° horizontal resolution are from the National
Oceanic and Atmospheric Administration (NOAA) (Liebmann and Smith 1996). Monthly SST data are from the Hadley Centre Sea Ice and Sea Surface Temperature datasets (HadISST1) of the Met Office (Rayner et al. 2003) with a horizontal resolution of $1^\circ \times 1^\circ$. The quality and precision of the reanalysis data improved dramatically after the 1970s due to the inclusion of the satellite observations into the assimilation (Liu et al. 2014). Therefore, to be consistent with the period of OLR, all results presented in this study are based on the period from 1 January 1975 to 31 December 2011.

An anomalous time series of a given variable was generated by removing its annual climatology and its long-time linear trend. The MJO signal was obtained using a 30–90-day Lanczos bandpass filter with 200 days of smoothing (Duchon 1979; Cai et al. 2013). Filtered signals thus obtained may contain other intraseasonal signals, but the MJO should be the dominant one. The method proposed by the Wheeler and Kiladis (1999) using the eastward spectral to reconstruct the MJO signal may overestimate the MJO strength (Ling et al. 2013), because all eastward spectral power does not belong to eastward-propagating perturbations; only that which is incoherent with its westward counterparts does (Hayashi 1982).

The El Niño index used in this study for EP and CP El Niño follows the definition of Ren and Jin (2011):

\[
\begin{align*}
N_{EP} &= N_3 - \alpha N_4 & \text{and} & \quad \alpha = \begin{cases} 2/5 & N_3 N_4 > 0 \\ 0 & \text{otherwise} \end{cases}, \\
N_{CP} &= N_4 - \alpha N_3
\end{align*}
\]

(1)

where $N_{EP}$ and $N_{CP}$ represent the EP and CP El Niño index, respectively; $N_3$ and $N_4$ denote the Niño-3 ($5^\circ$S–$5^\circ$N, 150°W–90°E) and Niño-4 ($5^\circ$S–$5^\circ$N, 160°E–150°W) index, respectively. These two indices have less simultaneous correlation, and they can clearly capture the main features of these two types of El Niño (Ren and Jin 2011).

A 3-month running mean was first applied to the indices. If the amplitudes of both types of El Niño indices averaged in boreal winter (November–March) are less than a half standard deviation, a normal year was then identified. There are seven normal years (1980/81, 1981/82, 1989/90, 1992/93, 1993/94, 2001/02, and 2003/04) from 1975 to 2011. If the values of either index exceed one standard deviation for five consecutive months including at least one month in boreal winter, then a mature phase of an El Niño event is identified. According to this definition, three EP El Niño events (1976/77, 1982/83, and 1997/98) and six CP El Niño events (1977/78, 1990/91, 1994/95, 2002/03, 2004/05, and 2009/10) are identified.

The methods used in this paper include composite analyses, partial correlation and multivariate linear regression. Because $N_{EP}$ and $N_{CP}$ have certain simultaneous correlation, the partial correlation was used to calculate the correlation between one of them and a third party time series in order to minimize the influence of such simultaneous correlation. For example, the partial correlation between the time series of $N_{EP}$ and MJO OLR is calculated, based on Anderson (2003, 136–144), as

\[
r_{EP,OLR} = \frac{r_{EP,OLR} - r_{EP,CP} r_{CP,OLR}}{\sqrt{(1 - r_{EP,CP}^2)(1 - r_{CP,OLR}^2)}},
\]

where $r_{EP,OLR}$ represents the partial correlation coefficient between $N_{EP}$ and MJO OLR with minimized the influence of the time series of $N_{CP}$, and $r$ represents the correlation coefficient between two time series indicated in the subscript. The $t$ test was used as the significance test. Because of one more variable used in the calculation, the degree of freedom should be $n - 3$, where $n$ is the total number of time series.

Multivariate linear regression, also known as partial regression, was used in this study to reduce the impacts of simultaneous correlation between $N_{EP}$ and $N_{CP}$. The regression coefficient represents the contribution of one independent variable to the dependent variable and minimizes the influence of other variables. For example, the multivariate regression of the indices of the two types of El Niño to MJO OLR is calculated, also based on Anderson (2003), as

\[
\hat{\text{OLR}} = a_0 + b_{EP} N_{EP} + b_{CP} N_{CP},
\]

where $\text{OLR}$ represents the regressed time series of MJO OLR, and $b_{EP}$ ($b_{CP}$) represents the directly contribution of EP (CP) El Niño to the regressed MJO OLR while partly eliminating the influence of CP (EP) El Niño. They are calculated as follows:

\[
\begin{align*}
b_{EP} &= \frac{S_{EP} S^2_{EP,OLR} - S_{EP,CP} S_{CP,OLR}}{S_{EP}^2 S_{CP}^2 - S_{EP,CP}^2}, \\
b_{CP} &= \frac{S_{CP} S^2_{CP,OLR} - S_{CP,EP} S_{EP,OLR}}{S_{CP}^2 S_{EP}^2 - S_{CP,EP}^2}.
\end{align*}
\]

(4)

where $S$ represents the variance of the variable or the covariance between the two variables indicated in the subscript. The $F$ test was used for statistical significance test of $b_{EP}$ and $b_{CP}$.

3. Characteristics of SST and OLR in the two types of El Niño

Monthly mean SST anomalies averaged over $10^\circ$S–$10^\circ$N partially regressed onto the two types of El Niño
indices (averaged over December–February) clearly show their differences in the distribution and intensity of SST anomalies (Fig. 1, color). Positive SST anomalies in EP El Niño are mainly to the east of date line (Fig. 1a), in contrast, they are concentrated between 160°E and 130°W with the maximum center located in central Pacific near the date line in CP El Niño (Fig. 1b). The intensity of SST anomalies is also stronger in EP El Niño. Meanwhile, the intensity and range of negative SST anomalies over the western Pacific are also more prominent, and positive SST anomalies appear over the eastern Indian Ocean after the occurrences of EP El Niño (Fig. 1a).

SST anomalies can directly affect the activities of atmosphere through its thermal effects, such as the ascending motion and precipitation (Wang and Sobel 2011). OLR is well in describing the convective activities in tropics. Monthly mean OLR anomalies averaged over 10°S–10°N partially regressed onto the two types of El Niño indices (averaged from December to February. Dashed contours are for negative values, and zero contours are omitted. Results passing the significant test at 90% confidence level are presented for anomalous SST and stippled for anomalous OLR; 0 (1) represents El Niño onset (next) year.

Fig. 1. Evolution of longitudinal distribution of monthly mean anomalous SST (colors, K) and OLR (contours; interval = 2 W m⁻²) averaged over 10°S–10°N partially regressed onto (a) EP and (b) CP El Niño indices averaged from December to February. Dashed contours are for negative values, and zero contours are omitted. Results passing the significant test at 90% confidence level are presented for anomalous SST and stippled for anomalous OLR; 0 (1) represents El Niño onset (next) year.

The anomalous activities of convection reflect the impacts of El Niño events on the atmosphere and further illustrate that these impacts are different for these two types of El Niño. Previous studies have shown the important role of feedback from convective latent heating in the dynamics of the MJO (Lau and Peng 1987; Li et al. 2002). The significant difference of convective activities in these two types of El Niño could lead to different activities of the MJO.

4. Relationships between the MJO and the two types of El Niño

Figure 2 shows composite of longitudinal evolution of the MJO in terms of the monthly mean amplitude and anomalies of zonal wind at 850 hPa and OLR averaged
over 10°S–10°N in two types of El Niño events. The monthly mean amplitude is calculated by taking the square root of the variance of MJO daily zonal wind and OLR within a 3-month moving window following Hendon et al. (2007).

The amplitude of MJO zonal wind at 850 hPa is enhanced over the equatorial western Pacific during the predeveloping and developing stages of EP El Niño, but it is weakened over the eastern Indian Ocean and western Pacific during the mature phase (Fig. 2a). Such enhanced MJO amplitude over the western Pacific occurs prior to CP El Niño (Fig. 2b), but it also appears at the mature and decaying phases without eastward movement. The composite results of the MJO amplitude coincide with those of MJO kinetic energy in Yuan et al. (2015). Furthermore, the occurrence of the strengthened MJO amplitude over the western Pacific is earlier and farther west in EP than in CP El Niño. The composite results in the amplitude of MJO OLR (Figs. 2c,d) show almost the same characteristics. The enhanced MJO amplitude is mainly over the western Pacific, and the weakened MJO amplitude is near the date line in boreal winter before the occurrence of EP El Niño (Figs. 2a,c). After EP El Niño onset, the MJO over the eastern Pacific is strengthened significantly. But it does not occur in CP El Niño, which shows a different relationship between the MJO and SST during boreal winter. Some differences in Fig. 2 do not pass the
significance test, which may be due to the limited sample size of El Niño events. These differences in the MJO amplitude in the two types of El Niño suggest their different impacts on the MJO, especially over the equatorial Indian Ocean and western Pacific.

The different features of the MJO in the two types of El Niño raise a question as whether these differences are physically related to El Niño. Figure 3 shows the monthly mean amplitude of MJO zonal wind at 850 hPa and OLR averaged over 10°S–10°N partially regressed onto the two types of El Niño indices averaged from December to February. Results passing the significant test at 90% confidence level are stippled; 0 (1) represent El Niño onset (next) year.

![Figure 3](http://journals.ametsoc.org/doi/abs/10.1175/JCLI-D-15-0486.1)

Fig. 3. Evolution of longitudinal distribution of monthly mean amplitude of MJO (a),(b) zonal wind at 850 hPa (m s⁻¹) and (c),(d) OLR (W m⁻²) averaged over 10°S–10°N partially regressed onto (left) EP and (right) CP El Niño indices averaged from December to February. Results passing the significant test at 90% confidence level are stippled; 0 (1) represent El Niño onset (next) year.

winter before the onset of EP El Niño, and then the center of the enhanced MJO gradually moves eastward along with the development of EP El Niño. Notable suppressed MJO activities appear over the eastern Indian Ocean after the onset of EP El Niño and expand eastward gradually into the western Pacific (Fig. 3a). However, there are no significantly enhanced MJO activities before the onset of CP El Niño. The MJO amplitude over the eastern Indian Ocean and Maritime Continent does not show significant changes after the onset of CP El Niño, but it is significantly increased over the central Pacific (Fig. 3b). The enhanced MJO activities over the western Pacific may play an important role in the onset of EP El Niño, but they do not occur before the onset of CP El Niño. Furthermore, the MJO
amplitude over the central Pacific is increased during the mature phase of CP El Niño, but it is decreased over the western Pacific in EP El Niño. These suggest different roles of the MJO in the initiation of the two types of El Niño and different impacts of the two types of El Niño on the MJO.

The evolutions of the monthly mean amplitude of MJO OLR averaged over 10°S–10°N partially regressed onto the two types of El Niño indices (averaged over December–February) are shown in Figs. 3c and 3d. Different characteristics of MJO OLR under the two types of El Niño are similar to the MJO zonal wind at 850 hPa. The amplitude of the MJO OLR is enhanced over the western Pacific before the onset of EP El Niño, although not as noticeably as MJO zonal wind at 850 hPa. However, the amplitude of MJO OLR is increased significantly over the eastern Pacific after the onset of EP El Niño (Fig. 3c), which may relate to the heating effect of the increased SST in EP El Niño there. The MJO OLR over the western and central Pacific is strengthened during the mature and decaying phases of CP El Niño. The MJO OLR between 130°E and 170°W is also strengthened during the developing phases of CP El Niño, as is the MJO zonal wind at 850 hPa. This will be discussed later.

Only the MJO activities during the El Niño events were taken into consideration in Fig. 2, whereas all MJO activities during both cold and warm ENSO phases were used in the regression in Fig. 3, which may contain information other than the relationship of the MJO and El Niño, such as the relationship between the MJO and La Niña. However, both results agree well with each other and clearly suggest the close relationships between MJO activities over the western Pacific in boreal spring and summer and the development of EP El Niño in the following winter. Such a relationship is absent for CP El Niño. This confirms the seasonal dependence of the relationship between the MJO and El Niño (Hendon et al. 2007) and also demonstrates that this seasonal dependence is not the same for the two types of El Niño.

To further analyze the sensitivity of El Niño to the preceding MJO activities over the western Pacific, their lag correlation was calculated. MJO activities over the western Pacific are represented by the amplitude of its zonal wind at 850 hPa averaged over 10°S–10°N, 120°E–180°. The lagged partial correlation between the El Niño index and MJO amplitude over the western Pacific as a function of starting month is shown in Fig. 4 for both types of El Niño. MJO activities over the western Pacific show evidently simultaneous correlation with the EP El Niño index from July through October (Fig. 4a). There is a significant lagged correlation between the MJO amplitude over the western Pacific in boreal spring and summer and the EP El Niño index up to 2–11 months later. However, a strong simultaneous correlation between the MJO amplitude over the western Pacific and CP El Niño index only occurs from January to June (Fig. 4b). The MJO over the western Pacific in boreal spring and summer does not show significant lagged correlation with the CP El Niño index in the following fall or winter.

Hendon et al. (2007) pointed out that this significant relation of MJO activities over western Pacific in boreal spring and summer with the EP El Niño may partly stem from significant self-correlation of SST. To exclude this possibility, the MJO amplitude linear regressed onto the El Niño indices at zero lag was first removed before calculating the lagged correlation. Results (Fig. 4c) clearly show that the correlation of the MJO over the western Pacific in boreal spring and early summer with the EP El Niño index up to 2–11 months later is still significant, which indicates that the strengthened MJO activities over the western Pacific in boreal spring and early summer may play an important role in the development of EP El Niño in the following autumn and winter. No such significant lag correlation appears in CP El Niño (Fig. 4d). It further demonstrates that MJO activities over the western Pacific may have no noticeable interaction with the development of CP El Niño. The results of MJO OLR have the similar characteristics (not shown). The summer MJO has a significant negative relationship with the EP El Niño index, leading it by 2–12 months (Figs. 3a,c), indicating that EP El Niño may suppress MJO activities in its decaying summer (Lin and Li 2008).

As indicated in Figs. 3b and 3d, there are strong MJO activities over the central Pacific before the occurrence of CP El Niño. The lag correlation relationship between MJO activities over the central Pacific and CP El Niño has been evaluated. The results (not shown) suggest that such strong MJO activities over the central Pacific before the onset of CP El Niño mainly stem from significant self-correlation of SST.

The MJO not only promotes the development of CP El Niño but also is affected by the El Niño (Li and Smith 1995; Tam and Lau 2005; Chen et al. 2015). Different distributions of anomalous SST in the tropics between the two types of El Niño will lead to different distributions of moisture and the large-scale circulation, and furthermore will lead to different impacts on the characteristic of the MJO, such as its intensity and propagation.

The characteristics of convective signals of MJO propagation over the Indian Ocean and Pacific are also examined. The propagation of MJO OLR that regressed onto its time series averaged over 10°S–10°N,
120°–150°E during boreal winter (November–March) of the two types of El Niño and normal years is shown in the left column of Fig. 5. The eastward propagation of the MJO, one of its most prominent features, is very different between these two types of El Niño (Figs. 5a,c). The intensity of the MJO is weakened during both types of EP El Niño over the western Pacific around 130°E compared to normal years because of the weakened upward branch of the Walker circulation. The eastward propagation is more continuous, its speed is little slower, and the propagation range is longer in CP El Niño than EP El Niño. The MJO intensity over the central Pacific is also stronger in CP El Niño than in normal years and its propagation is more continuous (Figs. 5c,e). The propagation of the MJO was also evaluated using the zonal wind at 850 hPa (Figs. 5b,d,f) and the MJO index averaged over different longitude ranges (150°E–180°, 150°–160°E, and 160°E–180°). The results are almost the same.

5. Causes for the different impacts of the two types of El Niño on the MJO

To explain the different characteristics of MJO propagation and intensity under the two types of El Niño, composites of the anomalous large-scale background circulation during boreal winter of the two types of El Niño are investigated. The composite anomalous zonal-vertical circulation (Fig. 6) shows that anomalous convergence (divergence) occurs over the equatorial eastern (western) Pacific in EP El Niño. Anomalous westerlies are over the equatorial Pacific from 150°E to 90°W. The Indian Ocean is dominated by easterly anomalies (Fig. 6a). Anomalous ascending (descending) motions occur over the eastern (western) Pacific in EP El Niño. For CP El Niño, a similar anomalous circulation exists but is weaker, and the anomalous convergence and ascending motion are mainly over the central Pacific (Fig. 6b).
western Pacific becomes dry in both types of El Niño, and the eastern Pacific becomes moist in EP El Niño while the central Pacific becomes moist in CP El Niño. The anomalies of moisture are related to the anomalous circulation and SST (Fig. 6, blue line) of the two types of El Niño. The anomalous circulation and moisture lead to insufficient moisture flux convergence in the lower troposphere over the western Pacific, especially in EP El Niño (Fig. 6, red line), which is a reason for the weakened intensity of the MJO over the Maritime Continent and western Pacific during El Niño (Fig. 5). The anomalous large-scale convergence as well as the positive specific humidity anomaly in the middle and lower
troposphere over the eastern (central) Pacific (Fig. 6) are mainly due to the underneath high SST of EP (CP) El Niño, which favors initiation and development of deep convections. This is the reasons for the enhanced convection activities over the eastern (central) Pacific during boreal winter of EP (CP) El Niño (Figs. 1 and 2).

It has been shown that the low-level tropospheric moistening to the east of MJO convection center might be important for its eastward propagation (Andersen and Kuang 2012; Ling et al. 2013; Zhao et al. 2013; Li 2014; Hsu et al. 2014). The moisture budget equation at pressure level is used to diagnose the tropospheric moistening in this study as follows (Hsu and Li 2012):
\[ \frac{\partial q}{\partial t} = -u \frac{\partial q}{\partial x} - v \frac{\partial q}{\partial y} - \frac{\partial q}{\partial p} \frac{Q_z}{L}, \]

where \( q, u, v, \omega, Q_z, \) and \( L \) denote the specific humidity; zonal, meridional, and vertical velocity; atmospheric apparent moisture sink; and latent heat of condensation, respectively. The first three terms on right-hand side of Eq. (5) are moisture tendency due to the zonal, meridional, and vertical advection. They mainly reflect the moisture exchange between the MJO convection and the environment. The moisture tendency due to the atmospheric apparent moisture sink, which is the fourth term on the right-hand side of Eq. (5), is mainly dominated by the inner physical processes in the subgrid scale, including the cumulus convection, large-scale condensation, and evaporation (Chikira 2014).

The time evolution of regressed large-scale MJO moisture tendency due to advection, which is the sum of the first three terms on the right-hand side of Eq. (5), and the corresponding zonal–vertical circulation averaged over 10°S–10°N are further compared between the two types of El Niño during boreal winter in Fig. 7. Both the low-level moisture advection and zonal–vertical circulation of MJO are stronger and deeper, and their zonal scale is larger in CP El Niño. The low-level positive moisture advection is to the east of the MJO convection center as reflected by the regressed MJO OLR. The large-scale circulation and positive moisture advection have a notable eastward propagation feature in both types of El Niño. Such MJO zonal–vertical circulation and low-level moistening are totally lost in EP El Niño when active convection of the MJO moves to around 170°E at day 12 (Fig. 7g), whereas they are still very strong in CP El Niño (Fig. 7h). This may be the reason for the different intensity and propagation features of MJO in the two types of El Niño (Fig. 5).

The time evolution of moisture tendency due to the atmospheric apparent moisture sink is almost the same as that due to the advection except for the opposite sign (not shown). To further explore the relative importance for each individual term on the right-hand side of Eq. (5), they are first regressed onto the reference MJO OLR time series and then averaged over 1000–500 hPa and 170°E–170°W, which is at the low-level troposphere to the east of the MJO convection center at day 8, and the results are shown in Fig. 8a. It is clearly shown that the value of local moisture change (\( \frac{\partial q}{\partial t} \)) is very small compared to the moisture advection and atmospheric apparent moisture sink in both EP and CP El Niño. In EP El Niño, meridional and vertical advection of moisture lead a negative moisture tendency, while the zonal advection and atmospheric apparent moisture sink lead a positive one. The total effect of moisture advection is to reduce the moisture to the east of MJO convection center. The positive atmospheric apparent moisture sink to the east of the MJO convection center corresponds to the negative latent heating release (evaporation) there. Even if the evaporation to the east of MJO convection contributes a positive moisture tendency, it is almost canceled out by the moisture advection there; furthermore, such negative latent heating corresponding to the positive atmospheric apparent
moisture sink will further induce the downward motion and lead the atmosphere to be more stable there. Therefore, the eastward propagation signal of MJO almost disappears after day 8 in the central Pacific Ocean in EP El Niño (Fig. 7g).

However, in CP El Niño, the effort for each term is almost opposite to that of EP El Niño. The zonal advection and atmospheric apparent moisture sink have negative effects to the moisture tendency, while the meridional and vertical advection lead to a positive moisture tendency. The upward motion delivers the moisture from the bottom to the upper level in the troposphere and enhances condensation, which corresponds to the negative effort of the atmospheric apparent moisture sink. Such positive latent heating released from the condensation will further induce upward motion and lead to the moisture convergence there. The importance of diabatic heating in the lower level in MJO simulation has been already indicated by Li et al. (2009). The adequate moisture and strong upward motion in the central Pacific to the east of MJO convection center will lead to the MJO signal continuing to propagate eastward after day 8 in EP El Niño (Fig. 7h).

Chikira (2014) regarded the net effect of moisture tendency due to the vertical advection and cloud processes as the “column process.” The atmospheric apparent moisture sink is almost dominated by the cloud processes. It is reasonable and convenient to use the column process to diagnose the moisture tendency; however, it also conceals many aspects of the moisture change. Such a column process works as a moistening...
factor both in CP and EP El Niño to the east of the MJO convection center and day 8 (Fig. 8a) even though the value is much smaller in EP El Niño. In EP El Niño, the downward motion delivers the moisture out of the convection and works as a drying factor, while the evaporation works as a moistening factor. The net effect of the column process leads the atmosphere more stable. In CP El Niño, the vertical advection works as a moistening factor while the condensation works as a drying factor to the atmosphere. The net effect leads the atmosphere to become more unstable. When analyzing the low-level moistening to the east of MJO convection, we need to consider not only the tendency of moisture, but also the source of the moisture and the stability of atmosphere (the release of latent heating), which is important in maintaining the eastward propagation of MJO and may be neglected using analysis of the column process. Therefore, both the different of vertical advection of moisture, which is the dominant component of the moisture advection term, and the corresponding atmospheric apparent moisture sink are the most important factors accounting for the different eastward propagation characteristics of MJO between CP and EP El Niño in the central Pacific.

Based on the mass continuity, the vertical advection term can be further decomposed into the zonal $q(\partial u/\partial x)$ and meridional $q(\partial v/\partial y)$ moisture convergence and the vertical moisture flux convergence $(\partial q/\partial p)$ following Hsu and Li (2012). The vertical averages (1000–500 hPa) of these three terms to the east of the MJO convection center at day 8 are shown in Fig. 8b. It is clearly shown that the zonal and meridional moisture convergences have somehow opposite effects with each other, and the vertical moisture flux convergence is relative small compared to them. Figure 8b clearly shows that the strong positive vertical advection of moisture ($\omega \partial q/\partial p$) is mainly attributed to the zonal moisture convergence ($q \partial u/\partial x$).

To identify the relative contribution of eddy–eddy and eddy–mean flow interactions following Hsu and Li (2012), the specific humidity and zonal and vertical velocity can be further decomposed into three components, the low-frequency background state (LFBS; with a period longer than 90 days), the MJO (30–90 day) component, and the high-frequency (with a period shorter than 30 days) component following Zhao et al. (2013):

$$q = q + q^*; \quad u = \bar{u} + u' + u^*; \quad \omega = \bar{\omega} + \omega' + \omega^*.$$  \hspace{1cm} (6)

where the overbar, prime, and asterisk denote the LFBS, MJO, and high-frequency components, respectively. Therefore, the terms $q(\partial \bar{u}/\partial x)$ and $\omega(\partial q/\partial p)$ can be further divided into nine terms as follows:

$$q \frac{\partial \bar{u}}{\partial x} = \bar{q} \frac{\partial u}{\partial x} + \bar{q}^* \frac{\partial u^*}{\partial x} + \bar{q}^* \frac{\partial u^*}{\partial x} + \bar{q} \frac{\partial u'}{\partial x} + \bar{q}^* \frac{\partial u^*}{\partial x} + q^* \frac{\partial u^*}{\partial x}$$

$$+ \bar{q} \frac{\partial u^*}{\partial x} + q^* \frac{\partial u^*}{\partial x} + q^* \frac{\partial u^*}{\partial x} \quad \text{and}$$

$$\omega \frac{\partial q}{\partial p} = \bar{\omega} \frac{\partial q}{\partial p} + \bar{\omega}^* \frac{\partial q^*}{\partial p} + \bar{\omega} \frac{\partial q}{\partial p} + \bar{\omega}^* \frac{\partial q^*}{\partial p} + \omega^* \frac{\partial q^*}{\partial p} + \omega^* \frac{\partial q^*}{\partial p}$$

$$+ \bar{\omega} \frac{\partial q^*}{\partial p} + \omega^* \frac{\partial q^*}{\partial p} + \omega^* \frac{\partial q^*}{\partial p}. \quad \text{(7)}$$

The vertically integrated (from 1000 to 500 hPa) terms from the right-hand side of Eqs. (7) and (8) averaged over 10°S–10°N, 170°E–170°W at day 8 shown in Fig. 9. It indicates that the zonal convergence of moisture is mainly attributed to the convergence of LFBS specific humidity induced by the MJO zonal wind (Fig. 9a). This term is much stronger in CP than EP El Niño. The other terms from the right-hand side of Eq. (7) contribute little to the zonal convergence of moisture. The vertical advection of moisture is mainly attributed to the advection of LFBS specific by the MJO vertical velocity (Fig. 9b). This term is positive in CP El Niño and negative in EP El Niño. The combined efforts of LFBS specific humidity and MJO wind play an important role in both zonal moisture convergence and vertical moisture advection. They are may be the key reason for the different propagation characteristics of MJO between these two types of El Niño.

The composite results of anomalous LFBS specific humidity (Fig. 6) and the regressed MJO zonal–vertical circulation at day 8 (Figs. 7e,f) show that both anomalous LFBS specific humidity and MJO circulation differ significantly between the two types of El Niño. They are
weaker in EP than in CP El Niño to the east of MJO convection center. The strong anomalous moistening over the eastern Pacific due to the warming sea surface there in EP El Niño is much too far east of the convection center and the MJO easterly is mainly to the east of 150°W. Such a distribution of specific humidity and zonal wind cannot lead to low-level tropospheric moistening near the date line and support the initiation of the convection to the east of an existing convection center of the MJO. In contrast, maximum anomalous specific humidity is located just east of the date line and the anomalous easterly of the MJO in CP El Niño prevails. The anomalous easterlies lead to robust moistening to the east of an existing convective center and set a favorable state for the generation of deep convection there. The MJO vertical velocity is prominently upward in CP El Niño in the lower troposphere to the east of its convection center at day 8. This is not so in EP El Niño. Such a distribution of the vertical velocity in CP El Niño can transport moisture from the planetary boundary layer to the free atmosphere to help the generation of deep convection. Therefore, the different distributions of LFBS moisture and MJO circulation lead to the different characteristics of MJO eastward propagation in the two types of El Niño.

6. Summary and discussion

El Niño events are classified into the eastern Pacific (EP) type and the central Pacific (CP) type according to the locations of their maximum positive SST anomalies. The differences in the intensities and locations of SST anomalies between these two types of El Niño lead to their different impacts on the tropical atmosphere, especially on convection. Enhanced OLR over the western Pacific appears in boreal spring and early summer before the onset of CP and EP El Niño. During the mature stages, OLR is strengthened over the eastern (central) Pacific in EP (CP) El Niño, and weakened over the western Pacific.

Further analyses show that MJO activities over the western Pacific in boreal spring and early summer are closely related to EP El Niño 2–11 months later. In contrast, there is no such significant lagged relation between the MJO and CP El Niño. MJO activities are
enhanced over the Indian Ocean and western Pacific prior to EP El Niño and reduced during its mature and decaying stages. There is no strengthened MJO over the Indian Ocean or western Pacific closely related to CP El Niño before its occurrence. But MJO activities near the date line are strengthened during the mature and decaying phases of CP El Niño, which may be due to positive SST anomalies there.

El Niño impacts the intensity and propagation of the MJO mainly through the anomalous circulation related to El Niño. Anomalous divergence and descending motions over the western Pacific during El Niño lead to the weakened convection and insufficient moisture flux convergence there, which result in reduced MJO activities through the feedback effect of convective heating. The anomalous circulation related to EP El Niño is stronger, so its impacts on the MJO are also more prominent. The increased intraseasonal (30–90 day) oscillation over the eastern Pacific during the mature phase of EP El Niño may be due to increased SST there and may not directly relate to the MJO over the Indian and Pacific Oceans. The eastward propagation of the MJO is more continuous and farther eastward during CP El Niño, because strong and deep low-level atmospheric moistening to the east of the MJO convection center can still be maintained even if the MJO passes the date line. Such low-level moistening totally disappears during EP El Niño when the MJO reaches around 170°E. The robust low-level moistening in CP El Niño is primarily due to the LFBS specific humidity advection by the MJO upward motion, which is induced by the convergence of LFBS specific humidity by the zonal gradient of MJO zonal winds. Both LFBS specific humidity and MJO circulation are stronger to the east of the MJO convection center in CP than EP El Niño, which leads to stronger low-level moistening and more continuous eastward propagation of the MJO. Furthermore, the different characteristics of latent heating release corresponding to the atmosphere’s apparent moisture sink are also important in leading to different eastward propagation of MJO in the two types of El Niño.

Detailed results in this study may be influenced by the limited sample size of El Niño events used in the diagnosis, but the main evolution characteristics of the MJO in the two types of El Niño are robust. There are only three EP El Niño events after 1975. However, four more EP El Niño events (1951/52, 1963/64, 1965/66, and 1972/73) were further identified during the period from 1948 to 1975, which are only covered by the NCEP reanalysis and SST data. The composite results using all the seven events (not shown) are almost the same as those shown in this study. Regression results using a longer period (from 1948 to 2011) also show the same results. The relationship between the MJO and the two types of El Niño investigated in this study is only the linear relationship between them. Their nonlinear relation needs further investigation. A comprehensive understanding of the interaction between the MJO and the two types of El Niño may need further studies using numerical simulations.

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