Tropical–Extratropical Interactions Associated with East Asian Cold Air Outbreaks. Part II: Intraseasonal Variation

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ABSTRACT

Intraseasonal variations of East Asian cold air outbreaks (CAOs) in relation to the tropical atmosphere during 34 winters (DJF) are investigated. This study is a continuation of Part I, which discussed the interannual variability of East Asian CAOs. Two types of quantitative East Asian CAOs, western and eastern CAOs, are examined. Their variations are identified by the zonal integration of equatorward flux of cold air mass (CAM) below 280 K at 45°N over 90°–135°E and 135°E–180°. A day-lagged regression analysis reveals that peaks of intraseasonal western and eastern CAO events are preconditioned by large-scale tropical convection anomalies resembling particular phases of the Madden–Julian oscillation (MJO). Western CAO events tend to occur when the convective phase of the MJO crosses over the Maritime Continent. In contrast, eastern CAO events are triggered by the MJO over the western Pacific. Observations of MJO-related atmospheric anomalies indicate the important roles of poleward Rossby wave trains in affecting extratropical East Asian CAOs. The barotropic Rossby waves develop negative geopotential height anomalies in mid-latitude East Asia, which then induce a low-level equatorward cold airflow. Several experiments in an atmospheric model using prescribed MJO-like heating anomalies demonstrate that the Maritime Continent MJO and the western Pacific MJO clearly affect the equatorward CAM flux over the western and eastern CAO regions, respectively. Compared with the western CAO, the eastern CAO shows a more robust response to the MJO because of stronger wave activity during the western Pacific MJO.

1. Introduction

Cold air outbreak (CAO) events frequently occur during the winter season and cause significant damage to societies. CAO events are characterized by the equatorward flow of a polar cold air mass (CAM) that leads to sharp temperature drops, strong winds, and sometimes heavy precipitation (Chang et al. 1979; Chang and Lau 1982; Zhang et al. 1997). CAOs in East Asia have attracted a great deal of research, since the CAOs within the East Asian winter monsoon are the most vigorous CAOs as a result of the huge mountainous region of the

Tibetan Plateau and the large thermal contrast between the Eurasian continent and the Pacific Ocean (Zhang et al. 1997; Garreaud 2001; Wang and Chen 2014).

A recent study by Iwasaki et al. (2014) proposed a convenient method to diagnose the geographical distributions of polar CAM streams from their generation to disappearance. By defining the airmass conservation below a designated threshold potential temperature ($\theta_T = 280$ K), two distinct equatorward CAM streams were found in boreal winter, namely, the East Asian stream and the North American stream (Iwasaki et al. 2014). These streams indicate the ensemble effect of intermittent CAOs in the respective regions, which are attributed to the downstream development of synoptic disturbances from the high mountains under the control of quasi-stationary ultralong waves. The variability of hemispheric-scale CAOs can be understood with a conceptual charge and discharge model (Kanno et al.

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2015). Some regional analyses of CAOs in East Asia have been conducted. [Shoji et al. \(2014\)](#) investigated the synoptic evolution of East Asian CAOs using quantitative CAO indices. Their interannual variability was discussed in a recent study by [Abdillah et al. \(2017\)](#). Basically, the East Asian CAOs are driven geostrophically by an eastward pressure gradient force of the Siberian high and the Aleutian low. The variability of East Asian CAOs on various time scales is known to be greatly affected by various extratropical phenomena, such as atmospheric blockings and upper-tropospheric wave trains across the Eurasian continent ([Lau and Lau 1984](#); [Takaya and Nakamura 2005a,b](#); [Park et al. 2014](#)), the Arctic Oscillation ([Gong et al. 2001](#); [Park et al. 2011](#)), the North Pacific Oscillation–west Pacific pattern ([Linkin and Nigam 2008](#); [Abdillah et al. 2017](#)), and the Pacific–North American pattern ([Abdillah et al. 2017](#)).

The East Asian equatorward stream was found to exhibit two major patterns that correspond to the two CAO types: the western CAO (W-CAO) and the eastern CAO (E-CAO) ([Shoji et al. 2014](#); [Abdillah et al. 2017](#)). These two CAOs are greatly influenced by the stationary ultralong waves. W-CAO is characterized by the developing Siberian high and Okhotsk low. The widespread cooling as a result of W-CAO is seen in most parts of continental East Asia, and its anomalous equatorward flow can reach Southeast Asian countries. This type of CAO is closely related to the typical East Asian cold surge investigated in many previous studies ([Chang et al. 1979](#); [Zhang et al. 1997](#); [Compo et al. 1999](#)). On the other side, E-CAO occurs over the northwestern Pacific Ocean. It is driven by the strengthened Aleutian low and anomalous high over eastern Siberia. Its direct impact is located mostly over northern and eastern Japan, but E-CAO has other important implications. E-CAO is accompanied by a large warming over western North America (i.e., Alaska and vicinity) because of the severe loss of cold air mass over there ([Abdillah et al. 2017](#)). The loss of cold air mass in the eastern flank of the Aleutian low is a compensation for CAOs in the western flank. Poleward warm advection from the south also contributes to the warming. Therefore, E-CAO greatly influences the polar cold airmass balance over the North Pacific. In Japan, an E-CAO-like pattern was recently observed to be associated with heavy snowfall events over the Kanto region ([Yamazaki et al. 2015](#)). Furthermore, marine CAOs are usually related to the intense energy exchange between the ocean and atmosphere ([Chechin and Pichugin 2015](#)). It potentially affects the ocean circulation and increases the heat flux from the ocean. Several studies have been done on the marine CAOs (e.g., [Kolstad and Bracegirdle 2008](#); [Papritz and Spengler 2017](#)). In [Abdillah et al. \(2017](#),

hereafter [Part I](#)), both W-CAO and E-CAO were shown to be connected to the El Niño–Southern Oscillation (ENSO)-related climate variability, but they have distinct patterns. The former tends to be enhanced during a La Niña event, while the latter is strengthened during an El Niño event. The remote influences of ENSO are linked by anomalous tropical convections over the Maritime Continent and central Pacific, which excite a poleward Rossby wave train to the extratropics.

The Madden–Julian oscillation (MJO) is the most dominant convective mode of tropical intraseasonal variability ([Madden and Julian 1972](#); [Zhang 2005](#)). Deep and large-scale convection associated with the MJO induces atmospheric disturbances in both the tropics (e.g., [Hidayat and Kizu 2010](#)) and the extratropics ([Matthews et al. 2004](#); [Seo and Son 2012](#)). The impact on the extratropics is often explained by a poleward Rossby wave train propagating from the vicinity of the MJO convective anomalies. The Rossby wave is excited by an upper-level divergent flow that meets with a large meridional vorticity gradient in the subtropical jet ([Sardeshmukh and Hoskins 1988](#)). The extratropical circulation anomalies can extract energy from the mean flow by barotropic conversion. [Hsu \(1996\)](#) showed that the MJO influence on extratropical weather is a mixture of the direct Rossby wave response to the tropical heating and interactions between the divergent and rotational flow embedded in the Rossby wave–like disturbance. Previous studies reported the importance of MJO on the precipitation and surface temperature over North America ([Higgins et al. 2000](#); [Lin et al. 2010](#)). Using a composite of MJO phases, [Jeong et al. \(2005\)](#) suggested the importance of the MJO over the Indian Ocean on the development of extreme cold surges over continental East Asia. The Indian Ocean MJO was also shown to increase East Asian precipitation in the coastal areas through changes in the vertical velocity near the entrance of the East Asian jet and a low-level moisture supply ([Jeong et al. 2008](#)). [He et al. \(2011\)](#) investigated the pentad-scale evolution of East Asian weather using regression analysis on the two leading MJO modes. They showed that the MJO over the Maritime Continent enhances winter monsoon circulation and brings cold and dry weather over East Asia, while the Indian Ocean MJO and western Pacific MJO lead to the occurrence of cold weather over western (continental) and eastern (oceanic) East Asia, respectively ([He et al. 2011](#)). However, the influence of the MJO to the aforementioned East Asian CAOs has not been explored.

This study is a part of a series aiming to investigate the relationships between East Asian CAOs and the tropics. We are interested in exploring the interactions at an intraseasonal time scale by using the newly defined

quantitative CAO indices. Here, in Part II, particular attention is paid to the possible influences of the MJO on the CAOs. This article is organized as follows. [Section 2](#) presents the data and methods. [Section 3](#) reveals MJO signatures in association with intraseasonal East Asian CAO events. [Section 4](#) discusses the mechanism of the interactions between the MJO and East Asian CAOs based on both observation and model simulation. [Section 5](#) presents more detailed discussions on the influence of the MJO. A summary and the key findings of this study are presented in [section 6](#).

2. Data and methodology

a. Data

Atmospheric variables are obtained from the Japanese 55-year Reanalysis (JRA-55), which has a 6-h resolution, a $1.25^\circ \times 1.25^\circ$ horizontal resolution, and 37 vertical pressure levels ([Kobayashi et al. 2015](#)). Daily interpolated outgoing longwave radiation (OLR) is provided by NOAA/OAR/ESRL Physical Sciences Division ([Liebmann and Smith 1996](#)) and is utilized to denote the tropical convection; it has a 1-day temporal resolution and a $2.5^\circ \times 2.5^\circ$ horizontal resolution.

b. East Asian CAO index

The definition of the East Asian CAO index follows the equations in [Shoji et al. \(2014\)](#) and [Part I](#), which include cold airmass equations formulated by [Iwasaki et al. \(2014\)](#). The horizontal flux of a cold air mass $\mathbf{F}$ is given by

$$\mathbf{F} \equiv \int_{p(\theta_T)}^{p_s} \mathbf{v} dp, \quad (1)$$

where $\mathbf{v}$ is the horizontal wind field vector; p_s and $p(\theta_T)$ are the pressures at the surface and at a designated potential temperature θ_T , respectively. The variable θ_T is set to be 280 K because this value is a good representation of the global-mean equatorward flow of polar air mass associated with the extratropical direct (ETD) circulation, from the viewpoint of a mass-weighted isentropic zonal mean ([Iwasaki and Mochizuki 2012](#); [Iwasaki et al. 2014](#)). [Shoji et al. \(2014\)](#) defined western and eastern East Asian CAO indices (CAOIs) based on the longitudinal integration of equatorward flux at 45°N from 90° to 135°E :

$$\text{W-CAOI} \equiv \frac{a \cos \phi}{g} \int_{\lambda=90^\circ\text{E}}^{\lambda=135^\circ\text{E}} F_{-v} d\lambda \Big|_{\phi=45^\circ\text{N}} \quad (2)$$

and from 135°E to 180° :

$$\text{E-CAOI} \equiv \frac{a \cos \phi}{g} \int_{\lambda=135^\circ\text{E}}^{\lambda=180^\circ} F_{-v} d\lambda \Big|_{\phi=45^\circ\text{N}}, \quad (3)$$

respectively. The variable F_{-v} denotes the southward component of $\mathbf{F}$ with $\theta_T = 280\text{ K}$. The symbols ϕ , λ , a , and g denote latitude, longitude, Earth radius, and gravitational acceleration, respectively. [Part I](#) showed that W-CAOI and E-CAOI are highly correlated with EOF2 (explained variance of 29%) and EOF1 (31%) patterns of winter-mean East Asian equatorward CAM flux, respectively.

c. Methodology

Interannual variability associated with ENSO is excluded from the time series because the tropical atmosphere, especially convection, is highly correlated with ENSO. In addition, ENSO was found to have remote impacts on the East Asian CAOs ([Part I](#)). Therefore, the influence of ENSO must be removed to emphasize intraseasonal signals in the analysis. We follow the ENSO removal procedure provided by [Wheeler and Hendon \(2004\)](#), except we use the monthly Niño-3 SST index (5°S – 5°N , 150° – 90°W) as a parameter to represent ENSO evolution. The monthly values of the Niño-3 index are converted to a daily basis. Then, a linear regression is calculated from the daily field data at each grid point, separately for each month, which results in 12 regression parameters. The monthly regression parameters are interpolated to a daily basis to yield a 365-day seasonally dependent relationship. Finally, the ENSO influence is diminished by subtracting the regression relationship of each field at each grid point.

We perform a time-lagged correlation and regression among CAOIs and field variables. In this approach, the atmospheric anomalies that precede or follow the CAO event can be determined objectively. The analysis period covers 34 winters from 1979 to 2012, in which the 1979 winter denotes 1979/80 winter. In the lagged analysis, the days of the predictand are fixed at 1 December–28 February, and the days of the predictor depend on the lags that could be in November–March. A preliminary analysis using unfiltered daily data shows that there are potential precursors of a CAO event from the tropics resembling MJO-like patterns. To focus our attention on the interaction between the MJO and CAOs, the datasets are temporally bandpass filtered (BPF) during the MJO period (30–80 days) using a Lanczos filter ([Duchon 1979](#)) from 1 November to 31 March with 320 daily weights for each winter. A sensitivity test for choosing the appropriate band shows the highest connectivity between CAOIs, and tropical OLR is achieved using a 30–80-day BPF dataset (see

appendix A). The minimum period of 30 days neglects any high-frequency waves over the tropics and extratropics. This band also excludes low-frequency variation related to seasonality.

In section 4, we reconstruct an MJO cycle based on a combined lagged analysis from two leading EOF patterns of tropical OLR. This approach enables us to determine the timing of intraseasonal CAO events relative to an MJO cycle and to explain its mechanism. Furthermore, we carry out several numerical experiments using a linearized GCM to demonstrate the MJO influence. Full details on the MJO cycle and model configuration are shown in section 4.

A significance test for the correlation coefficient is calculated based on the two-sided Student's t test at 95% and 99% confidence levels. The time series at each grid point must be decorrelated first to obtain the effective number of degrees of freedom n . As in Livezey and Chen (1983), the autoregressive properties of both time series A and B are taken into account:

$$\tau = 1 + 2 \sum_{i=1}^N C_A(i) C_B(i),$$

where N and $C(i)$ denote the number of samples and the autocorrelation at lag i , respectively. The number of degrees of freedom is then estimated by

$$n = \frac{N}{\tau}.$$

Then, the statistical significance is evaluated locally at each grid point and for each time frame.

3. Evolution characteristics of CAO indices

Day-to-day variations of unfiltered and 30–80-day filtered CAO indices are shown in Fig. 1a. The time interval between two adjacent peaks of an unfiltered CAO is usually a few days to one week (Shoji et al. 2014). The autocorrelation feature of an unfiltered CAO exhibits a clear single peak with a high positive correlation that persisted from approximately day -2 to day $+2$ in both W-CAOI and E-CAOI (Fig. 1b). This highlights the time scale of intermittent CAO events. The synoptic conditions during the W-CAO event and E-CAO event suggest the relative importance of the Siberian high and Aleutian low, respectively (Shoji et al. 2014), which also appears in their interannual variability (Part I). The W-CAO event appears to persist in a shorter time than the E-CAO (Fig. 1b), which is possibly related to the shorter lifetime of the anomalous Siberian high (because of the rapid southward progression)

FIG. 1. (a) Time series of unfiltered and filtered W-CAO indices in 1989 winter. Autocorrelation properties of (b) unfiltered and (c) filtered CAO indices from all 34 winters.

compared with the stationary Aleutian low (Shoji et al. 2014). However, the peak-to-peak time interval of a 30–80-day filtered CAO is several weeks (Fig. 1a). The autocorrelation feature indicates that the filtered CAO lasts long and has significant periodicity (Fig. 1c). Both filtered W-CAOI and E-CAOI have consistent autocorrelation features. There are two minimum peaks at lags near 20–25 days that exhibit a -0.8 correlation coefficient. The variability of W-CAOI and E-CAOI at this time scale appears to be affected by intraseasonal variations associated with the MJO, as will be discussed in the following paragraphs.

4. Relationships between the MJO and East Asian CAOs

a. Evolution of tropical atmosphere during a W-CAO event

The lagged correlations/regressions with the filtered CAOs are used to capture the atmospheric evolution associated with the intraseasonal CAO events. In the

FIG. 2. Lagged correlations between tropical OLR anomalies (averaged at 15°S–15°N) with (a) W-CAOI and (b) E-CAOI shown in longitude–time diagrams (0.05 contour interval). Horizontal lines denote day 0. Stippled areas are significant at the 95% confidence level.

extratropics, the atmospheric anomaly patterns during W-CAO and E-CAO events are comparable with those documented in Shoji et al. (2014), except with more long-lasting signatures (not shown). Figure 2a shows a longitude–time diagram of the lagged correlation between tropical OLR (15°S–15°N) anomalies and the W-CAOI from day –21 to day +21. In general, day 0 corresponds to occasions when the filtered W-CAOI is a maximum. The significant negative correlation leading day 0 suggests a potential precursor to the intraseasonal W-CAO event (Fig. 2a). It is characterized by a peak of the negative OLR anomalies over the western Maritime Continent (105°E) approximately one week before the event. The negative correlation propagates eastward crossing the Maritime Continent near day 0 and arrives at the western Pacific afterward. Approximately two weeks after the event, a significant positive OLR anomaly develops (Fig. 2a). This is related to the characteristics of the intraseasonal W-CAOI, which has a large periodicity (Fig. 1c). Therefore, the positive OLR anomaly in Fig. 2a leads to a minimum of equatorward flow in the W-CAO region.

Figure 3 presents the temporal evolution of the spatial feature of OLR and 850-hPa wind field anomalies associated with a W-CAO event. We focus on the evolution of the W-CAOI maximum during the timeline (from approximately day –10 to day +10). The field anomalies shown here are constructed from a lagged regression and indicate the values according to one standard deviation of the W-CAOI. In general, the W-CAO event is accompanied by a large-scale feature of the organized tropical convections (Fig. 3). It is

characterized by a wide area of negative OLR anomalies over the Maritime Continent whose signals are more apparent before the event at day –10 and day –5 (Figs. 3a,b). At day 0, a strong equatorward outflow from midlatitude East Asia is observed (Fig. 3c), indicating the peak of the W-CAO event. Following the W-CAO event at days +5 and +10, the large-scale negative OLR anomaly persists and propagates eastward and the low-level northerly wind dominates over the subtropical and tropical regions (Figs. 3d,e). These findings are consistent with the result of Shoji et al. (2014), who showed that the anomalous low-level northerly flux associated with a W-CAO event can penetrate to the low-latitude regions. Here, we observe a development of significant negative OLR anomalies over the northern Maritime Continent (i.e., the Philippines and vicinity) after day 0 (Figs. 3d,e). This postevent signature possibly indicates the response of precipitation to the anomalous equatorward flow. Several past studies have documented the influence of an East Asian cold surge on an increase in precipitation over the South China Sea and Philippines (Compo et al. 1999; Chang et al. 1979, 2005; Chen et al. 2015a).

b. Evolution of tropical atmosphere during an E-CAO event

The signature of tropical OLR anomalies correlated with the E-CAOI is shown in Fig. 2b. Leading day 0, the OLR anomaly resembles a clear dipole pattern with the negative anomaly over the western Pacific and the positive anomaly over the Indian Ocean. The OLR anomalies propagate eastward, similar to the characteristics

FIG. 3. Lagged regressions between OLR (color shaded; 0.5 W m^{-2} contour interval) and 850-hPa wind field anomalies (vectors; m s^{-1}) with W-CAOI. Temporal evolution at (a) day -10 , (b) day -5 , (c) day 0 , (d) day $+5$, and (e) day $+10$. Purple contours and vectors denote significant correlations at the 95% confidence level.

observed during a W-CAO event (Figs. 2a,b). The correlations in the E-CAOI are more significant than those in the W-CAOI. The spatial variation of OLR and 850-hPa wind anomalies is shown in Fig. 4. An E-CAO event is clearly led by the large-scale dipole pattern over the western Pacific and Indian Oceans (Figs. 4a,b). This dipole pattern resembles an MJO wet phase over the western Pacific, and it could be a precondition for an E-CAO event. At day 0, the CAO occurs over the northwestern Pacific Ocean. Following the event, the

FIG. 4. As in Fig. 3, but for regressions with E-CAOI.

negative OLR anomaly weakens over the equatorial Pacific (Figs. 4d,e), while the positive OLR anomaly dominates from the Indian Ocean to the Maritime Continent. In contrast to those in a W-CAO pattern, the postevent feature of an E-CAO does not exhibit a penetration of low-level northerly flow anomaly to the tropical regions (Figs. 4d,e). The large portion of the extratropical mass flux is stirred eastward (Figs. 4c,d) because of the development of the Aleutian low (Shoji et al. 2014; Part I).

c. Observational evidence revealed in the MJO life cycle

Figures 2–4 indicate possible influences of the MJO on the East Asian CAOs. Further analysis with a reference

FIG. 5. Two leading EOF modes extracted from tropical OLR anomaly fields (20°S – 20°N , 50°E – 160°W). OLR (a) EOF1 and (b) EOF2 patterns. Contours denote the OLR anomalies obtained from simultaneous regressions with the corresponding PC indices (2 W m^{-2} contour interval). (c) The autocorrelation of the PC1 index (solid line), and the lagged correlation between the PC2 and PC1 index (dashed line).

from MJO events is needed to investigate the interactions in detail. Previous studies often used simultaneous composites of MJO phases to study the impact of the MJO (e.g., Vecchi and Bond 2004; Jeong et al. 2005). Considering the forcing from the tropics takes time to affect the midlatitude weather, it may be difficult to see the evolution and direct impact of the MJO using the simultaneous composite.

In this study, we perform an analysis based on a combination of lagged regressions/correlations with two MJO indices. First, the MJO indices are created from the two leading EOF modes of the filtered OLR anomaly in the tropics (20°S – 20°N , 30°E – 120°W). The OLR EOF1 shows a clear dipole pattern of tropical convection over the western Pacific and Indian Oceans, whereas the OLR EOF2 shows a broad area of

convection anomaly centered over the Maritime Continent accompanied by a reversed anomaly center over the central Pacific (Figs. 5a,b). Their variance contributions are quite balanced, 14.52% for EOF1 and 13.54% for EOF2. These OLR EOFs simply represent two different phases of the eastward-propagating MJO. The autocorrelation of the first principal component (PC1) index and the lagged correlation between PC1 and PC2 indices yield a sequence of $-\text{EOF2}$, $+\text{EOF1}$, $+\text{EOF2}$, and $-\text{EOF1}$ that appears to be separated by approximately 12 days (Fig. 5c). We infer that a complete MJO life cycle has a nominal length of 48 days, which is identical to previous studies (e.g., Matthews et al. 2004). Progression through the MJO cycle is determined in terms of a phase angle varied from 0° ($t = 0$ days) to 360° ($t = 48$ days). The 0° phase corresponds to

FIG. 6. Temporal evolution of an MJO cycle calculated from the combined lagged analysis and OLR PC indices. Left and right axes denote the day of the MJO cycle and the corresponding eight phases of the MJO, respectively. (a) The regressions (contour lines; 2 W m^{-2} contour interval) and correlation coefficients (color shaded; 0.2) of OLR. The variation of correlation coefficients of (b) W-CAOI and (c) E-CAOI. Thin and thick solid lines are significant at 95% and 99% confidence levels, respectively.

occasions when PC2 is a maximum, and simultaneous regression maps for this phase are calculated using PC2 as the dependent variable. The 180° phase, when PC2 is a minimum, corresponds to $t = 24$ days; therefore, $-PC2$ is used to calculate the simultaneous regression. The 90° ($t = 12$ days) and 270° ($t = 36$ days) phases are denoted by simultaneous regressions using $-PC1$ and $+PC1$, respectively. Because of the periodicity in the analysis technique, $t = 48$ days is equivalent to $t = 0$. Regression maps at intermediate phases are constructed by linear combinations of lagged regression maps based on PC1 and PC2. For example, to determine regression maps at $t = 1$ – 11 days, lagged regressions maps of PC2 at lag from $+1$ to $+11$ days are combined with the maps of $-PC1$ at lag from -11 to -1 days, respectively, with certain weights. The weights are linearly distributed from $11/12$ to $1/12$ for PC2 and from $1/12$ to $11/12$ for $-PC1$. Therefore, the regression maps at $t = 6$ days (45° phase) are obtained by combining lagged regression maps of PC2 at lag $+6$ days and $-PC1$ at lag -6 days with equal weights of $6/12$. This also applies for lagged correlations. To calculate the statistical significance level, the numbers of degrees of freedom are combined using the same method. This approach enables us to depict a day-to-day evolution of the MJO life cycle.

Figure 6a shows a tropical OLR evolution of an MJO cycle in a longitude–time diagram. The negative OLR anomalies clearly propagate from the central Indian Ocean at $t = 0$ days to central Pacific at $t = 48$ days, while the positive OLR anomalies appear over the central Indian Ocean at $t = 24$ days. Figures 6b and 6c show the

evolution of the W-CAOI and E-CAOI, respectively. The maximum of the W-CAOI is approximately at $t = 27$ days, corresponding to MJO phase 5 or the late phase of the Maritime Continent MJO. Additionally, the maximum of the E-CAOI is seen at $t = 39$ days, corresponding to MJO phase 7 after the MJO crossing over the western Pacific. Because of the periodicity of the MJO, the minima of the W-CAOI and E-CAOI are observed at $t = 3$ days and $t = 15$ days, respectively. The minima of CAOIs differ by 24 days with their maxima. This time scale is nearly consistent with the time scale of the autocorrelations of CAO indices (Fig. 1c).

The remote impacts of the MJO on the extratropics were often explained by poleward Rossby wave trains, which are induced by the upper-level divergence/convergence of tropical convective anomalies (Matthews et al. 2004; Lin et al. 2010; Seo and Son 2012). Here, we explore the possibility of Rossby waves influencing the East Asian CAOs. Figure 7 shows atmospheric circulation anomalies at $t = 21$, 24, and 27 days represented by a 250-hPa geopotential height and its wave activity flux (Takaya and Nakamura 2001), a 500-hPa geopotential height, and the CAM amount and its flux below 280 K, as well as 850-hPa wind and temperature fields. The convection over the Maritime Continent induces an upper-level anomalous anticyclone over South Asia (Fig. 7a), which subsequently develops upper-level anomalous cyclonic circulation over the midlatitude East Asian coast. The extratropical response is quite barotropic. The development of a 500-hPa East Asian trough is clear and reaches its peak at $t = 27$ day (Fig. 7b). The trough

FIG. 7. Atmospheric anomalies at (left) $t = 21$, (center) $t = 24$, and (right) $t = 27$ days of the MJO cycle. (a) The 250-hPa geopotential height (shaded; 5 gpm contour interval) and wave activity flux $\geq 0.05 \text{ m}^2 \text{ s}^{-2}$ (vectors), (b) The 500-hPa geopotential height (shaded; 4 gpm contour interval), (c) CAM amount (shaded; 3 hPa contour interval) and its flux (vectors; hPa m s^{-1}) below 280 K, and (d) 850-hPa temperature (shaded; 0.2 K contour interval) and wind (vectors; m s^{-1}). Purple contours and vectors in (c) and (d) denote significant correlations at the 95% confidence level. Green contours indicate OLR anomalies with a 4 W m^{-2} contour interval (zero values omitted).

triggers more equatorward cold airflow in the west of the cyclone, which is the location of W-CAO (Fig. 7c). The evolution of low-level monsoon circulation is also observed (Fig. 7d). The anomalous 850-hPa northerly wind and cold temperatures extend from midlatitude East Asia to the tropical region. On the other hand, Fig. 8 exhibits atmospheric anomalies at $t = 33$, 36, and 39 days. The Rossby wave trains, which are induced by the convective anomalies over the western Pacific and Indian Oceans, are also evident (Fig. 8a) and even stronger than those in Fig. 7a. The center of the extratropical cyclonic anomaly resides over the North Pacific and overlaps the location of the Aleutian low (Fig. 8b); this induces a stronger equatorward cold airflow over the E-CAO region (Fig. 8c) and causes an

intense overturning circulation over the North Pacific Ocean (Fig. 8d).

Figure 9 shows the temporal evolution of the East Asian meridional cold airflow and 500-hPa geopotential height anomalies in an MJO cycle. This figure clearly shows that the variation of East Asian CAOs is strongly guided by the center of a midlatitude trough and ridge. The equatorward flow becomes stronger when the trough is apparent, but it becomes weaker when the ridge prevails. The equatorward flow anomaly starts to develop at approximately $t = 15$ days (phase 3) over the westernmost part. It then propagates eastward following the center of the trough developed by the MJO. The response over the E-CAO region is more robust than it is over the W-CAO region because the anomaly

FIG. 8. As in Fig. 7, but for (left) $t = 33$, (center) $t = 36$, and (right) $t = 39$ days.

height is larger over the eastern part than it is over the western part.

d. Linear model experiments

To demonstrate the impact of the MJO, we perform numerical experiments using a linear baroclinic model (LBM; Watanabe and Kimoto 2000) using prescribed MJO-like thermal forcing. The model configurations are generally similar with those in Part I, except the horizontal pattern of heating is approximated from MJO-like convection anomalies rather than idealized Gaussian patterns. Table 1 shows a list of experiments that can be divided into two groups: MJO-A and MJO-B experiments. Thermal forcing in MJO-A is estimated from tropical OLR anomalies (20°S–20°N, 40°E–120°W) when the MJO crosses over the Maritime Continent (–EOF2 pattern), whereas MJO-B mimics the heating pattern when the MJO crosses over the western Pacific

(+EOF1 pattern). OLR anomalies < 2 and $> -2 \text{ W m}^{-2}$, as well as minor convective centers, are excluded from the model input. The minimum or maximum thermal forcing is set to be 4 or -4 K day^{-1} . Furthermore, we perform sensitivity experiments to reveal the relative importance of different heating centers. The results are expected to provide answers for why the extratropical response in the E-CAO is stronger than in the W-CAO. The MJO-A_MC and MJO-A_CP experiments simulate atmospheric responses based on the Maritime Continent heating and the central Pacific cooling, respectively, while the MJO-B_WP and MJO-B_IO experiments simulate the responses based on the western Pacific heating and the Indian Ocean cooling, respectively.

Figure 10a shows the linear response of atmospheric circulation in the MJO_A experiments two weeks after the integration. The wave trains that are evident over

FIG. 9. Longitude–time diagram of anomalous meridional components of the 280-K CAM flux at 45°N (color shaded; 20 hPa m s^{−1} contour interval) and 500-hPa geopotential height averaged over 30°–40°N (black contours; 2 gpm contour interval). Purple lines enclose the significant meridional CAM flux at the 95% confidence level.

East Asia (Fig. 10a, left) deepen the East Asian trough, induce equatorward flow over the W-CAO region, and bring more cold air mass to East Asia (Fig. 10a, right). The sensitivity experiments of the MJO-A_MC and MJO-A_CP suggest that the East Asian responses are strongly controlled by the Maritime Continent heating (Figs. 10b,c). The response because of the central Pacific cooling is significantly weaker and even indicates an anticyclone over the North Pacific (Fig. 10c). In fact, the lagged regression of OLR anomalies onto the W-CAOI does not exhibit a strong signal over the central Pacific (Fig. 2). The importance of Maritime Continent heating to W-CAO was also mentioned in Part I.

Figure 11a shows the atmospheric responses in the MJO-B experiment. The resulting Rossby wave trains

appear from South Asia to North America, develop a cyclonic anomaly over the North Pacific, and trigger an equatorward cold airflow over the E-CAO region. The Rossby waves are substantially stronger than those in the MJO-A experiment, which is consistent with the observations. The sensitivity experiments of MJO-B_WP and MJO-B_IO clarify the important roles of both western Pacific heating and Indian Ocean cooling on the development of E-CAO (Figs. 11b,c). The response in E-CAO is greater than that in W-CAO because of the collaborative effects of in-phase Rossby waves induced by western Pacific heating and Indian Ocean cooling.

5. Discussion

a. Tropical atmosphere contributions to the East Asian CAOs

In summary, as with Part I, the tropics have a significant effect on the East Asian CAOs in two frequency bands: interannual and intraseasonal time scales. The seasonal mean of a CAO is affected by ENSO (Part I), whereas the intraseasonal CAO event is affected by the MJO in a 30–80-day band period. Nevertheless, we find no notable interactions between the tropics and short-term CAO events in the high-frequency band (<30 days), as indicated by the very small correlations (see appendix A). The tropical waves in this period (<30 days) are not strong enough to transfer a remote impact to the extratropics. It was documented that the shorter the time scale of the anomalous tropical heating, the more equatorially trapped the atmospheric response (Trenberth et al. 1998). Therefore, the timing of short-term CAO events is largely controlled by the internal extratropical dynamics.

b. Possible influence of the MJO on the occurrence probability of short-term CAO events

Despite the inability of the MJO to predict the timing of short-term (high frequency) CAO events, the MJO may play a significant role on the probability of its occurrence. Here, we create a composite analysis of the short-term CAO events in eight MJO phases. The CAO

TABLE 1. Name and description of experiments carried out in the linear model simulations.

Experiment	Description of heating source
MJO-A	Approximated from simultaneous regressions between −PC2 index and tropical OLR.
MJO-A_MC	As in MJO-A expt, but heating only over the Maritime Continent.
MJO-A_CP	As in MJO-A expt, but cooling only over the central Pacific.
MJO-B	Approximated from simultaneous regressions between +PC1 index and tropical OLR.
MJO-B_WP	As in MJO-B expt, but heating only over the western Pacific.
MJO-B_IO	As in MJO-B expt, but cooling only over the Indian Ocean.

Linear responses in MJO–A Experiments

FIG. 10. Northern Hemisphere atmospheric responses two weeks after the integration in (a) MJO–A, (b) MJO–A_MC, and (c) MJO–A_CP experiments denoted by anomalies of (left) a 250-hPa geopotential height (black contours with 10 gpm contour interval) and (right) a CAM (blue–red shading) overlapped with its flux (vectors). Light and dark gray shading denotes the prescribed positive and negative heating anomalies greater than $\pm 1 \text{ K day}^{-1}$ in the tropics.

events are identified based on the local maxima of unfiltered CAO indices; the maxima must be greater than or equal to 1.5 standard deviation. A total of 151 W-CAO events and 130 E-CAO events are captured during the analysis period. We count the number of CAO events that occurred in specific MJO phases and perform a significance test using the Monte Carlo bootstrap method (Efron and Tibshirani 1993; Li et al. 2016) (see appendix B).

Figures 12a and 12b exhibit the distribution of W-CAO events and E-CAO events, respectively, over an MJO phase–space diagram from Wheeler and Hendon (2004), whose real-time multivariate MJO (RMM) indices are obtained from the Australian Bureau of Meteorology website. The W-CAO events tend to occur after the MJO crossing over the Maritime Continent (Fig. 12a). For a comparison among the MJO phases, the

maximum number of W-CAO events is detected in phase 5 (20.1%) followed by phase 6 (18.0%), whereas the lowest frequency is observed in phases 8 (6.5%) and 1 (7.2%). They are statistically significant at 95% (Fig. 12c). However, the E-CAO events tend to occur after the MJO crossing over the western Pacific, where several extreme events are observed (phases 7 and 8) (Fig. 12b). The significant maximum occurrence is observed in phase 7 with 19% of the total frequency (Fig. 12d), while there is no clear minimum peak. Figures 12e,f show the frequency distributions of CAO events analyzed using the OLR EOF indices. The result is somewhat consistent with the CAO distribution in the RMM indices. The peak of W-CAO occurrence appears in phase 5 (Fig. 12e) and emphasizes the CAO occurrence during the poststage of the Maritime Continent's MJO. Additionally, the peaks of E-CAO occurrence are

Linear responses in MJO-B Experiments

FIG. 11. As in Fig. 10, but for MJO-B, MJO-B_WP, and MJO-B_IO experiments.

observed in phases 8 and 6 (Fig. 12f). The occurrence probability of E-CAO events seems rather sensitive to the definition of MJO indices. The E-CAO occurrences are distributed widely during MJO phases over the western to central Pacific.

The frequency distribution of short-term CAO events suggests that there are potential impacts of the MJO on the probability of CAO occurrence (Fig. 12) because of the favorable background conditions given by particular MJO phases. The MJO induces negative geopotential height anomalies from midlatitude East Asia to the North Pacific (Figs. 7–9), which cause CAOs to occur more often.

c. Relationships to previous work

Several studies have investigated the influences of the MJO on East Asia. Using a simultaneous composite analysis in eight MJO phases, Jeong et al. (2005) concluded that the MJO over the Indian Ocean (phases 2 and 3) tends to cause cold events and cold surges with a more dominant influence at phase 3. It seems at first to be inconsistent with our finding related to the W-CAO

development that appears at phase 5. To validate the consistency, we show the evolution of low-level temperature anomalies for the eight MJO phases (Fig. S2 in the supplemental material). Figure S2 is comparable with Fig. 1 in Jeong et al. (2005). The temperature anomalies associated with the MJO phases shown in their result are similar to our result to a broad extent. Here, the broadest significant cooling over continental East Asia is observed at phase 3 (Fig. S2c), which does not contradict with the finding of Jeong et al. (2005). To get a better view, 850-hPa wind anomalies are shown in Fig. S2, and a 280-K anomalous CAM flux is shown in Fig. S3 of the supplemental material. At phase 3, significant northerly winds are mainly located over inland China (Fig. S2c). The northerly CAM flux toward inland China is observed at phase 3, but its amplitude and its region are rather limited (Fig. S3c). However, phase 5 exhibits strong northerly wind anomalies emanating from 45°N (Figs. S2e and S3e). Significant low-level temperature drops are largely observed near the southern Korea Peninsula, Japan, and southern China (Fig. S2e). Another important signature at phase 5 is that the northerly

FIG. 12. Scatterplot distribution of short-term (a) W-CAO events (blue circles) and (b) E-CAO events (red circles) over the MJO phase-space diagram using RMM indices (Wheeler and Hendon 2004). The size of the circles is proportional to the intensity of the CAO event. Historical MJO tracks are drawn with gray lines. The percentage of CAO events are calculated for each MJO phase (MJO intensity ≥ 0.5) in (c),(d) RMM indices and (e),(f) OLR EOF indices. Asterisks indicate statistical significance at the 95% confidence level using Monte Carlo simulation.

wind and cold temperature anomalies extend to the East China Sea and propagate to the South China Sea and Southeast Asian countries (Figs. S2e,f).

The difference in conclusions from those of Jeong et al. (2005) is related to the difference in CAO definition. One of the cold surge criteria in their study was based on temperature drops over the middle or southern regions of China. Using this definition, most of the cold events are captured at phase 3 (Figs. S2c and S3c). In this study, we define W-CAOI as an equatorward cold air mass flux at 45°N , 90° – 135°E , whose magnitude reaches a maximum at phase 5 (Figs. 6b and S3e). Nevertheless, our temperature anomaly maps are quite consistent, although some minor differences may exist

due to the difference in analysis techniques and the utilization of a time filter. It is very clear that the pathway of anomalous northerlies shifts eastward following the MJO propagation (Figs. 9, S2, and S3). The enhanced equatorward flux that appears at phase 3 extends and propagates eastward at phases 4 and 5. It then gradually evolves into an E-CAO-type flux at phases 6–8. Our result demonstrates that the location of anomalous northerly flow in East Asia is sensitive to the MJO propagation. This northerly cold flux is controlled by the propagation of the East Asian trough, as shown in Fig. 9.

Our result is also consistent with a study by He et al. (2011). They investigated the pentad-scale evolution of East Asian weather using regression analysis with the

two leading MJO modes. They showed that the MJO over the Maritime Continent enhances East Asian winter monsoon circulation and brings cold and dry weather over East Asia, while the Indian Ocean MJO and western Pacific MJO lead to the occurrence of cold weather over the western (continental) and eastern (oceanic) East Asia, respectively (He et al. 2011). In our study, phase 5 in Fig. S2e (3 days after the peak of the Maritime Continent MJO) exhibits strong northerly wind and cold temperature anomalies in East Asia, indicating the enhanced winter monsoon circulation documented in He et al. (2011), while the dipole patterns between the western and eastern parts appear at phases 3 and 7 (Figs. S2c,g), which is also consistent with their result. He et al. (2011) suggested that, to some extent, the enhanced local Hadley circulation associated with the Maritime Continent MJO strengthens the East Asian winter monsoon. This could explain why the northerlies at phase 5 are quite strong near subtropical coastal China and the East China Sea (Fig. S2e) despite the cold air outbreak from midlatitude East Asia being not so strong (Figs. S2e and S3e).

6. Concluding remarks

We have investigated the interactions between tropical atmosphere and East Asian CAOs on an intraseasonal time scale (30–80-day period). Two quantitative East Asian CAOs are examined: W-CAO and E-CAO. Lagged regressions and correlations are used to determine the evolution of CAO events. It is found that the intraseasonal W-CAO and E-CAO events are preconditioned by MJO-related large-scale convections with the signals preceding the CAOs by approximately 1 week. The precursor for W-CAO is the MJO over the Maritime Continent, while the precursor for E-CAO is the MJO over the western Pacific. The remote impact is delivered by poleward Rossby wave trains excited by anomalous tropical convections, which in turn develop negative geopotential height anomalies in midlatitude East Asia and affect the low-level CAM flux. Model simulations using a linearized GCM show that the response in E-CAO is stronger than it is in W-CAO because the Rossby wave trains during the western Pacific MJO are more significant because of the in-phase relationships of circulation patterns excited by anomalous convections over the western Pacific and Indian Oceans.

It is clear that the location of intraseasonal northerly anomalies in East Asia is sensitive to the MJO propagation. The equatorward flow starts to develop at phase 3 over continental East Asia (Jeong et al. 2005), particularly over the western part of the W-CAO region,

which then extends and shifts eastward following the development of the East Asian trough at phases 4 and 5. At phase 5, the W-CAOI reaches its maximum, and the low-level winter circulation in East Asia becomes more active (He et al. 2011). The equatorward flow then gradually evolves into an E-CAO-type flux at phases 6–8. These results imply the complexity of the MJO's influence over East Asia because East Asia covers a vast region elongating from the subtropics to the polar region and from the Tibetan Plateau to the North Pacific, and they require careful interpretation depending on the region of interest.

Furthermore, the MJO has a potential impact on the occurrence probability of short-term or high-frequency CAO events, as shown by the composite analysis. Comparing among the MJO phases, approximately 38% (34%) of short-term W-CAO (E-CAO) events occur at phases 5 and 6 (phases 7 and 8) following the MJO over the Maritime Continent (western Pacific). The MJO develops a favorable background condition in the extratropics that facilitates more frequent CAO events. A more detailed study is needed to investigate the MJO influence on the transient CAOs.

In Part I of our study, we investigated the interannual variability of CAOs and its relationship with tropical climate. Two leading EOF modes of East Asian equatorward flow were identified. Both are consistent with W-CAO and E-CAO. ENSO-related convection anomalies affect the intensity of seasonal-mean East Asian CAO indices. La Niña tends to induce a strong W-CAO winter, and El Niño tends to induce a strong E-CAO winter. Here, in Part II, the intraseasonal CAO events are affected by particular phases of the MJO. The intraseasonal W-CAO is triggered by the MJO over the Maritime Continent and is consistent with Part I according to the location of convection. While the intraseasonal E-CAO is mainly controlled by the MJO over the western Pacific. Based on the numerical experiments in Part I and Part II, the E-CAO flux can be affected by tropical heating over the western to central Pacific. These studies emphasize the importance of tropical convections on medium- and long-range CAO forecasting. Despite the different time scales between the MJO and ENSO, the characteristics of MJO evolution and its teleconnection sometimes depend on ENSO phases (Moon et al. 2011; Feng et al. 2015). A more detailed study on the interactions between CAO–MJO–ENSO is needed to gain a better understanding on the CAO variation.

Several studies have documented that CAOs in East Asia can enhance the tropical precipitation in Southeast Asian countries (Chang et al. 1979; Compo et al. 1999; Chen et al. 2015a). This impact on the precipitation field

is due to an enhanced northerly moisture flux, vertical wind shear, and interactions with the topography. The response is indirect and sensitive to the preconditioning of the tropical environment. Because of this possibility of CAO influence on the tropics, interactions between CAOs and the MJO can be rather complex. The CAOs, especially W-CAO, may give positive feedback to the tropical convection. Although the current study provides some evidence regarding the development of tropical convection following a CAO event (Figs. 3d,e), a more in-depth analysis is needed to draw a solid conclusion about the impact. Some past studies showed that collaborative effects of the MJO and subtropical cold surge were shown to amplify northerly flux and tropical precipitation, which potentially cause heavy rainfall and flood events (Hattori et al. 2011; Lim et al. 2017). Chen et al. (2015b) specifically discussed the positive effect of the East Asian winter monsoon on MJO intensification. However, most of the works evaluated the cold surge/northerly flow over a subtropical or low-latitude region, which is near the tropics. Future studies are needed to explore the possibility of extratropical CAOs influence on tropical precipitation and the MJO.

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APPENDIX A

Bandpass Filter and Connectivity between Tropical OLR and CAOs

A 30–80-day band period is selected for the analysis. This band is chosen because 1) it is a common MJO period where the highest spectra of MJO-related tropical OLR were found (e.g., Wheeler and Hendon 2004) and 2) the

FIG. A1. Connectivity between East Asian CAOs and tropical OLRs denoted by max absolute lag correlations (find a max from lag -10 to lag 0 days) between CAO indices and area-averaged OLRs for different time filter schemes. (top) Correlations between the W-CAO index and OLR over the Maritime Continent (15°S – 15°N , 100° – 130°E). (bottom) Correlations between the E-CAO index and OLR over the Indian Ocean (15°S – 15°N , 70° – 100°E) and the western Pacific (15°S – 15°N , 140° – 170°E).

strongest connectivity between leading tropical OLR and East Asian CAO indices is observed near this band period (Fig. A1). If we include a high-frequency band (<30 days), then the correlation is greatly reduced (Fig. A1). This simply indicates no notable interactions exist between East Asian CAOs and tropical disturbances at the short time scale during the preconditioning of a CAO event. Nevertheless, despite the inability of the MJO to predict the timing of high-frequency CAO events, the MJO has a potential impact on its occurrence probability (see section 5).

APPENDIX B

Statistical Significance Tests

Statistical significance levels for correlation coefficients are determined based on the two-sided Student's t test. The effective number of degrees of freedom n is achieved by removing autoregressive properties, as in Livezey and Chen (1983). Figure B1 shows a distribution of n from all correlations at every grid point and in all indices. Variable n is distributed from 90 to 200, which corresponds to critical correlation coefficients of 0.21–0.14 and 0.28–0.19 for the 95% and 99% confidence levels, respectively.

FIG. B1. A histogram showing the n used in the analysis. The top x axis denotes the critical value of the correlation coefficient for each block of n at 95% and 99% confidence levels.

In section 5, a composite analysis of short-term CAO events in the MJO phases is calculated. The number of events in each phase is tested using Monte Carlo simulation. The test is computed as follows: 1) the CAO events that occurred when the MJO intensity ≥ 0.5 are counted; 2) all CAO events satisfying step 1 are distributed randomly into eight MJO phases, resulting in a random number of CAO events for each phase; 3) step 2 is repeated 1000 times; 4) the probability density functions of random CAO events in all phases are averaged; 5) the highest and lowest 2.5% values are defined as the top and bottom thresholds; and 6) CAO composites above the top and below the bottom thresholds are considered significant.

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