Equatorial Circumnavigation of Moisture Signal Associated with the Madden-Julian Oscillation (MJO) during Boreal Winter

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Abstract

In order to describe the connection from an event of MJO to the next in the boreal winter, the eastward propagation of MJO is studied, focusing on that over the western hemisphere. Propagation signal is identified by EEOF analysis, performed on the bandpass filtered OLR for the period of 1979–2000. Besides NOAA OLR, total precipitable water (TPW), and surface winds from Special Sensor Microwave/Imager (SSM/I), precipitation observed from Microwave Sounding Unit (MSU), and reanalysis and operational analysis data of the European Centre for Medium-Range Weather Forecasts (ECMWF), are utilized for the composite.

Positive TPW anomalies are found, synchronizing with tropospheric and surface zonal wind anomalies. They propagate eastward all around the equator in the boreal winter. They propagate at a speed of about 6 ms^{-1} , with a Kelvin-Rossby coupled mode structure in the eastern hemisphere, and at about 20 ms^{-1} as an envelope of a radiating response in the western hemisphere. Within the envelope in the western hemisphere, faster propagating signals corresponding to $30\text{--}40 \text{ ms}^{-1}$ exist in the fields of TPW, zonal wind at 200 and 700 hPa, surface zonal wind. It is especially clear in the geopotential anomalies at 1000 hPa. This fast propagation speed of $30\text{--}40 \text{ ms}^{-1}$ is consistent with a first-baroclinic dry Kelvin wave mode recently rediscovered by Milliff and Madden (1996), and Bantzer and Wallace (1996). TPW increases under surface easterly anomalies along the equator. After the preceding TPW accumulation for 5–7.5 days, convective anomalies begin to occur as a part of the next cycle of the MJO from the eastern Atlantic to the western Indian Ocean.

These results suggest a following conceptual model for propagations and event-to-event connections of MJO. Equatorial Kelvin wave generated by convection of the MJO propagates eastward emanating from a warm pool region at a faster speed ($30\text{--}40 \text{ ms}^{-1}$) in the western hemisphere. Elevated topography of the South American and African continent, blocks the wave propagation. After being blocked several days by topography, they continue to proceed. As a result, the signal propagates at 20 ms^{-1} on average. Frictional convergence with lower easterlies of the dry Kelvin wave results in the associated propagation of TPW positive anomaly. Although it does not induce deep convections over large-scale subsidence regions, once it enters over the warm water in the western Indian Ocean, it helps to induce active convections for the next cycle of MJO.

1. Introduction

Madden-Julian Oscillation (MJO), first discovered by Madden and Julian (1971; 1972), is

one of the most prominent atmospheric phenomena in the tropics. Although its definition is not obvious, it is generally characterized as convection and circulation anomalies, which have zonal wave-number 1–3, and propagate eastward with a 30–90 day periodicity (comprehensive review can be seen in Madden and Julian, 1994). The spectral characteristic of convection shows narrower zonal scale and

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broader period range than that of circulation (Salby and Hendon 1994).

These features of MJO change with seasonal cycle. Longer periodicity in convection is found in the boreal winter than in the summer (Hartmann et al. 1992), and their amplitude is stronger in the boreal winter (Salby and Hartmann 1994). Furthermore, propagation characteristics of convection also alter (Wang and Rui 1990). In the boreal winter, convective anomalies associated with the MJO propagate eastward along the equator from Africa, or from the western Indian Ocean to near the date-line (Lau and Chan 1985; Knutson and Weickmann 1987). In the boreal summer, in contrast, they move northward or northeastward as well as eastward over the Indian Ocean, and the western Pacific (Lau and Chan 1986). In spite of these differences, throughout the year, convection related to the MJO is primarily confined over the warm pool region, from the eastern Indian Ocean to the western Pacific Ocean.

This is why most observational studies (Hendon and Salby 1994; Maloney and Hartmann 1998; Hsu and Weng 2001; Lawrence and Webster 2001; Kemball-Cook and Weare 2001) have stressed on how the convection evolves over the warm pool region. Their researches have advanced our understanding of the eastward movement of the large scale convective system associated with the MJO, especially over the warm pool region. For example, Hendon and Salby (1994) and Maloney and Hartmann (1998) showed that frictional convergence caused by a Kelvin wave might play a key role in evolving the MJO.

However, there is still no clear explanation what determines the periodicity of the convection associated with the MJO. Many general circulation models (GCMs) have defects of being unable to simulate the MJO sufficiently (Slingo et al. 1996): MJOs in numerical models have shorter periods than observed, and underestimated amplitude of variability. Recent composite studies (Maloney and Hartmann 1998; Kemball-Cook and Weare 2001) showed that moisture build-up takes place under surface easterly anomalies, before convection associated with the MJO over the warm pool region. This might determine migration speed of the MJO, but not yet the periodicity.

Propagating disturbances in the western hemisphere, where convection is almost absent, have been recently reinvestigated. Madden and Julian (1972) first discovered a fast propagating signal at about 20 ms^{-1} of surface pressure over the Pacific. Later, Knutson et al. (1986) indicated that upper tropospheric zonal wind anomalies circulate all around the equator. Their propagation speed in the western hemisphere is estimated as about 15 ms^{-1} . Recently, a number of observational studies (Milliff and Madden 1996; Bantzer and Wallace 1996) employing spectrum analysis showed that tropospheric dynamical signals, such as temperature, geopotential, and zonal wind cross over the Pacific at a faster speed of about 40 ms^{-1} emanated from the organized convection associated with the MJO over the warm pool region. These signals are identified with the first baroclinic equatorial Kelvin wave mode. In addition, some composite studies (Matthews 2000; Madden et al. 1999) demonstrated that corresponding surface disturbances of sea-level pressure, or surface zonal wind, tend to propagate all around the equator. They employ careful treatment of the phase of the MJO to construct the composite. Matthews (2000) suggested that circulating low sea-level pressure may trigger the convection over the Indian Ocean as a next cycle of the MJO. However, its physical process and likelihood was not clear yet.

In this study, the life cycle of the MJO is investigated focusing on the behavior of the moisture field associated with the MJO to clarify the physical process of its recurrence. In other words, whether and how the next cycle of the MJO convection is triggered by the former cycle, using satellite derived total precipitable water for a relatively long period is investigated. This paper consists of five sections. Data, and temporal and spatial filters are referred to in section 2. Section 3 describes the analysis method. The results are shown in section 4, and discussion and conclusions are provided in section 5.

2. Data

Propagating characteristics of the MJO were assessed using outgoing longwave radiation (OLR) from the National Oceanic and Atmospheric Administration (NOAA), and a composite was constructed based on it. OLR is a

good proxy for deep convection in the tropics. Daily OLR on $2.5^\circ \times 2.5^\circ$ grid points were used from 1979 to 2000. Daily total precipitable water (TPW) retrieved from Special Sensor Microwave/Imager (SSM/I) for the period of July 1987–2000 were utilized to analyze the moisture variation associated with the MJO. SSM/I provides global water vapor distribution over oceans with a good accuracy, the root mean square (rms) retrieval accuracy is 1.2 mm in the absence of rain (Wentz 1997). TPW was originally mapped on $1^\circ \times 1^\circ$ grids and area averaged data onto $2.5^\circ \times 2.5^\circ$ grids was used in this study.

Atmospheric motions associated with the MJO were described with daily reanalysis, and operational analysis data on $2.5^\circ \times 2.5^\circ$ grids produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). Reanalysis data were used for 1979–1993 and operational data were used for 1994–2000.

Daily surface winds (Atlas et al. 1996) over oceans produced by the combination of SSM/I wind speed, ECMWF analyses, and conventional surface observations for the period July 1987 to 2000 were also utilized. SSM/I provides surface wind speed data with an rms accuracy of 0.9 ms^{-1} (Wentz 1997), while ECMWF analyses assign wind directions. Surface winds were originally mapped on $1^\circ \times 1^\circ$ grids. We used area-averaged data that was mapped on $2.5^\circ \times 2.5^\circ$ grids. Precipitation data obtained by a Microwave Sounding Unit (MSU) from 1979 to October 1996 were also employed. This global precipitation data show good performance over the ocean compared with other global precipitation data such as the GOES precipitation index (GPI) or conventional observations (Spencer 1993). Precipitation data were daily gridded on $2.5^\circ \times 2.5^\circ$ grids over the ocean and missing over land.

All data mentioned above were processed using temporal and spatial filters in order to extract the fluctuations associated with the MJO. Firstly, to remove low-frequency disturbances longer than a seasonal cycle, a 90-day high pass Lanczos filter (Duchon 1979) was applied to all the data except for the data from SSM/I and MSU, because these data include missing values. The feature of the filter is to reduce the amplitude of what is called Gibbs phenomenon (detailed explanation can be found in Emery

and Thomson 2001 etc.). It is designed to reduce the amplitude by half at a 90-day frequency. Regarding SSM/I and MSU data, removal of a weighted 91-day running mean with 1:2:1 weights, (1-2-1 91-day running mean removal) was applied instead of employing a 90-day high pass filter. Secondly, in order to remove high-frequency disturbances, a 5-day running mean is applied to the high pass filtered (90-day high pass or 1-2-1 91-day running mean removal) data. Hereafter, these band-pass filtered data are called Intra-Seasonal Filtered (ISF) data. Finally, to focus on large-scale dynamics, two-dimensional spatial smoothing (Sardeshmukh and Hoskins 1984) was applied to these data. In short, spatially smoothed variables can be expressed as a weighted average over the neighborhood, with a weighting function depending only on the distance from a given grid. In this study, total wave number 36 was adopted, with which weighting function is reduced half at about 5° away from a given grid. This smoothing, thus, did not modulate fields strongly.

3. Analysis method

In order to determine the phase of the eastward propagating boreal winter mode MJO, the extended empirical orthogonal function (EEOF) method (Weare and Nasstrom 1982) following Lau and Chan (1985) was utilized, who succeeded in capturing globally eastward propagating signals associated with MJO. ISF OLR was first normalized in a way that the standard deviation at each grid for the entire period becomes unity. Next, the EEOF analyses are applied to the normalized OLR anomalies, over the equatorial region from 20°N to 20°S for the periods from December to May in 1979–2000. In this EEOF process, reduced data is used on $5^\circ \times 5^\circ$ grids for computational economy. The reason why normalized OLR anomalies were used first is as follows. The amplitudes of OLR variability relating to MJO differ much from location to location; larger over the warm pool region from the eastern Indian Ocean to the western Pacific Ocean. As a result, applying EEOF to the non-normalized OLR anomalies tends to be controlled primarily by variations over warm pool regions. Normalization can effectively reduce this local dependency and enables us to focus on the eastward

Fig. 1. Time evolutions of the first two EEOFs of OLR anomaly for the winter mode. Contour intervals are 0.01, and zero contours are suppressed. Values greater (less) than 0.01 (−0.01) are light (dark) shaded.

propagating signal of convection associated with MJO.

Like Lau and Chan (1985), we employed three time step extension and 5-day lag for EEOF analyses are employed. Using other-day lag, such as 4-day or 6-day, showed much similar result, however, utilizing 5-day lag seemed the best choice to represent the cyclic feature of MJO. The results of the first two EEOFs are shown in Fig. 1. The contributions of the first two EEOFs are 7 and 6% of the total variances respectively. In general, n time step extension of EOF results in the decrease of contribution to each mode by one- n th (Weare and Nasstrom 1982). In this case, three-time step extension is expected to reduce the contribution by about one-third. Thus, although the contribution of the first two EEOFs are low, they are statistically significant, while higher modes are not significant according to the North's test of significance (North et al. 1982). Convective signals of EEOF₁ and EEOF₂ having zonal wave number one, and some higher modes move eastward, and connect at day 0 and day 10; day 10 of EEOF₁ corresponds to day 0 of EEOF₂, and day 10 of EEOF₂ corresponds to day 0 of EEOF₁ in opposite sign. Convective anomalies thus show circulating characteristic and a

period of convective patterns circulating all around the equator is expected to be about 40 days.

Next, we projected the OLR data onto the first two EEOFs for the entire period for 1979–2000 to obtain time series of first two principal components (hereafter referred to as PC_1 and PC_2). Using entire period instead of wintertime only allows us to collect as many events as possible.

As an example of the correspondence between the first two PCs and OLR variation, a Hovmöller diagram of the ISF OLR, averaged along the equator and time series of first two PCs in 1990, are shown in Fig. 2. Negative OLR anomalies appear to circulate all around the equator from January to May, when first two PCs are large in amplitude. While negative OLR anomalies appear to be primarily confined to the eastern hemisphere in the boreal summer when first two PCs are small. Thus these first two PCs reflect globally eastward propagating convective signals well.

Time series of first two PCs for the entire period used in this study are shown in Fig. 3. There is a strong seasonal cycle, and inter-annual variation of the amplitude of the first two PCs. The amplitude is large in November

Fig. 2. (a) Hovmöller diagram of 90-day high pass filtered and 5-day moving averaged OLR anomalies averaged between 7.5°N and 7.5°S along the equator for 1990. Contour interval is 10 Wm^{-2} . Solid (dashed) contours indicate negative (positive) values less (greater) than -5 Wm^{-2} (5 Wm^{-2}). Shades are for values less than -15 Wm^{-2} . (b) First (solid) and second (dash) normalized eigenvalues time series correspondence to first and second eigenvectors for EEOF analysis for the period from December to May.

to June, especially in January to April. In interannual timescale, there is no apparent relationship between the amplitude and underlying basic state such as El Niño and the Southern

Oscillation (ENSO). For example, in 1987/88 El Niño winter, the amplitude is relatively large, although, in 1997/98 El Niño winter, the amplitude is considerably small. Similar conclusions have been obtained by previous studies (Hendon et al. 1999; Slingo et al. 1999; Lawrence and Webster 2001).

Next, utilizing first two PCs, the composite life cycle of the MJO is constructed. PC_1 always leads PC_2 by about a quarter cycle, and the vector $\mathbf{P}$ consisted of PC_1 and PC_2 is considered as:

$$\mathbf{P}(t) = (PC_1(t), PC_2(t)) = |\mathbf{P}(t)|e^{i\alpha(t)}, \quad (1)$$

$$\alpha(t) = \tan^{-1}(PC_2(t)/PC_1(t)), \quad (2)$$

where t is time in real space, and $\alpha(t)$ is a phase angle of the $\mathbf{P}(t)$ ranging from 0 to 2π . 16 phases are defined by dividing at an equal angle: phase 0 corresponds to $\alpha = 0$, and phase 4 corresponds to $\alpha = 0.5\pi$. Since a period of one cycle of the MJO is expected to be about 40 days, an interval of categories corresponds to about 2.5 days. Thus from one time increment to 2.5 days is converted in interpreting composite results. In order to extract the significant events of the boreal-winter-mode of MJO, the events in which the value $|\mathbf{P}|$ exceeds 1.3σ are composited, where σ is the standard deviation of $|\mathbf{P}|$ for the entire period of 1979–2000. The threshold is decided by taking into consideration that σ is calculated by all seasons instead of winter time only. The events selected in this study occurred mostly in the boreal winter as shown with shades in Fig. 3. At least one event occurred per season, and the maximum numbers of events that occurred per season is four to five, e.g., in 1996/1997 winter. Altogether the numbers of composited events differ slightly from one category to another, ranging from 51 to 63 for the period 1979–2000.

Statistical significance of the composite results should be assessed carefully because temporal and spatial filters were applied to the data. To assess the statistical significance, the degree of freedom (DOF) needs to be specified. However, in case of using a filter, the exact DOF is not known. Following Madden et al. (1999), a Student- t test using minimum DOF was applied to judge the significance of the composite results in this study. In short, the

Fig. 3. Time series of PC_1 (thick lines), PC_2 (thin lines), and events of MJO (shades) that are used to construct the composite. The criterion of identifying events is based on the square root of PC_1 and PC_2 . See text for details.

composited event numbers in each category were adopted for DOF. The process of significance test is described in the appendix in detail.

4. Composite life cycle of the MJO

The composite results are shown in this section. First of all, an overview of propagating

signals in some variables is given in subsection a. Second, the propagation of TPW signal in association with surface wind anomalies in subsection b is focused on. Third, how the propagating TPW signal connects to the deep con-

vective activity of the next MJO cycle is shown in subsection c. Then, a detailed description about going over topography are shown in subsection d. Finally, the relationship between accumulation of TPW and precipitation over the equatorial western Indian Ocean, where initiation of the MJO is considered to occur, is shown in subsection e.

a. Relationship between convection, circulation and moisture: Hovmöller diagrams

An overview of the composite results using Hovmöller diagrams is given in this subsection. The life cycle of the OLR and upper-tropospheric circulation (zonal wind and velocity potential) anomalies associated with the MJO, as indicated in Figs. 4a–c, show characteristics similar to the well-known features (e.g., Knutson and Weickmann 1987; Hendon and Salby 1994). The OLR anomalies are confined primarily to the eastern hemisphere; however, the upper-tropospheric circulation anomalies circulate all around the equator. As indicated with straight lines in Fig. 4, the circulation anomalies together with the OLR anomalies propagate at a speed of about 6 ms^{-1} in the eastern hemisphere, and the circulation anomalies accompanying intermittent signal of OLR propagate at a speed of about 20 ms^{-1} in the western hemisphere. The contrast of phase speed between the eastern hemisphere and the western hemisphere is consistent with, but somewhat faster in the western hemisphere than the results of Knutson et al. (1986) and Hendon and Salby (1994) ($5\sim 6 \text{ ms}^{-1}$ in the eastern hemisphere, $10\sim 15 \text{ ms}^{-1}$ in the western hemisphere). The phase relationship be-

Fig. 4. Hovmöller diagrams for composite anomalies averaged between 7.5°N and 7.5°S . Light (dark) shades indicate positive (negative) values significant at the 99% confidence level. Thick lines indicate constant phase speed corresponding to 20 and 6 ms^{-1} . (a) OLR, (b) velocity potential at 200 hPa , (c) 200 hPa zonal wind. Contour intervals are 3 Wm^{-2} in (a), $1 \times 10^6 \text{ m}^2\text{s}^{-1}$ in (b), and 1.0 ms^{-1} in (c).

tween the circulation and convection anomalies changes during a life cycle of the MJO. In the western hemisphere, and in the regions where OLR anomalies exist, zonal wind anomalies lag behind OLR anomalies by about a quarter cycle. In the eastern hemisphere, the phase difference becomes slightly smaller.

In the present study, we would like to address the point that not only the upper-tropospheric circulation anomalies but also positive TPW anomalies circulate all around the equator as shown in the longitude-time section plot in Fig. 5a. They show coherent propagation, lagging slightly less than a quarter cycle from the upper level circulation anomalies. In the eastern hemisphere, they propagate at a speed of about 6 ms. In the western hemisphere, although there are some discontinuities over continents, positive TPW anomalies propagate at a speed of about 20 ms on average. There also is a faster propagating signal found in the western hemisphere, which will be discussed later. The positive TPW anomaly leads the negative OLR anomaly by about a quarter cycle over the regions from 100°W to 60°E, and by about 1/8 cycle from 60°E to the date line, suggesting that the increase of the TPW induces the convection associated with the MJO. It is interesting to find the accumulation of TPW positive anomalies almost simultaneously at the east and west coast of the African continent, so that the lead of positive TPW anomaly at the east coast (40–60°E) is larger than further eastern longitudes. This early onset at the African east coast is also found in the MSU precipitation (Fig. 5b), as well as in the surface convergence (Fig. 5c) and very slightly in the OLR anomaly. It is suggested that the accumulation of the TPW associated with the very fast mode initiates the deep convection that will induce the next cycle of the MJO. This will be examined in detail in the following subsections.

A robust relationship, between TPW anomalies and lower level atmospheric dynamical disturbances, can be seen in the fields of zonal wind and geopotential as shown in Fig. 6. The almost simultaneous accumulation of TPW at the African west and east coasts mentioned previously are associated with fast propagating low geopotential anomaly over the African continent (Fig. 6c). While the surface convergence

Fig. 5. Same as Fig. 4 except for (a) SSM/I precipitable water, (b) MSU precipitation, and (c) divergence derived from SSM/I surface wind data averaged between 2.5°N and 2.5°S. Shades are drawn in the same manner as in Fig. 4 but for at the 95% confidence level. Contour intervals are 0.5 mm in (a), 0.5 mm day⁻¹ in (b), and 0.02 day⁻¹ in (c).

Fig. 6. Same as Fig. 4 except for (a) 700 hPa zonal wind, (b) 1000 hPa zonal wind, and (c) 1000 hPa geopotential. Contour intervals are 0.2 ms^{-1} in (a), 0.2 ms^{-1} in (b), and $10 \text{ m}^2\text{s}^{-2}$ in (c).

at the east coast primarily consists of the zonal wind component as seen in Fig. 6b, it is not found in the west coast. It will be shown that the meridional wind components contribute to the TPW increase at the west coast, in the following subsection. Zonal wind anomalies at 700 hPa and at the surface are in phase and show circumnavigating characteristics. They are coherent with TPW anomalies and easterly anomalies almost correspond to the increasing phase of TPW. Fast propagating signals, with a phase speed of $30\sim 40 \text{ ms}^{-1}$, also exist in the western hemisphere. Although surface zonal wind anomalies of SSM/I are missing over the American or the African continent, surface zonal wind anomalies also seem to circulate all around the equator.

Geopotential anomalies at 1000 hPa exhibit clear fast-propagating characteristics ($30\sim 40 \text{ ms}^{-1}$) in the western hemisphere. They show coherent propagation and also show a circumnavigating feature; in the eastern hemisphere, they propagate at a speed of about 10 ms^{-1} . Their propagation is obstructed by the topography over Central America and Africa, and after several days they continue to propagate. This propagating characteristic is something different from the other fields at the surface. For example, compared to surface zonal wind (Fig. 6b), blocking over central America is less obvious. As a result, surface geopotential anomalies start to lag behind surface zonal wind anomalies over central America, and the difference become a quarter over Africa. This phase difference may play some role in accelerating surface easterly anomalies over Indian Ocean at phase 6–10 that will be discussed later. Topographic blocking effects were also described in Matthews (2000), and in Nitta (1992) by Sumatra Island. The vertical profile of the zonal wind anomalies (not shown) have one node, suggesting that these fast eastward propagating dynamical disturbances have a first baroclinic structure. The characteristics of these fast moving perturbations are consistent with the first-baroclinic mode equatorial Kelvin wave recently rediscovered by Milliff and Madden (1996), Milliff et al. (1998), and Bantzer and Wallace (1996). The fast propagating signal in the TPW field mentioned in the previous paragraph is considered to be associated with this fast-moving baroclinic Kelvin mode.

Fig. 7. Composite maps of precipitable water (left panel), and 1000 hPa winds and OLR in contours and shades (right panel) for phase 0 (a)(b), phase 2 (c)(d), phase 4 (e)(f), and phase 6 (g)(h). Light (dark) shades in left panels indicate significant positive (negative) values at 95% confidence level. Contour interval are 0.5 mm and zero contours are depressed. Thick arrows in right panels are those which pass the significance test at 99% confidence level. Negative OLR at the 99% confidence levels are shaded, and positive OLR at the 99% confidence levels are contoured at 5, 10, 15 Wm^{-2} .

b. Detailed characteristics of circulating TPW and atmospheric motions associated with the MJO

Next we examine how TPW anomalies in association with circulation anomalies propagate eastward in details. To this end, composited anomaly maps of TPW, OLR and surface winds anomalies at phase 0, 2, 4, and 6 are shown in Fig. 7. TPW anomalies and their 95% significant regions are drawn in the same manner within Fig. 5. Vectors represent anomalous winds, and 99% significant winds are depicted with thick vectors. OLR anomalies at 99% significance are drawn by shades (negative values), and contours (positive values).

First of all, the composite result at phase 0 is shown. This is the time just after the negative OLR anomalies associated with MJO reach maximum amplitude in its life cycle (see Fig. 4a). A convective anomaly is centered over the equatorial eastern Indian Ocean, and extends to the western Pacific (Fig. 7b). To the east and west of the convective anomalies, Kelvin-Rossby wave responses are found in the field of surface winds anomaly: the former over the Pacific Ocean, and the latter over the Indian Ocean. To the east of the center of the convective anomalies, positive, significant TPW anomalies are present (Fig. 7a). Two maxima of positive TPW anomalies, are found around 100°E, 10°S, over the southeast part of the convective

center, and around 150°E , 0° , to the east of the convective center. Further eastward extension of positive TPW anomalies can be seen along the ITCZ, where surface easterly anomalies are present. Over the eastern part of the positive TPW anomalies (around 160°W , 15°N), there is weak, but significant convective anomaly.

By phase 2, positive TPW anomalies have moved eastward (Fig. 7c). Its eastern edge has reached the west coast of South America, and has become large in amplitude. At that time, surface winds converge from northeast in the northern hemisphere, and from southeast in the southern hemisphere, to the west of South America (Fig. 7d). There are two other local maxima in the TPW fields. One is over the central Pacific east of the dateline in the northern hemisphere, where significant convective signal is found. The other is over the western Pacific between New Guinea and Australia southeast of the strong convection. The center of the negative OLR anomalies is now located over the maritime continent. To the east, and west of the convective anomalies, Kelvin-wave and Rossby-wave responses still exist and surface easterly anomalies as a Kelvin-wave response extends far east of the positive TPW anomalies.

At phase 4, there still exist three local maxima of positive and significant TPW anomalies (Fig. 7e). Eastern parts of the positive TPW anomalies are amplified and western parts of that are diminished in amplitude. In addition, positive but non-significant TPW anomalies appear over the Atlantic Ocean. Thus positive TPW anomalies have moved eastward as a whole. Over the central Pacific far east of the dateline in the northern hemisphere, convective anomaly still exists corresponding to the local TPW maximum (Fig. 7f). Convection and circulation anomalies have also moved eastward.

Finally, by phase 6 positive TPW anomalies have crossed over the Atlantic Ocean, and now cover a wide range along the equator from the western Pacific to the western Indian Ocean (Fig. 7g). Anomalous surface easterlies have also moved eastward and then are over the western Indian Ocean. Anomalous convection is found over Sahara, and Argentina (Fig. 7h). Accumulation of TPW starts to the east of the African east coast that will connect to the next

cycle of the MJO as discussed in the next subsection. After all, positive TPW anomalies cross over the Pacific and Atlantic in association with surface easterly anomalies; the latter always precedes the former.

c. Initiation of the next cycle of MJO

We now focus on how the next cycle of MJO is evolved. Composite anomaly maps of TPW, surface winds from SSM/I, and OLR at phase 7–11 are shown in Fig. 8: left panels for TPW, and right panels for surface winds and OLR anomalies.

At phase 7, anomalous convection over the Sahara and South America that is found at phase 6 is still present, and the former have strengthened (Fig. 8b). Corresponding convergences at 700 hPa and divergence at 200 hPa exist over these regions (not shown). TPW anomaly patterns have moved eastward as a whole (Fig. 8a). The equatorial Atlantic is covered with moist anomaly. And positive TPW anomaly over the western Indian Ocean is amplified up to 2.5 mm. On the other hand, positive TPW anomalies over the eastern Pacific have weakened, compared to phase 6 (Fig. 7g).

Subsequently (phase 8), anomalous convection develops near the equator extending from South America to the Sahara (Fig. 8d). Associated convergences at 700 hPa also exist (not shown), especially strong over the east coast of equatorial South America. No significant large-scale circulation associated with the convection is found. On the other hand, surface easterlies over the Indian Ocean have shifted slightly eastward and strengthened. Consequently, anomalous surface convergences in the western Indian Ocean have been amplified, and TPW anomalies in the western Indian Ocean have increased (Fig. 8c). By this time, positive TPW anomalies over the equatorial eastern Pacific have disappeared.

At phase 9, a new weak convective anomaly appears over the western Indian Ocean (Fig. 8f), where the TPW peak has been found. In addition, preexisting convection extended from South America to Africa splits into western and eastern part, with the former strengthened and the latter shifted eastward. Over the Indian Ocean, anomalous surface easterlies became amplified, accompanied by the eastward shift of positive TPW anomalies (Fig. 8e). Positive

Fig. 8. Composite maps of precipitable water (left panel) and 1000 hPa winds and OLR in contours and shades (right panel) for phase 7 (a)(b), phase 8 (c)(d), phase 9 (e)(f), phase 10 (g)(h), and phase 11 (i)(j). Light (dark) shades in left panels indicate significant positive (negative) values at 95% confidence level. Contour interval are 0.5 mm and zero contours are depressed. Thick arrows in right panels are those which pass the significance test at 99% confidence level. Negative OLR at the 99% confidence levels are shaded, and positive OLR at the 99% confidence levels are contoured at 5, 10, 15 Wm^{-2} .

TPW anomalies now cover the equatorial western Indian Ocean. The maximum TPW anomaly is located at around 60°E . Moist anomalies over the equatorial Atlantic still exist.

By phase 10, developed anomalous convection has shifted eastward (Fig. 8h). Anomalous convection over Africa has developed with

strengthened cyclonic circulation over the equatorial Atlantic Ocean. Convection over the western Indian Ocean also has been strengthened. In contrast, anomalous convection over the east coast of South America decayed. Anomalous surface easterlies over the Indian Ocean are still observed. Corresponding TPW anomalies

have elongated along the equator (Fig. 8g). However, the peak is still at around 60°E.

Finally by phase 11, convective anomalies over the western Indian Ocean have developed (Fig. 8j), which may be considered as the onset of the MJO. In addition, anomalous convection over equatorial Africa is still moderately strong, and cyclonic circulation anomalies west of the convection still exist. Anomalous convections over the east coast of South America have been weakened further. Over the Indian Ocean, surface easterly anomalies have shifted eastward and a wide strip of positive TPW anomalies exist along the equator (Fig. 8i). This is located to the east of the convective anomaly, suggesting following eastward shift of convection center.

After all, accumulation of TPW was always found ahead of the convective anomalies throughout the composite life cycle of the MJO. Particularly notable is that after the significant convective anomaly ceases eastward propagation beyond the central Pacific Ocean, a fast signal of surface easterlies associated with TPW accumulation maintains the eastward propagation, and a new deep convective activity starts over the western Indian Ocean following the TPW accumulation. In accumulation of TPW over the western Indian Ocean, the phase difference between surface geopotential and surface zonal wind may assist the process. According to the linearized Kelvin wave theory, zonal wind acceleration is proportional to the inverse longitudinal gradient of geopotential field. Thus, surface easterly anomalies over the Indian Ocean are accelerated at phase 7–10 (the phase relationship can be found in Figs. 6a and 6b).

d. Orographic effect of Africa

In this subsection, we describe how the surface easterlies are related with the upper zonal wind disturbances where exist high mountains of over 4000 m. To this end, composite vertical profile of zonal wind anomalies (upper panels) and TPW anomalies (lower panels) both averaged from 2.5°N to 2.5°S as a function of longitude are made (Fig. 9). By phase 3, zonal wind anomalies having first baroclinic vertical structure have reached Africa as a front of Kelvin wave (see Figs. 4, 6). There also exist, first baroclinic zonal wind anomalies in opposite phase

over the Indian Ocean, which is ascribed to the convective anomaly of the previous cycle. At this time, TPW accumulate over the eastern Pacific, while TPW anomalies are nearly zero over the Atlantic and are negative over the Indian Ocean.

At phase 4, the upper-level westerly anomalies migrate over Africa first, while the lower-level easterly anomalies seem to be blocked by the topography. TPW anomalies over the eastern Pacific have increased, and slightly positive values are found over the Atlantic Ocean. At phase 5, the lower-level easterly anomalies then travel over Africa, and catch up with the upper level zonal wind anomalies. Then anomalous lower-level easterlies extend across a broad area from the eastern Pacific to the western Indian Ocean. TPW has risen over the Atlantic and just to the east of Africa. By phase 6, lower easterly anomalies have almost completely jumped over Africa. Afterwards, zonal wind anomalies move eastward and get strengthened (phase 7): surface anomalous easterlies are confined to the Indian Ocean. By this time, positive TPW anomalies over the western Indian Ocean have increased to ~3 mm.

Looking over Figs. 4–6 again, faster propagating ($30\sim40\text{ ms}^{-1}$) signals are found in the fields of TPW, zonal wind at 200 hPa and surface, and 1000 hPa geopotential anomalies in the western hemisphere. These faster signals correspond to the first baroclinic free wave shown by Milliff and Madden (1996). They propagate eastward gathering moisture, and its propagation is sometimes blocked for several days over mountains, and it continues to proceed. As a result, they seem to move at a speed of about 20 ms^{-1} on average. Gathering moisture under large-scale subsidence regions such as eastern Pacific and western Atlantic Ocean does not induce deep convections that produce atmospheric disturbances. Once they have reached over the eastern Atlantic to the Indian Ocean, where SST is high enough, moisture gathering induce deep convections that generate new atmospheric disturbances that is identified as the next cycle of MJO.

e. Relationship between TPW, precipitation and vertical structure of the atmosphere

Finally, in order to examine the relationship between TPW and precipitation at the equato-

Fig. 9. Composite structures of zonal wind anomalies (upper panels) and TPW anomalies (lower panels) averaged between 2.5°N to 2.5°S. In upper panels, ordinate indicate pressure in hPa. Light (dark) shades are positive (negative) values significant at the 99% confidence level, and shades change at increments of greater (less) than 4 (-1) ms^{-1} . Contour intervals are 0.5 ms^{-1} . In lower panels, ordinate indicate water amount in mm. Light (dark) shades denote negative (positive) anomalies, and significance interval at 95% averaged from 120°W to 120°E at a phase are indicated by error bars.

rial western Indian Ocean (0° , 60°E), where large-scale deep convection associated with the MJO onset occurs, the composite time series in Fig. 10 are shown. In addition, to get physical insights on convective development, vertical profiles of the pressure-velocity and humidity from ECMWF analyses data are also shown.

Positive TPW anomaly leads precipitation by about 5–7.5 days. The increase of TPW almost coincides with anomalous surface easterlies. During the TPW accumulation stage, upward motion is primarily confined to the lower troposphere, and the specific humidity increases in the lower troposphere with a maximum at about 700 hPa. Several days later, when precipitation anomaly reaches its peak at phase 12, upward motion is strongest and significant in the upper troposphere. The significant positive specific humidity anomalies exist in the upper troposphere, while the lower atmosphere starts to dry and there already is no significant upward motion. Afterwards, as precipitation dies out TPW decreases rapidly. Atmosphere becomes dryer in the same manner as moistening, but with opposite sign.

5. Discussion and conclusion

In this study, to examine the connection from an event of MJO to the next during boreal winter, when eastward propagation of MJO is predominant, the composite life cycle of MJO using EEOF analysis of boreal wintertime OLR anomalies is made. In particular, the composite results are analyzed focusing on the behavior of water vapor field in the western hemisphere, associated with winter mode MJO using SSM/I data. Positive TPW anomaly propagates all around the equator (Fig. 5). It always coheres with lower zonal wind anomaly. In short, it moves at a speed of about 6 ms^{-1} in the eastern hemisphere, and moves at a speed of about 20 ms^{-1} in the western hemisphere. It has been shown that the eastern hemisphere, where circulation strongly couples with convection, frictional convergence caused by forced Kelvin wave response, plays a key role in accumulating moisture east of the convection (Hendon and Salby 1994; Maloney and Hartmann 1998).

We emphasize that radiating Kelvin waves in the western hemisphere is also accompanied by TPW accumulation. Increase of TPW occurs under surface easterly anomalies.

Fig. 10. Composite life cycle of TPW, precipitation, specific humidity, and pressure velocity anomalies over the equatorial western Indian Ocean (average over 55°E – 65°E , 5°N – 5°S). Upper panel shows TPW (solid line) and precipitation anomaly (dashed line). Left and right axis denotes TPW in mm and precipitation in mm day^{-1} respectively. Shaded areas correspond to anomalous surface easterlies. Confidence intervals at 95% for the average valued of TPW and precipitation are indicated with closed circles and open circles. Lower panel shows specific humidity anomalies (contour) and vertical p -velocity (vectors) obtained from ECMWF data. The contour interval is $5 \times 10^{-5} \text{ mm}$. Positive (negative) specific humidity anomalies over than 90% significant level are shown with light (dark) shades, and 90% significant vertical velocity anomalies are shown by thick vectors. Abscissa is the composite phase number. All variables are obtained by averaging a rectangular box with $10^\circ \times 10^\circ$ centered on the reference point (0° , 60°E).

There are two candidates for the mechanism of accumulation of TPW: surface convergence, and evaporation. Contour interval of divergence (0.02 day^{-1}) in Fig. 5c could be translated to boundary-layer moisture convergence (0.6 mm day^{-1}), based on the assumption that the boundary layer is 150 hPa height, and boundary-layer averaged specific humidity is 0.02 kg kg^{-1} . The amplitude of variation of TPW anomaly in the western hemisphere ($2\sim 3 \text{ mm}$), could be explained by boundary layer moisture convergence. Thus, surface convergence produced by a radiating Kelvin wave is an important factor for creating propagation of positive TPW anomaly.

Next, the radiating Kelvin wave in the western hemisphere is discussed in detail. The propagation speed of 20 ms^{-1} is still too small for the tropospheric dry first baroclinic Kelvin wave. Although, within the envelope propagating at a speed of about 20 ms^{-1} , faster propagating signals corresponding to $30\text{--}40 \text{ ms}^{-1}$ exist in the fields of TPW, upper- and lower-tropospheric zonal wind, and it is especially clear in the geopotential height anomalies at 1000 hPa. These zonal wind anomalies have first baroclinic structure, and correspond to the fast equatorial Kelvin wave mode recently rediscovered by Milliff and Madden (1996) and Milliff et al. (1998). This wave is blocked over elevated topography of South American, and African continents. We examined this feature in the case of the African continent. Upper-level zonal wind anomalies travel over the continent first, and then lower-level zonal wind anomalies catch up with them several days later. Similar features were described about the topographic effects of the Sumatra Island by Nitta (1992) based on small numbers of cases. Until the lower-level zonal wind anomalies travel over the continent, and propagation of upper-level zonal wind anomalies slows down. Thus, the average speed of about 20 ms^{-1} might be present as an obscure signal resulted from fast propagating ($\sim 40 \text{ ms}^{-1}$) wave occasionally blocked over continents.

The final point is the relationship between accumulation of TPW, and convection development of MJO will be discussed. The positive TPW anomaly does not connect to deep convection in the western hemisphere. However, when it enters into the eastern hemisphere, the

connection between accumulation of TPW and convection development becomes evident. Over the western Indian Ocean, when the MJO initiates, it is shown that a maximum of precipitation anomaly is observed after the TPW peak lagging by 5–7.5 days. Recent studies based on satellite data (Maloney and Hartmann 1998), and on radiosonde data (Kemball-Cook and Weare 2001) also suggested that moisture build-up is present before convection associated with MJO occurs. At the middle stages of the TPW accumulation, lower to middle level specific humidity, increases and upward motions are primarily confined to the lower to middle troposphere. Several days later, the precipitation anomaly reaches its peak, with stronger upward motions and significant positive specific humidity anomaly in the upper troposphere. This suggests that there is a shallow convection stage before active convection. There may be some reasons that account for the time lag between moisture build-up and deep convection. One possible explanation may be the entrainment effect. Recently, Brown and Zhang (1997) demonstrated, by calculating a simple parcel model, that the warm pool atmosphere above the boundary layer can be dry enough to discourage the growth of deep convective clouds by depleting parcel buoyancy through entrainment. Additional reason may be the existence of the quasi stable layer at 0°C level (Johnson et al. 1996; Johnson et al. 1999). This stable layer might prevent convection from developing deeply for several days. In a separate study, the development of convection associated with MJO observed during TOGA-COARE is examined using high resolution GMS histogram data in detail. In that study, we showed that there exist two stages that most clouds populated below trade inversion, and melting level, before mature convection. During these stages, atmosphere below each stable layer had moistened. This result suggests that the relationship between the existence of stable layers, and moistening of atmosphere, is an important factor to understand the development of convection associated with MJO.

As a conclusion, we put the ideas about the eastward propagation properties and event-to-event connections of the boreal winter MJO together. Over the warm water pool, it is considered that frictional convergence associated with

the forced Kelvin wave response to the east of convection (Hendon and Salby 1994) accounts for the accumulation of water vapor (Kemball-Cook and Weare 2001; Maloney and Hartmann 1998), that determines the slow eastward propagation. In the western hemisphere, various studies in 1980s have shown that “dry atmospheric disturbances” in the western hemisphere somehow connect to the next event of the MJO. The connection was especially clear in the upper tropospheric wind disturbances (Knutson and Weickmann 1986). However, there remained two unsolved issues. The first is that the faster propagation speed of 20 ms^{-1} is still too slow for the tropospheric dry first baroclinic Kelvin wave. Secondly, it cannot be understood how the upper tropospheric wind disturbances could trigger the active convection of the next cycle of the MJO. We showed that within the envelope, a faster eastward propagating mode ($30\text{--}40 \text{ ms}^{-1}$) exists in the western hemisphere in the fields of upper and lower tropospheric zonal winds, 1000 hPa geopotential, and TPW. These faster propagating signals correspond to the fast equatorial Kelvin wave mode recently rediscovered by Milliff and Madden (1996), Milliff et al. (1998) and Bantzer and Wallace (1996). Elevated topography of the South American and African continent, blocks the wave propagation. After several days blocked by topography, they continue to proceed. As a result, the signal propagates at a speed of about 20 ms^{-1} on average. Frictional convergence with lower easterlies of the dry Kelvin wave results in the associated propagation of TPW positive anomaly. Although it does not induce deep convections at the longitudes under large-scale subsidence, once it enters over the warm water in the western Indian Ocean, it helps to induce active convections for the next cycle of MJO.

In this study, we focused on the MJO events that mainly occur in boreal winter and their initiation process. In the boreal summer, a different mechanism may exist for convective initiation of the MJO. For one thing, eastward-propagating intraseasonal signals of convection and circulation are relatively weak (Salby and Hendon 1994) and significant northward propagations at Asian monsoon longitudes are also observed. For another thing, intraseasonal periodicity is shorter than that during boreal winter

(Hartmann et al. 1992). We are currently working on clarifying the differences in the propagation properties of the MJO during boreal summers and boreal winters.

Appendix

Process of significance test

In the appendix, we describe how to assess the significance of the composite results following Madden et al. (1999). According to Student's t test, significance of the mean depends on the sample means, the pooled estimate of the common standard deviation, and the DOF. In using temporal and spatial filter, the DOF is uncertain.

First of all, we express any variable as $X_{i,j,k}$, where i, j, k represents the i -th grid, j -th day, and k phase respectively. Composite average for phase k is expressed as;

$$\bar{X}_{i,k} = \frac{1}{N_k} \sum_j \sum_{i'}^{I_j} w_{i'} X_{i',j,k}. \quad (3)$$

I_j is the number of spatial grids at i -th grid on day j , and $w_{i'}$ is the weight at i' -th grid both used in the spatial smoothing. J_k is the total number of days used to construct the composite for phase k . N_k is the total number of grids for phase k .

$$N_k = \sum_j \sum_i^{I_j} w_i. \quad (4)$$

Then the grand average at grid i (X_i) and the variance at grid i for phase k ($S_{i,k}^2$) is

$$\bar{X}_i = \frac{1}{16} \sum_{k=1}^{16} X_{i,k}, \quad (5)$$

$$S_{i,k}^2 = \frac{1}{N_k} \sum_j \{ (X_{i,j,k} - \bar{X}_{i,k}) w_i \}^2, \quad (6)$$

and a pooled variance, S_i^2 , is

$$S_i^2 = \frac{1}{\sum_{k=1}^{16} K_k} \sum_{k=1}^{16} K_k S_{i,k}, \quad (7)$$

where K_k is the number of events selected for a given phase. Number of events is not J_k , which includes continuous days, but is the total number of intermittent days. Using these parameters, Student's t (T) can be expressed as

$$T_{i,k} = \frac{\bar{X}_{i,k}}{[S_i^2/K_k]^{1/2}}. \quad (8)$$

Thus corresponding DOF here is K_k , although the true DOF must be a somewhat larger number. In this study we utilize T to assess the significance of composite results, imposing higher criteria on significance tests than necessary.

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