Cumulative Positive Contributions of Propagating MJO To The Quick Low-Level Atmospheric Response During El Niño Developing Years

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Abstract

Based on Australian Bureau of Meteorology (BoM) El Niño alert system, this study investigates the atmospheric and oceanic conditions during El Niño developing years between 1982 and 2016. It is found that there is a 2–5-month lag to establish steady low-level atmospheric (or the Southern Oscillation Index, SOI) response than the steady El Niño-pattern Sea Surface Temperature Anomaly (SSTA), which is defined as the critical period in this research. According to the duration of this critical period, the quick and slow steady atmospheric response years can be identified among all El Niño–Southern Oscillation (ENSO) developing events. The quick establishments of the Sea Level Pressure Anomaly (SLPA) in the tropical atmosphere are proved to be closely related to the subseasonal Madden–Julian Oscillation (MJO) events. In the quick response years, the MJO events can even propagate to the eastern Pacific, which lead to cumulative negative Outgoing Longwave Radiation (OLR) and SLP anomalies there, and make a positive contribution to the quick atmospheric response at the end of critical period. However, the eastward-propagation of MJO events is mainly restricted in the tropical Western Pacific in the slow response years, causing slow steady atmospheric response with almost no contributions from MJO. Furthermore, observations and several simulations are used to understand this propagation differences of the MJO between quick and slow response years.

Key Points

- A time-lag is proved to exist between the steady SST and SLP anomalies in the El Niño developing year, which defined as the critical period.
- The key factors of MJO propagations are found to be different during the critical periods of quick and slow response years by using observation and several simulations.
- The different cumulative effects of eastward MJO propagation in the critical periods are proved to dominate the establishment of steady low-level atmospheric response over the eastern Pacific on the interannual timescale.

Introduction

The El Niño–Southern Oscillation (ENSO) events, which are the most prominent events with interannual variation in the tropical area, affect climate and weather globally (Bjerknes 1969; Wyrki 1975; Philander 1983; Sohn and Park 2010). Corresponding to the El Niño-pattern Sea Surface Temperature Anomaly (SSTA), the coupled ocean-atmosphere system is also characterized by a major updraft of deep atmospheric convection over the central-eastern Pacific and a Sea Level Pressure Anomaly (SLPA) over the tropical Pacific (Albrecht and Cox 1975), which is measured as the Southern Oscillation Index (SOI).

Using the SOI and Niño3.4 indexes, the evolutions of atmosphere and ocean during El Niño can be unveiled (Nicholls 2008). However, considering the steady maintaining of the anomalies, a time-lag is found between the establishment of steady SLPA and the steady El Niño-pattern SSTA in the El Niño developing year (Fig. 1). The time-lag is defined as the critical period in this article, which can be divided into two categories according to the durations (the detailed definition of the critical period is in section 3). The time-lag coupling between atmosphere and ocean was firstly discovered by Bjerknes (1969). Furthermore, Johnson et al. (2019) pointed out that the atmospheric response lags 4 months to the steady El Niño-pattern SSTA during 2015/16 El Niño. Other studies proposed that the subseasonal signals such as the Madden–Julian Oscillation (MJO) should be further considered (Madden and Julian 1972; Chen et al. 2020; Liu et al. 2020), but this issue has not been explored systematically. The objective of the present work is to analyze the subseasonal MJO effects on interannual atmospheric response during El Niño developing years.

As the most pronounced subseasonal signal, MJO is known as a planetary scale eastward-propagating mode with approximately period of 30–90 days (Rui and Wang 1990; Li et al. 2015). Plenty researches proved that the MJO events can affect the atmospheric response during El Niño through direct or indirect ways (Power and Smith 2007; Murakami et al. 1986; Song 1994). The direct effects of MJO events on the atmospheric response is through the large-scale wind field (Power and Smith 2007; Xiang et al. 2014). During the MJO Pacific phases, the strong westerlies of the propagating MJO events can weaken the low-level easterlies of Walker circulation during the neutral ENSO condition (Zhang 2005; Guan et al. 2019; Chu et al. 2020). Modulated by the atmospheric circulation field of MJO, Murakami et al. (1986) revealed that the intensity of Walker circulation has a subseasonal variability. Besides, the ascending branch of Walker circulation differs substantially considering the effects of MJO (Song 1994). Due to the above direct effects from MJO, Slawinska et al. (2014) put forward a low-frequency variable index to evaluate the subseasonal variation of Walker circulation in a cloud-analyzed model. The indirect effects of MJO events on the atmospheric response during El Niño are the corresponding oceanic

processes (Wang and Weisberg 2000). The MJO events trigger the Westerly Wind Bursts (WWB) at lower troposphere, resulting in eastward transport of warm water from the western to the eastern Pacific (Yoshida 1959; Penland 1996; Chen et al. 2019; Wang et al. 2019). Meanwhile, the MJO events associate WWB force the oceanic Kelvin waves, that in turn propagating eastward contributes to the deepening of thermocline, yielding the warming of central-eastern Pacific (McPhaden 2002; Lybarger and Stan 2019). Furthermore, some studies about the MJO effects turned to longer timescales, pointing out a decadal shift of the relationship between MJO and ENSO (Gushchina and Dewitte 2011, 2018).

Since boreal summer to autumn (May–October) are always within the critical period we focused (Fig. 2), the special Intra-Seasonal Oscillation (ISO) events in boreal summer should be considered. In boreal summer, the ISO events are regarded as the Boreal Summer Intra-Seasonal Oscillation (BSISO) events, which differ significantly in the characteristics of northward propagation and variation of 10–90 days (Rui and Wang 1990; Chen et al. 2020). The study of BSISO events effects on atmospheric response are relatively deficient (Liu et al. 2016; Wu and Cao 2017). Liu et al. (2016) revealed the ENSO effects on the BSISO, which the magnitude of 10–20-day ISO is enhanced during El Niño developing summer; whereas the 30–60-day ISO is enhanced when La Niña decays. For the decadal timescales, Wu and Cao (2017) implicated that compared to 1990s, BSISO had a closer link with ENSO during 1980s. Most of the previous works diagnosed the intensity and position of Walker circulation under modulation of MJO events, but less attentions were paid to the different atmospheric response time during the El Niño events.

Above all, there remain several questions still need to be answered: does it exist a relatively quick or slow atmospheric response during El Niño events commonly? Furthermore, how does the quick or slow atmospheric response phenomenon relate to subseasonal signals such as MJO? Based on the above issues, the outline of the paper is as follows. Section 2 contains introduction of the data, the El Niño alert system to identify the atmospheric response time, and the numerical model we used. Then, we classify the quick and slow atmospheric response years, and figure out the main distinction between them in Section 3. Section 4 analyzes the different performances of MJO events in observations between the quick and slow steady atmospheric response years. An idealized atmospheric model and multivariable diagnostic analyses are used to explore the essential factors of the diversity of MJO events between the quick and slow response years. Section 5 presents how the subseasonal MJO events with “shorter period” finally affect the interannual atmospheric response with “longer period”. Finally, the summary and discussion are presented in Section 6.

Data And Method

2.1 Data introduction

The data sets used in this research include daily SLP from the National Aeronautics and Space Administration (NASA) Modern-Era Retrospective analysis for Research and Applications, Version 2 reanalysis (MERRA2; Gelaro et al. 2017), and daily SST from the Optimum Interpolation Sea Surface Temperature (OISSTV2; Reynolds 2007), dataset developed by the National Oceanic and Atmospheric Administration (NOAA). The observed daily 3-dimension variables include horizontal and vertical winds from National Centers for Environmental Prediction (NCEP)-DOE2 with a resolution of 2.5° latitude $\times$ 2.5° longitude (Kanamitsu 2002). Outgoing Longwave Radiation (OLR) is a good proxy for atmospheric convection and is widely adopted to study the MJO (Wheeler and Hendon 2004; Li et al. 2020). The daily OLR with a horizontal resolution of 2.5° latitude $\times$ 2.5° longitude is obtained from National Center for Atmospheric Research (NCAR)/ NOAA. To demonstrate the location and intensity of MJO events, the phase and amplitude data of BoM Real-Time Multivariate MJO (RMM) indexes are used (Wheeler and Hendon 2004). In addition, the monthly precipitation data with a horizontal resolution of 2.5° latitude $\times$ 2.5° longitude from the Global Precipitation Climatology Project (GPCP) are applied (Allan 2014). All the data sets we used in this research are from year 1982 to 2016.

2.2 Model introduction

A 2.5-layer idealized model is used to investigate the key factor of the distinctive MJO characteristics during the critical periods (Wang and Xie 1997). The model is designed for analyzing equatorial waves and coupled convection. Previous studies (Wang and Xie 1997; Liu et al. 2016) provided robust evidences that this model is capable of simulating the basic features of MJO in boreal summer, such as the one-wave propagation structure and the corresponding rain band. In this linearized model, the continuity and momentum equations are written at the upper (200hPa) and lower (850hPa) troposphere in a spherical p coordinate. To predict geopotential height anomalies, the model applies a stationary barotropic Planetary Boundary Layer (PBL). The horizontal resolution of the model is 5° latitude $\times$ 2° longitude with a 10-minute time step over the field (0° – 360° , 40° S– 40° N). Central difference integration scheme is adopted in both space and time for higher accuracy.

The SST and u-component of wind at 200 hPa and 850 hPa are adopted to force the model. The input SST data are obtained from NOAA high-resolution blended analysis dataset with a resolution of 0.25° latitude $\times$ 0.25° longitude, and the wind data are derived from NCEP/DOE AMIP-II Reanalysis (Reanalysis-2) with a coarse resolution of 2.5° latitude $\times$ 2.5° longitude. All the input data are interpolated into the resolution of 5° latitude $\times$ 2° longitude to adapt the model resolution. To solve the matrix inversion more quickly, the Fourier transform is applied to truncate 25% of the zonal grids points within PBL (Bracewell 2002). The deep convection parameterized scheme is following Kuo's (1974) work, which bases on analyses of heat and moisture budget. In addition, the precipitation efficiency coefficient is an adjustable key parameter to measure the moisture convergence feedback within PBL in this model (Wang and Xie 1997).

2.3 The El Niño alert system

Nowadays, NOAA and BoM El Niño alert systems are widely used. The NOAA alert system categorizes El Niño cycle into 3 classes, which include inactive, watch El Niño and advisory El Niño (<https://www.climate.gov/news-features/understanding-climate/el-ni%C3%B1o-and-la-ni%C3%B1a-alert-system>). In contrast, the BoM El Niño alert system includes 4 phases: inactive, watch El Niño, alert El Niño and El Niño phases (<http://www.bom.gov.au/climate/enso/outlook/#tabs=Outlook>). In particular, the BoM alert system emphasizes the oceanic and atmospheric statuses by using the SSTA and SOI indexes, which are the key points of this research. Thus we applied BoM alert system in the present research and focused on the El Niño phases in the atmosphere and ocean, respectively.

Based on monitoring the evolution of key indicative indexes (Niño3.4 and SOI) and the major simulations of climate systems, the BoM El Niño alert system can give the outlook of ENSO. Niño3.4 is the averaged SSTA of the specific region (5°N – 5°S , 170°W – 120°W), which this region has large variability during El Niño and is important for the rainfall typically located in the far western Pacific (Rasmusson and Carpenter 1982). The BoM El Niño phase is identified when the overlapping 3-month Niño3.4 values exceed $+0.8$ for 5 consecutive months (<http://www.bom.gov.au/climate/enso/outlook/#tabs=Criteria>). The SOI is a standardized index according to the different SLP between Tahiti ($148^\circ05'\text{W}$, $17^\circ53'\text{S}$, eastern Pacific) and Darwin station ($130^\circ59'\text{E}$, $12^\circ20'\text{S}$, Australia).

The calculation equation of SOI is showed at <https://www.ncdc.noaa.gov/teleconnections/enso/indicators/soi/>. The SOI can measure the large-scale oscillation in air pressure over the tropical Pacific, such as the Southern Oscillation. There are several criterions for SOI to indicate El Niño condition. For instance, BoM cites the sustained SOI values below -7 as an El Niño, while the thresholds of SOI values is below -6 in some other websites (https://www.weatherzone.com.au/climate/indicator_enso.jsp?c=soi; <https://www.eldersweather.com.au/climate/?mode=soi&period=weekly>). The above standards are dependent on specific time periods they studied and different calculation methods of SOI (Ropelewski and Jones 1987; Power and Kociuba 2011). To eliminate the uncertainty caused by the different periods and methods, here we calculated the averaged SOI ($\text{SOI} < -4.9$) during the period when the monthly Niño3.4 reaches the El Niño condition as the atmospheric criterion of El Niño condition. Several other criterions of SOI had also been tried in the supplementary information and have no effect on the results of the research (Fig. S3–S5).

3 Whether the atmosphere and the ocean reach the steady El Niño conditions at the same time?

Firstly, the oceanic and atmospheric conditions in two special El Niño cases (1982/83 and 1997/98) were diagnosed. Considering the atmospheric alter time, we plotted the SSTA patterns when monthly Niño3.4 first exceeds $+0.8$ and the corresponding SLPA which lags three months (Fig. 1). The maximum SSTA at this time in year 1982 is $+2.1^\circ\text{C}$ over the equatorial eastern Pacific (85°W – 80°W , 5°S – 5°N); whereas in year 1997, the maximum SSTA is $+3.0^\circ\text{C}$ with a much wider domain (95°W – 80°W , 15°S – 15°N) (Rinsland et al. 2001; Garcia et al. 2004). The SLPA results suggest that the related low-pressure regions can extend to the central Pacific (140°W – 130°W) in year 1982 three months after the Niño3.4 reaching the criterion of El Niño condition. However, a weaker negative SLPA is restricted in the northeast tropical Pacific (100°W – 90°W) 3 months after the 1997/98 El Niño SSTA sets up. It indicates that the SLPA in tropical atmosphere establishes more quickly and strongly during 1982/83 El Niño case.

Furthermore, the evolution of dekadal Niño3.4 and SOI in all 6 El Niño cases from year 1982 to 2016 were plotted in the supplementary information (Fig. S1). Table 1 present the dekads when the oceanic (Niño3.4) and atmospheric conditions (SOI) reaching the steady El Niño statuses in different El Niño cases. The table results show that the earliest onset date of ocean (Niño3.4) reaching steady El Niño condition is April, and the latest date is November. On average, the ocean (Niño3.4) reaches steady El Niño condition in July. Previous researches also proved that boreal summer (June–August) is the developing stage during El Niño evolution (Trenberth et al. 2001; Levine and McPhaden 2016). The earliest onset date of SOI reaching the steady El Niño condition is October; whereas the last date is January in El Niño mature year. It shows that there is a time-lag to establish steady SOI anomaly than the steady El Niño-pattern SSTA among all El Niño cases between 1982 and 2016. We furtherly defined the time interval between the steady atmospheric response (SOI)

relative to the steady El Niño SSTA (Niño3.4) from the El Niño developing summer to autumn as the critical period. The critical period always exists for all the El Niño events, which lasts 12 dekads on average. Typically, the critical period lasts for only 3 dekads in 2009/10 El Niño case; however, the critical period in 2015/16 lasts for 25 dekads. According to the duration of this critical period, quick (less than 12 dekads, which is the average response time of all the El Niño events) and slow steady atmospheric response years can be identified.

The classifications of quick and slow response years are not sensitive to different atmospheric or oceanic thresholds. The same classifications can be obtained by applying the Equatorial Southern Oscillation Index (ESOI) developed by NOAA in the supplementary information (Fig. S2–S3) or using different threshold values of the SOI and Niño3 indexes (Fig. S4–S5).

Table 1. The dekad when the ocean (or the Niño3.4) and atmosphere (or the SOI) reaches the steady El Niño conditions in different El Niño cases. The details of the oceanic and atmospheric evolutions are in Fig. S1. The 2nd and the 3rd column are the dekad when ocean and atmosphere reaching the steady El Niño conditions, respectively. According to the durations of the time interval between the 3rd and the 2nd column, quick (bolded and underlined) and slow steady atmospheric response years can be identified. The 1982/83, 1991/92, 2009/10 El Niño cases are quick response years, and the 1987/88, 1997/98, 2015/16 El Niño cases are slow response years. The classifications exceed the 95% confidence level through the results of T-test.

El Niño cases	The dekad when ocean reaches El Niño condition	The dekad when atmosphere reaches El Niño condition
<u>1982/83</u>	Aug (2nd)_1982	Oct (1st)_1982
1987/88	Jun (2nd)_1987	Jan (1st)_1988
<u>1991/92</u>	Nov (1st)_1991	Dec (2nd)_1991
1997/98	Jun (1st)_1997	Dec (2nd)_1997
<u>2009/10</u>	Sep (3rd)_2009	Nov (1st)_2009
2015/16	Apr (2nd)_2015	Jan (1st)_2016

The composite evolutions of dekadal Niño3.4 and SOI indexes during the El Niño cycles are distinctive in the quick and slow response years (Fig. 2). In the quick response years, the composite Niño3.4 index (blue solid line) reaches steady El Niño condition in September (blue star symbol) and then promotes rapidly during the critical period. The SOI (blue dashed line) exhibits a weak positive value in May and then decreases steadily to establish steady atmospheric response during the critical period (60 days). In the slow response years, the onset dates of Niño3.4 and SOI indexes reaching steady El Niño condition are May of El Niño developing years and January of mature years, respectively. Specifically, the Niño3.4 index increases slightly from May and reaches its maximum intensity in November (red solid line). The evolution of SOI in the slow response years is significant different from quick response years, which takes much longer time (180 days) to reach steady atmospheric El Niño condition although Niño3.4 gets steady El Niño condition earlier. In slow response year, the SOI develops from May with a weak negative value, indicating the atmospheric response during El Niño begins to set up. From May to September, the evolution of SOI is relatively steady with no obvious downward trend. The SOI curve turns to rise in the autumn (SON), but it sharply decreases to reach a stable El Niño condition in December (red circle symbol).

We further plotted the different spatial precipitation rate and horizontal wind fields during periods 1 and 2 of quick and slow response years (Fig. 3). The most notable differences of the precipitation rate patterns are the longitudinal anomalies which locate over the equatorial Pacific (Zhao et al. 2020). During periods 1 (Fig. 3a), the negative precipitation rate anomalies over the equatorial Pacific indicate the precipitation in the quick response years is relatively less here. However, the precipitation rate modes during periods 2 are almost opposite to that during periods 1, which display as an obvious positive precipitation rate center over the equatorial Pacific. The horizontal wind field indicates a shift from easterlies to westerlies over the equatorial Pacific. The different patterns of precipitation rate and horizontal wind are physically reasonable. Before September, both the quick and slow response years have not reached the steady atmospheric El Niño condition (Fig. 2). However, the SOI presents a weak negative anomaly in the slow response years, indicating the SLP of Tahiti station (eastern Pacific) is weaker than Darwin station (Australia). It may be due to the updraft of the Walker circulation has already moved from Maritime Continent (MC)-western Pacific to central-eastern Pacific. Instead, the SOI value is positive during periods 1 of quick response years, suggesting the main precipitation area is over MC-western Pacific. Therefore, during periods 1 of

quick response years, the intensity of precipitation over the equatorial Pacific is weaker than the slow response years. During periods 2 of quick response years, the atmospheric condition has already reached the steady El Niño condition, whereas the slow response years have not reached yet. Thus, the establishment of steady atmospheric response in quick response years leads to the enhanced convective activity over the central Pacific. It proves that the atmospheric responses (wind and precipitation) experience a considerable change after the critical period in the quick response years.

In this section, we revealed a time-lag of steady atmospheric response relative to the steady El Niño-pattern SSTA in the El Niño developing year, which is defined as the critical period. According to the duration of critical period, the quick response years and slow response years are classified. What cause the quick or slow atmospheric response? What's the mechanism of the critical period? Hereafter, this study will focus on these key issues.

4 Different characteristics of MJO events during the critical periods in the quick and slow response years

4.1 Observed MJO propagation differences

Since the different time-lag of the critical periods between the SOI quick and slow response years is behaved on subseasonal timescales, it's reasonable to hypothesize that the atmospheric response can be modulated by MJO events, which is the most significant subseasonal signal over the tropical area (Rui and Wang 1990; Li et al. 2015).

Firstly, we used 30–90-day band passed filtered OLRA to show the composite MJO life cycles in the critical periods of quick (Fig. 4a) and slow response years (Fig. 4b). The most remarkable similarity is the one-wave eastward propagation, which is recognized as the typical structure of the MJO. However, the range and extensity of eastward-propagating mode differs between quick and slow response years. In the quick response years, there are significant negative OLRA over the tropical western Indian Ocean in phases 1–2, indicating that active convective activity of MJO initiates from here. In phases 3–4, the convective center of MJO propagates eastward to the western MC. In phases 5–6, the eastward-propagating speed increases and the convection is enhanced after passing over the MC. In phases 7–8, the convective center continues to propagate eastward, extending to central-eastern Pacific (120°W). In slow response years, the eastward-propagating mode is not so obvious. In phase 1–2, the initial position of MJO convection is further east and the intensity is relatively weaker compared to quick response years. In phases 3–4, the convective center starts to extend northward. In phases 5–6 and 7–8, there barely exhibits eastward-propagating mode. Although the convective center still exists in the slow response years, the extension of MJO events is restricted in the west of 180°. The above results indicate that the MJO events can even propagate to the eastern Pacific in the quick response years. However, the eastward-propagation of MJO events is mainly suppressed over the tropical Western Pacific in the slow response years.

To show the RMM phases distributions of MJO in different dekads during the critical periods of quick and slow response years, we showed the phase-space diagram (Fig. 5). Different from the canonical phase-space diagram, we used the yellow and green dots to represent the MJO phases in the first and last dekads of the critical period, respectively. The distributions of the MJO phases in the first dekad show that the performance of initial MJO events is significantly distinctive. In the quick response years, the MJO phases in the first dekad concentrate in phase 1; whereas in the slow response years, the distributions are relatively scattered. It shows that the initial location of MJO events in the quick response years is further west and concentrated to western Indian Ocean compared to the slow response years. In the last dekad, the center of MJO convection can even reach the central-eastern Pacific and maintain relatively stable in the quick response years. In the slow response years, the MJO phases in last dekad are mostly distribute in the phases 5–7, which locates over the western-central Pacific, further west compared to quick response years. Since the critical period lasts 6 dekads in the quick response years, thus we showed the phase contributions of the 6th dekad in slow response years to get a comprehensive contrast. The distributions of 6th dekad suggest that the center of MJO convection are mostly locate in phases 5–6, which is districted by the "MC Barrier" (Ling et al. 2019; Li et al. 2020). It is clear that the propagation mode in the quick response years is more significant from the diagnoses of spatial propagation modes and the MJO phase-space diagrams.

To figure out the distinctive eastward-propagating modes of MJO events in the quick and slow response years, it is crucial to discriminate the eastward-propagating theories of MJO. The classical theory is the equatorial Wave–Conditional Instability of the Second Kind (Wave–CISK) feedback (Lau and Peng 1987; Sui and Lau 1989), which can explain the eastward propagation over the Indian Ocean and western Pacific. The Wind Induced Surface Heat Exchange feedback theory (WISHE, Emanuel 1987; Neelin 1987) complements the vacancy of propagation over the eastern Pacific. Furthermore, Wang and Li (1994) hypothesized the fictionally coupled Kelvin–Rossby wave theory, which includes the frictional convective coupled Kelvin–Rossby wave and evaporation–wind feedback process. Through the analyses of the column integrated Moisture Energy (MSE), previous works (Sobel et al. 2014; Wang et

al. 2017) reached a new understanding from the perspective of moisture mode. The above studies thus provide us with dynamic or moisture views to study the MJO propagation.

In the views of above-mentioned researches, the different RMM phase-vertical level distributions of atmospheric divergence, vertical velocity and specific humidity were displayed in [Fig. 6](#). Below the atmospheric boundary layer (red line), the different divergence between quick and slow response years features as a regional negative anomaly ([Fig. 6a](#)). A center of negative divergence anomaly appears from the MC to the western Pacific (phases 3–5) at the low-level of troposphere (1000–800hPa), indicating a pronounced convergence updraft in the quick response years. The enhanced convergence is more concentrated in the MC compared to the mean state, which benefits the cumulus convection of MJO passing over the “MC Barriers” and extending to the central-eastern Pacific. In [Fig. 6b](#), the results of vertical velocity show significant positive anomalies over the Indian Ocean and MC at the mid-troposphere (700–200hPa), suggesting a notable updraft here in the quick response years. The center of downdraft (negative vertical velocity anomalies) locates over the Pacific (phases 6–8), which does not benefit the development of convection. Consistent with the results of [Fig. 6a](#), the environmental dynamic conditions enhance the convective activity over the MC in the quick response years. [Fig. 6c](#) shows the distributions of specific humidity, which provides moisture to the cumulus convection of the MJO events. It indicates that moisture in the quick response years is more abundant over the Indian ocean and the upper level of the MC. Over the upper level of Pacific, the moisture is relatively less than the slow response years. Thus, both the dynamic conditions and moisture promote the convection over the MC in the quick response years, which benefits cumulus convection of MJO events passing over the MC and then propagating into the western-central Pacific. Over the Pacific, the atmospheric convergence plays a dominant role to maintain the eastward-propagation of MJO events in the quick response years.

The above observations show that in the quick steady atmospheric response years, the environmental convergence, vertical velocity and abundant specific humidity are favorable for convective center of MJO events to pass over the “MC barriers”. Over the Pacific, the significant convergence plays an important role for the propagation of MJO events. In sharp contrast, the extension of MJO events is restricted at dateline in the slow response years, because the background mean states cannot maintain the eastward-propagation of MJO events. In addition, the atmospheric vertical shear of zonal wind is important for the propagation of MJO events, and we will show the diagnosis in [section 4.2](#) (Zhang 2001; Mcphaden 2002).

4.2. Key factors of different MJO propagations between quick and slow response years in model simulations

The above observation analyses thus illustrated the variables and processes which may favor the pronounced eastward-propagating mode of MJO events in the quick response years. However, it is difficult to distinguish their relative contributions in observations. Hereafter, we designed several numerical simulations based on a 2.5-layer idealized atmospheric model developed by Wang and Xie (1997). The details of the model were introduced in [section 2.2](#). In our simulations, the environmental wind field, moisture convergence within the atmospheric PBL, the position of MJO initial atmospheric turbulence, and the SSTA pattern are designed as the forcings. [Fig. 7](#) shows the different forcing fields of the quick ([Fig. 7a](#)) and slow response years ([Fig. 7b](#)), which were inputted into the control runs later. The SSTA results imply that in boreal summer (May–October), the positive SSTA over the equatorial central-eastern Pacific in the quick response years are not as strong as the slow response years. The maximum SSTA are 1.0°C and 1.6°C in the quick and slow response years, respectively. The SSTA presents a weaker negative anomaly over the MC in the quick response years, which are relatively warmer than slow response years ([Fig. 7c](#)). The area of positive vertical shear of zonal wind is wider in the quick response years. We marked the initial atmospheric disturbances of MJO events according to the previous method (Zhang 2005; Straub and Katherine 2013). The MJO initial disturbance (red circle) of quick response years locates at 77°E with amplitude of 1.86 RMM (calculated from observation, red circle in [Fig. 7](#)), whereas in the slow response years, it locates at 87°E with amplitude of 1.49 (from observation, blue circle in [Fig. 7](#)).

In conclusion, compared to slow response years, the quick response years have relatively warmer SSTA over the tropical central-eastern Pacific, but colder SSTA over the MC. Over the north of the equator, the vertical shear of zonal wind of quick response years is larger than slow response years. The slow response years has stronger vertical shear of zonal wind over the south of equator. The position of MJO initial atmospheric disturbance is stronger and further west in the quick response years ([Fig. 7c](#)).

Using the associated MJO initial atmospheric disturbances and mean states ([Fig. 7](#)), two control experiments were conducted to simulate the development of MJO events in the quick and slow response years ([Fig. 8](#)). It shows that the MJO events initiate from the equatorial Indian Ocean, and then motivate geopotential height anomalies and corresponding precipitation. In the control run of the quick response years ([Fig. 8a](#)), the center of MJO precipitation splits into two centers in the central Indian Ocean from day 3. Then the

intensity of MJO precipitation enhances and the precipitation center propagates to the MC. During day 9 to day 13, the precipitation pattern is steady over the eastern Pacific, which exhibits notable precipitation over the Indian Ocean and the western Pacific. In contrast, the precipitation is relatively weaker in the slow response years, presenting as several weak signals in the Indian Ocean (Fig. 8b). Compared to the precipitation, the different subseasonal potential height anomalies between the quick and slow response years are more notable. In the quick response years (Fig. 8a), the range of the geopotential height anomalies of MJO extends steadily during day 1 to day 13. On the 13th day, the range of the negative geopotential height anomalies can extend to the tropical eastern Pacific (90°W). However, in the slow response years (Fig. 8b), the intensity of the anomalous geopotential height is much weaker. Furthermore, the propagation range is districted in the Indian Ocean at this point. During day 3 to day 9, the anomalous geopotential height is nearly stationary. From the day 9, the range of the geopotential height anomalies begin to shrink to central Indian Ocean. On the 13th day, the propagation is strongly suppressed over the MC. The above two control simulations confirm that this model can capture the eastward-propagating mode of MJO events. It proves that the oceanic and atmospheric backgrounds of the quick response years benefit the MJO propagation, leading to the stronger propagation of MJO events to eastern Pacific, which is more distinguishable to make the differences of control runs between quick and slow response years (CTL Diff in Fig. 9).

Four sensitive experiments were designed to explore the crucial factors of the distinctive MJO behaviors between the quick and slow response years. The experiment design is presented in Table 2. The sensitive simulations (Diff1–4) are used to distinguish the relative contribution of the MJO initial disturbances, vertical shear of zonal wind (U200 minus U850), SSTA pattern and moisture convergence feedback within the PBL between the quick and slow response years by the comparisons with the CTL Diff. The precipitation efficiency coefficient was adopted to measure the PBL feedback in Diff4, following the work of Sui and Yang (2007).

Table 2. Experiment design. Solid circle symbol represents inputting the respective conditions in the runs of quick and slow response years. The line represents inputting the climatological mean condition of El Niño events from 1982 to 2016. CTL Diff is defined as the differences between the quick and slow response control runs (Fig. 8), which is forced by all the above different processes simultaneously. Diff1–4 each represents the difference between two simulations for quick and slow response years considering one special forcing.

	Different MJO Initial disturbance for quick and slow response years	Different U200 minus U850 for quick and slow response years	Different SSTA for quick and slow response years	Different moisture feedback in the PBL for quick and slow response years
Difference of two control runs (CTL Diff)	●	●	●	●
Sensitive Diff1	●	---	---	---
Sensitive Diff2	---	●	---	---
Sensitive Diff3	---	---	●	---
Sensitive Diff4	---	---	---	●

Considering the combined or respective effect of simulation forcings, the direct differences between the quick and slow response years are showed in Fig. 9. The original simulation results are displayed in the supplementary information (Fig. S6– S9). In the CTL Diff (Fig. 9a), the precipitation and geopotential height anomalies of MJO mostly locate in the MC and extend significantly after day 3. In the Diff1, the distributions of MJO precipitation only concentrate in the MC, and the intensity of anomalous geopotential height is not so strong compared to the CTL Diff. What's more, the anomalous geopotential height turns to shrink to the eastern MC in day 13, which is distinctly different from the CTL Diff. The results of Diff1 indicate that the MJO initial disturbances is not the influential factor for the different MJO behaviors. However, the geopotential height anomalies in the Diff2 and Diff4 are relatively stronger, which similar to the CTL Diff. The similarity between the CTL Diff indicates the vertical shear of zonal wind and moisture feedback within the PBL

contribute more to the different geopotential height propagation mode. However, the precipitation patterns in all the simulations (control and sensitive runs) are different from the geopotential height propagations, which are restricted in the tropical western Pacific. In order to separate the contribution of each forcing on the geopotential height propagation and precipitation, the spatial correlation coefficients are calculated between the sensitive Diffs and the CTL Diff ([Table 3](#)).

For the geopotential height anomalies, it shows that the moisture convergence feedback within the PBL (Diff4) is the most important factor, which is also discussed in previous works (Szoeke and Maloney 2019; Hagos et al. 2019). Instead, the most important forcing for precipitation is the atmospheric vertical shear of zonal wind (Neena et al. 2017; Girishkumar et al. 2015). It is noteworthy that the SSTA takes the second position for both the precipitation and geopotential height anomalies of MJO, which means an indispensable effect of the air-sea interaction on MJO (Kapur and Zhang 2012; Liu and Wang 2016).

In summary, both the observation and simulation results illustrated that the propagation of MJO events are obviously different in quick and slow atmospheric response years. In the quick response years, MJO events can propagate further east to 120°W although the location of MJO initial disturbance is further west during the El Niño developing year. However, in the slow response years, the eastward-propagation of MJO events is mainly suppressed around the dateline and the MJO signals are much weaker and scattered. The observed moisture, atmospheric divergence and atmospheric vertical velocity are proved to support MJO events breaking the “MC barrier” in the quick response years. Furthermore, environmental vertical wind shear (U200 minus U850), moisture convergence within the atmospheric PBL, the position of MJO initial turbulence, and the SSTA pattern were designed as the forcings in the simulations, respectively. It proves that the MJO responses on geopotential height propagation and precipitation are different. The moisture feedback within PBL and the SSTA dominant the MJO eastward propagation of geopotential height, while atmospheric vertical shear of zonal wind and the SSTA are more important for the MJO precipitation.

Table 3. The averaged spatial correlation coefficients of geopotential height and precipitation between sensitive Diff1–4 and the CTL Diff for the first 13 days. The values with bold and underline are the most important factor, and the bold value takes the second place for each row.

Spatial correlation coefficients	Diff1 (Initial disturbance) & CTL Diff	Diff2 (U200-U850) & CTL Diff	Diff3 (SSTA) & CTL Diff	Di &	Diff4 (PBL feedback) &CTL Diff
For Geopotential height	0.37	0.74	0.81	<u>0.87</u>	
For Precipitation	0.37	<u>0.85</u>	0.78	0.31	

5 How does the "shorter period" MJO affect the "longer period" atmospheric response?

The above sections illustrated the distinctive eastward-propagating modes and the key factors of MJO events in the quick and slow response years. However, as a relatively “shorter-period” and “higher-frequency” process, how does MJO events finally impact the atmospheric response with “longer-period”?

Firstly, the evolution of dekadly mean 30–90-day filtered OLRA, 30–90-day filtered SLPA, and unfiltered SLPA during the critical period in quick response years are showed ([Fig. 10](#)). The filtered values represent the subseasonal MJO signals, and the dekad-averaged values can be regarded as the effects of MJO on the mean state. As seen in [Fig. 10](#), the negative centers of filtered OLRA initiate from tropical eastern Indian ocean to pass over the MC during the first 3 dekads, and then enhance to reach the tropical central-eastern Pacific (120°W) on the 6th dekad. Co-occurring with the eastward propagation of filtered OLRA, the negative centers of 30–90-day filtered SLPA propagate from the Indian Ocean to the central-eastern Pacific during the critical period of quick response year, leading a negative SLPA over the Tahiti station in the last dekad. On the 6th dekad, the negative centers of filtered convective signal (OLRA), filtered SLPA and the unfiltered SLPA locate over the Tahiti station, making a positive contribution for SOI to reach the threshold of atmospheric El Niño condition. To summarize, the center of MJO convective signals can even propagate to the eastern Pacific (120°W) during the critical period in the quick response years, leading to a significant negative value of OLRA and SLPA over the Tahiti station. However, a weak negative center of unfiltered SLPA presents over the central-eastern Pacific in the first dekad and increases rapidly because of the MJO effects and local positive Bjerknes feedback (Bjerknes 1969).

Similarly, the filtered OLRA, SLPA and unfiltered SLPA during the critical period of slow response years are presented ([Fig. 11](#)). The negative centers of filtered OLRA locate over the two sides of MC on the 1st dekad and then concentrate to the eastern MC on the 2nd

dekad. However, the propagation path of the filtered OLRA experiences a westward-propagating mode from the 3rd to 4th dekad and then propagate to the western MC on the 6th dekad, according with the analysis in [Fig. 5b](#). Note that the subseasonal filtered OLRA turn weak and scattered after passing over the MC (from the 11th to 18th dekad), indicating a weak eastward-propagating mode over the Pacific. Corresponding to the weak filtered OLRA, the filtered SLPA of slow response years exhibits few obvious negative signals near the equator during the whole critical period. Furthermore, the negative center of the dekadly mean unfiltered SLPA expands from the central-eastern Pacific to the central Pacific more slowly only under the modulation of local Bjerknes feedback. The suppression of negative unfiltered SLPA during 9th to 14th dekad is consistent with the increase of SOI during August–October in [Fig. 2](#). The steady atmospheric response anomaly (unfiltered SLPA) establishes later (180 days) in the slow response year with the restricted MJO propagation (only in the western Pacific).

Comparing the evolution of filtered dekadly mean convective signals and unfiltered SLPA in the quick and slow response years, it is found that the subseasonal signals during the critical period in the quick response years are more active and can reach the key region (central-eastern Pacific) for the establishment of atmospheric response. However, in the slow response years, the intensity of MJO events are relatively weaker and the extension cannot pass over the dateline to influence the atmospheric response.

The above diagnoses suggest that the subseasonal convective signals can propagate to the central-eastern Pacific in the quick response years, hence, the key issue is whether the subseasonal signals make contribution to the local atmospheric response interannual anomaly? Previous researches pointed out the asymmetric intensities between the MJO inactive and active convective phases and illustrated the amplitude of convective active phase is stronger, suggesting that the constructive effects of MJO events to local convection will not be offset by the destructive effects (Cronin and McPhaden 1997; Zhang and Anderson 2003). Therefore, the averaged 30–90-day filtered SLPA, 30–90-day filtered OLRA, the variance of the filtered OLRA and the unfiltered SLPA during the whole critical periods in the quick and slow response years are presented in [Fig. 12](#). In the quick response years, the averaged filtered OLRA exhibits a relatively stronger convective activity over the eastern Pacific ([Fig. 12a](#)), which indicates the positive cumulative contributions to the steady atmospheric response interannual anomaly. The variance of filtered negative OLRA during the critical period are plotted to show the dispersion of the convective activity ([Fig. 12c](#)). The large variance centers mainly locate in the tropical Indian Ocean, and several weak centers near the Central America and Tahiti station, which basically coincides with the convective center. The cumulative MJO effects on the SLPA also presents that a strong center of negative values locate over the Tahiti station but a positive value over the Darwin station, resulting in a stronger magnitude of negative SOI value in the quick response years ([Fig. 12e](#)). Coupling with the El Niño-SSTA pattern, the unfiltered SLPA are the comprehensive results of local Bjerknes feedback and MJO signals with relatively high-frequency timescales ([Fig. 12g](#)). The pattern of unfiltered SLPA shows an ascending branch anomaly over the eastern Pacific, and the Tahiti station is near the negative center.

The subseasonal convective contribution of the slow response years are showed in right column of [Fig. 12](#), similarly. The averaged filtered OLRA present weak suppressed convection with a maximum intensity of $+2 \text{ W/m}^2$ over most area of the central-eastern Pacific ([Fig. 12b](#)), which the maximum magnitude only takes 25% of quick response years ([Fig. 12a](#)). The variance of filtered OLRA are weak too, and there are only several weak centers around the MC ([Fig. 12d](#)). The filtered SLPA shows a south-north dipole mode in the slow response years ([Fig. 12f](#)). Both the Darwin and Tahiti stations are in the inactive phase of convection to present a positive value of averaged filtered SLPA, causing a weaker intensity of negative SOI value. [Fig. 12e](#) indicates that the Tahiti station locates at the edge of the negative SLPA center in the slow response years, presenting a relatively higher SLPA compared to the quick response years. Furthermore, the relatively cumulative contributions of MJO signals during 3 successive critical periods of the slow response events are shown in the supplementary information ([Fig. S10-S11](#)), which is split up by using the critical period of quick response events. It shows that the cumulative contributions of subseasonal MJO to the interannual atmospheric responses are relatively weaker in the slow response years than that in the quick response years.

In this section, we analyzed the evolutions and cumulative contributions of subseasonal convective signals related with MJO and the unfiltered total SLPA during the critical periods of quick and slow response years. In quick response years, the averaged filtered OLRA and SLPA show typical eastward-propagating mode and extend to the eastern Pacific. The MJO convective signals exhibit suppressed convection at the Darwin station (higher SLPA) and enhanced convection at the Tahiti station (lower SLPA) during the critical period, resulting in the negative SOI cumulatively. Hence, the establishment of steady atmospheric response is more quickly coupling with the warm SST. However, in the slow response years, the subseasonal MJO convection signals are restricted in the western Pacific with almost no cumulative effects. The establishment of atmospheric response is slow only with the local Bjerknes feedback.

Summary And Discussion

This research started from the different SLPA responses of two special El Niño cases (1982/83 and 1997/98), and we found that there exists significant quick or slow establishment of atmospheric response over the tropical Pacific (Fig. 1). A time-lag is proved to exist among all the El Niño events, which is defined as the critical period between the steady SOI anomaly relative to the steady El Niño-pattern SSTA in the El Niño developing years. Furthermore, quick and slow response years can be identified according to the duration of this critical period. Since the different time-lag between the quick and slow SOI response years is behaved on the subseasonal timescales, it makes sense to research the role of MJO events. Thus, we analyzed the composite MJO life cycles of OLR and phase-space diagrams to describe the distinctive characteristic of MJO events between quick and low response years (Fig. 4, Fig. 5). During the quick response years, it is found that the convective centers of MJO events can pass over the dateline to the tropical eastern Pacific, which are supported by the states of moisture, atmospheric divergence and atmospheric vertical velocity to break the “MC barrier” and extend to the further eastern position in the observation (Fig. 6). Furthermore, an intermediate model was applied to distinguish the key variables and processes. Through the contrast of the four sensitive simulations to the CTL run, it reveals that the crucial factors for geopotential height and precipitation anomalies of MJO signal are different. PBL feedback is the key factor to the geopotential height anomaly of MJO, and atmospheric vertical shear of zonal wind is dominant for the MJO precipitation. Moreover, SSTA is proved to be important for both MJO geopotential height and precipitation anomalies (Table 3).

Further study investigated how MJO events with “shorter period” alter the steady atmospheric response interannual anomaly on the “longer period” during El Niño developing events. It is found that the quick and slow establishments of atmospheric response are closely related to the propagation mode and cumulative contributions of MJO events, which can be summarized as Fig. 13. Specifically, in the quick response years, MJO events can propagate to the eastern Pacific because of the associated atmospheric and oceanic states. The pronounced eastward-propagating mode of MJO during the critical period leads to the cumulative negative anomalies of OLR and SLP there and makes a positive contribution to the quick establishment of atmospheric response interannual anomaly accompany with Bjerknes feedback for a weaker El Niño SSTA. However, in the slow response years, the eastward-propagation of MJO events is mainly restricted in the Western Pacific and cannot pass over the dateline. Accordingly, despite a stronger positive SSTA in the tropical eastern Pacific during the critical period, the steady atmospheric response establishes later in the slow response years only with the local Bjerknes feedback.

Although previous works illustrated the relationship between MJO and ENSO, only less attention were paid to the time-lag between the establishment of steady atmospheric response and the steady El Niño SSTA during El Niño developing years, especially about the contribution from MJO. Our results first revealed that the propagation and cumulative effects of MJO dominant the onset date of the steady atmospheric response interannual anomaly over the tropical Pacific, accompany with the local Bjerknes feedback. This research may improve our understanding of inconsistent atmospheric responses in different El Niño developing years. On the other hand, it may make sense to improve the subseasonal prediction ability of ENSO (Xue et al., 2005).

However, the present study only investigated the canonical El Niño events (Larkin and Harrison 2005; Takahashi 2011), the different atmospheric responses and the MJO behaviors in La Niña events should be further discussed. It's also interesting to compare different MJO contribution to Eastern-Pacific (EP) El Niño and Central-Pacific (CP) El Niño (Larkin and Harrison 2005) with the difference between quick and slow response events. In addition, we used an intermediate atmospheric model in the simulations. Although this model is capable to simulate MJO in the previous work, a complex air-sea coupling model is necessary for further research (Jiang et al. 2019; Chen et al. 2020). Limited by the length of daily reanalysis data, the number of the total canonical El Niño samples is 8 during 1982-2016 (<http://www.bom.gov.au/climate/enso/outlook/#tabs=ENSO-Outlook-history>). Six of them have significant consecutive atmospheric response. We classified these El Niño events based on the distinctive quick or slow atmospheric response, giving a 95% confidence level of T-test. Although the sample sizes of the two types of El Niño are not abundant, they can promote the further understanding of MJO effects on ENSO (Kapur et al. 2012; Lybarger and Stan 2018) and improve the long-term forecasting capabilities of ENSO (Shi et al. 2011; Duan and Hu 2016). Similar small sample events such as the discovery of Indian Ocean Dipole (IOD), which was summarized from only 6 events (Saji et al. 1999). However, their results were confirmed again with the extension of the data (Sreekala et al. 2018). We will update the present research if a longer dataset is available and try to use the model data to extend the sample size in the next work.

Declarations

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Haibo Hu and Rongrong Wang contribute equally as first authors.

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Figures

Figure 1

The Sea Surface Temperature Anomaly (SSTA) (shadings; unit: °C) when the monthly Niño3.4 first exceeds +0.8 in 5 consecutive months and the Sea Level Pressure Anomaly (SLPA) (dashed black contours, from -2.2 to -0.4 by 0.2, unit: hPa) 3 months later in two El Niño cases: (a) 1982/83 and (b) 1997/98. The red rectangles (5°S–5°N, 170°–120°W) represent the calculation range of Niño3.4 index.

Figure 2

The composite evolutions of SOI and Niño3.4 indexes. X-axis represents the El Niño developing year (0) and El Niño mature year (1). The blue, red, and black lines represent the quick response years, the slow response years and the average of the all the El Niño years, respectively. The dashed lines are the values of SOI, and the solid lines represent the Niño3.4. The atmospheric threshold of steady El Niño (SOI = -4.9, yellow line) and the oceanic threshold of steady El Niño (Niño3.4 = +0.8, purple line) are showed too. The star and circle represent the time when Niño3.4 exceeds the oceanic threshold (red star) and the SOI exceeds the atmospheric threshold (red solid circle) in the slow response years; and in the quick response years when the Niño3.4 exceeds the oceanic threshold (blue star) and the SOI exceeds the atmospheric threshold (blue solid circle). The critical period is defined as the time interval between the red (blue) star and red (blue) solid circle in the slow (quick) response years. The critical period of slow response years is the summation of period 1, critical period of quick response year, and period 2.

Figure 3

The different precipitation rate and horizontal wind between the quick and slow response years during (a) Period 1 and (b) Period 2 in Fig. 2. The differences of precipitation rate (shadings, unit: mm/day) and wind (vector arrows, unit: m/s) at 925hPa are plotted, using fields of quick response years minus slow response years. All shading areas and arrows exceed 95% significance level.

Figure 4

MJO composites of 30–90-day OLRA (shadings, unit: W/m^2) during the critical periods of quick and slow response years. (a) (c) (e) (g) Quick response years. (b) (d) (f) (h) Slow response years. All shadings exceed 95% significance level. Phase1–Phase2 represents the averaged pattern of MJO Phase1 and Phase2. P3–P4, P5–P6, P7–P8 are similar to it. The blue shadings represent the positive anomalies of MJO convection. The red solid lines denote the propagation track of MJO events.

Figure 5

The distributions of the MJO RMM phases in the critical periods of (a) quick response years and (b) slow response years. The yellow and green dots represent the first and last dekad of the critical period, respectively. The critical period lasts for 6 dekads (60 days) and 18 dekads (180 days) in the quick and slow response years. For comparison, the 6th dekad is also marked as purple dots in Fig. 5b.

Figure 6

Different RMM phase–vertical level patterns between quick and slow response years (shadings, quick response years minus slow response years) for the domains of eastern MC ($10^{\circ}S-10^{\circ}N, 120^{\circ}-130^{\circ}E$) in the critical period. The contours are the climatology condition of El Niño events during 1982–2016. The X-axis are the corresponding location of RMM MJO phases. (a) Divergence (shadings, unit: $1/s$; contour from -3.5×10^{-4} to 3.5×10^{-4} by 0.5×10^{-4} . The red line is the height of the boundary layer, which obtained from ERA-Interim reanalysis data. (b) Vertical velocity. (shadings, unit: m/s ; contour from -0.028 to 0.02 by 0.004). (c) Specific humidity (shadings, unit: g/kg ; contour from -4×10^{-4} to 6×10^{-4} by 10^{-4}). All shadings areas exceed 95% significance level.

Figure 7

The SSTA (shadings, unit: °C), vertical shear of zonal wind (U200 minus U850, contours, every 2, unit: m/s) and MJO initial disturbance in the boreal summer of the (a) quick and (b) slow response years. The mean condition of El Niño events during 1982–2016 is removed. The first day when RMM index exceeds 1 is defined as the initial day of MJO and the associated amplitude is recognized as the magnitude of MJO initial disturbance. Positive (negative) values are contoured by black solid (dashed) line, and the zero line is doubled. (c) The difference between the quick and slow response years (quick response years minus slow response years). Red (blue) circle is the position of MJO initial disturbance in the slow (quick) response years, and the size of the circle represents the strength of the initial disturbance, which is calculated from observation.

Figure 8

The control runs of (a) quick response years and (b) slow response years. Precipitation rate (shadings; unit: mm/day) and geopotential height anomalies at 850hPa (dashed blue and red contours, from -200 to 100 every 15, unit: m2s-2) relative to the initial values are

plotted. Positive (negative) values are denoted by solid (dashed) lines. Specific mean states of SSTA, U200 minus U850, and MJO initial disturbances in Fig. 7a and Fig. 7b are input to force the runs of quick and slow response years, respectively.

Figure 9

The different simulation patterns (quick response years minus slow response years) of Table 2. Precipitation rate (shadings; unit: mm/day) and 850hPa geopotential anomalies relative to the initial values (dashed black contours, every 15, unit: m^2s^{-2}) from day 1 to day 13 are plotted. Positive (negative) values of geopotential anomalies are denoted by solid (dashed) lines, and red (blue) shading represents the positive (negative) precipitation anomalies. The effects of difference of the initial disturbance of MJO, the vertical shear of zonal wind, the SSTA, and the moisture feedback within the PBL between quick and slow response years are showed relative to the CTL Diff (the difference between quick and slow response years in Fig.8).

Figure 10

The dekadly mean 30–90-day filtered OLRA (shadings, unit: W/m²), 30–90-day filtered SLPA (black dashed lines; contours, unit: hPa; contour from -0.7 to -8 by 0.5) and unfiltered SLPA (red dashed lines; contours, unit: hPa; contour from -1 to -16 by 1) during the critical period (60 days) of quick response years. The position of the Darwin (130°59'E, 12°20'S) and the Tahiti (148°05'W, 17°53'S) are marked as black and red stars, respectively.

Figure 11

As Fig. 10, but during the critical period (180 days) of slow response years.

Figure 12

The left and right column show the cumulative contribution of 30–90-day filtered signals in the quick and slow response years, respectively. (a) Cumulative contribution of 30–90-day filtered OLRA (contours, units: W/m²) of the critical period in the quick response years. (c) The variance of negative OLRA of the critical period in the quick response years. (e) Cumulative contribution of 30–90-day SLPA (contours, units: Pa) of the critical period in the quick response years. (g) Averaged unfiltered SLP anomalies (contours, units: Pa) of the critical period in the quick response years. (b), (d), (f), (h) are similar as the left column but in the slow response years. The subseasonal cumulative contributions are defined as the time-average 30–90-day filtered values during the critical periods of quick (Fig. 10) and slow (Fig. 11) response years, respectively.

Figure 13

The schematic diagram of how the propagating MJO influences atmospheric response interannual anomaly during El Niño developing years. The left red oval of each row denotes the SSTA when Niño3.4 first reaches the El Niño condition. The red oval in right side represents the SSTA when the atmospheric response anomaly sets up. The gray vectors are airflow of vertical Walker circulation of z-plane, and the red dashed line represents the specific processes. (a) In the quick response years, MJO events propagate further to 120°W (process 1). The subseasonal MJO convection signals lead to cumulative positive contributions to the atmospheric response there, accompany with the local Bjerknes feedback (process 2), resulting in the quick atmospheric response (b) In the slow response years, the propagation of MJO is weak and restricted in the western Pacific (process 3). Hence, the subseasonal signals barely have contribution on the establishment of atmospheric response interannual anomaly (process 4).

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