An Ensemble Hindcast of the Madden-Julian Oscillation during the CINDY2011/DYNAMO Field Campaign and Influence of Seasonal Variation of Sea Surface Temperature

Hiroaki MIURA

Department of Earth and Planetary Science, The University of Tokyo, Bunkyo, Japan
Japan Agency for Marine-Earth Science and Technology, Yokohama, Japan

Tamaki SUEMATSU

Department of Earth and Planetary Science, The University of Tokyo, Bunkyo, Japan

and

Tomoe NASUNO

Japan Agency for Marine-Earth Science and Technology, Yokohama, Japan

(Manuscript received 8 April 2015, in final form 18 September 2015)

Abstract

The ensemble hindcast initialized during 12–16 October 2011 is performed using a global cloud-system-resolving model (CSRM) with a horizontal mesh size of approximately 14 km. The ensemble size is five and the duration of each simulation is 60 days. When sea surface temperature (SST) with a realistic time evolution is prescribed, not only the first but also the second Madden-Julian oscillation (MJO) event observed during the CINDY2011/DYNAMO period emerges in the ensemble mean, although the signal of the second MJO is unsatisfactory in each member. This result leads to a hypothesis that the second MJO is significantly constrained by the prescribed seasonal change of SST.

The analyses of the observational data indicate that an MJO favorable environment, in which SST of the southeastern Maritime Continent is higher than that of the Indian Ocean, is established in late November to early December. Humidity in the lower troposphere increases substantially in the southeastern Maritime Continent during the period. A pair of sensitivity tests using a global CSRM clearly shows that the eastward migration of convection during the second MJO is at least partly caused by the climatological seasonal change of SST. The results of this study indicates that it is inappropriate to treat the climatological seasonal change as the background of the MJO during this season, because its timescale is short enough to be comparable with the intraseasonal timescale of the MJO. We provide a perspective that a certain type of the MJO can be regarded as a transition process, responding to the eastward shift of the region of large-scale positive buoyancy production following the warmer SST.

Keywords  Madden-Julian oscillation; intraseasonal variation; air-sea interaction; large-scale buoyancy
1. Introduction

The Madden-Julian oscillation (MJO; Madden and Julian 1971, 1972) is known as the dominant intraseasonal (30–90 days) variability of atmospheric circulation and convective activity in the tropics. A coupling between large-scale atmospheric circulation and the packets of organized cloud systems in the eastern hemisphere is a distinct character of the MJO (e.g., Knutson and Weickmann 1987). The MJO modulates atmospheric circulation in the global scale and influences a variety of phenomena from spatially small and temporally short ones to larger and longer ones. The role of the MJO in bridging weather and climate is regarded as one of the important properties of the MJO (Zhang 2013). Because of the global impacts that the MJO exerts during its life cycle, there have been strong interest in improving the representation of the MJO in weather forecast models for better seasonal prediction not only in the tropics but also in the mid-latitudes (e.g., Vitart 2014). Interests in the MJO are evoked not only by its enormous social impacts. The elusive physical behavior of the MJO has motivated a large number of researches from the viewpoints of observations, theories, and numerical modeling (e.g., Zhang 2005).

Along with its intraseasonal timescale, characteristics of the MJO that are widely accepted include initiation with a strong signal of convective activity over the Indian Ocean and its slow eastward movement in the speed of approximately 5 m s$^{-1}$ toward the western Pacific. An envelope of active convection that develops over the Indian Ocean, however, does not always proceed to the western Pacific. An example is the case of November 2006, when the Mirai Indian Ocean cruise for the study of the Madden-Julian oscillation–convection Onset (MISMO) observation campaign was operated in the equatorial Indian Ocean (Yoneyama et al. 2008). From this observation, Yoneyama et al. (2008) raised a speculation that the condition of sea surface temperature (SST) of colder SST anomaly along the western coast of Sumatra and warmer SST anomaly in the western Indian Ocean during MISMO was not suitable for maintaining organized cloud systems.

This speculation is consistent with a sensitivity study by a global cloud-system-resolving model (CSRM) by Miura et al. (2009). They demonstrated an important role of the zonal SST difference between the Maritime Continent (MC) and Indian Ocean. In their study, when the SSTs of the MC and tropical Indian Ocean were the same, although an organized cloud system developed at a proper timing in the central Indian Ocean, the system dissipated quickly without an eastward movement. This numerical simulation suggested the significance of warmer SST of the MC than the Indian Ocean in allowing an eastward-moving organized system observed during the MISMO campaign.

The behavior of the MJO is neither spatially nor temporally uniform. The MJO signals in convection are generally stronger in the eastern hemisphere than in the western hemisphere (e.g., Knutson and Weickmann 1987). In the eastern hemisphere, convective signals are strong over the Indian and western Pacific oceans, whereas the signals are much weaker over the MC (e.g., Salby et al. 1994). In the Pacific, the interannual variability of the MJO is affected by the interannual variability of SST associated with the El Niño–Southern Oscillation (ENSO), because SST in the central Pacific is warmer/colder during El Niño/La Niña events (e.g., Hendon et al. 1999). In return, enhanced activities of the MJO that precede ENSO warm events may be related to the development of ENSO (e.g., Zhang and Gottschalck 2002). An interannual variability of SST in the Indian Ocean, known as the Indian Ocean Dipole (IOD; Saji et al. 1999), also influences the MJO (Izumo et al. 2010, Wilson et al. 2013). The positive IOD, which is the condition that SSTs in the southeastern Indian Ocean and the western Indian Ocean are anomalously cool and warm, respectively, is thought to be one of the reasons for the abrupt termination of convective activity in the Indian Ocean during MISMO (Yoneyama et al. 2008).

These geographical preferences and interannual variations of the MJO are related to the fact that warmer SST is preferable for sustained convective activities in the tropics. One reason is that the warmer temperature of the planetary boundary layer over warmer SST provides less stable stratification in the lower troposphere than over colder SST. Another is that warmer ocean mixed-layer contains a larger amount of thermal energy, which can supply more humidity to the atmosphere through surface latent heat flux. The warmer SSTs of the Indian Ocean and the western Pacific provide climatological environments that induce time-mean large-scale upward motion in those regions as a spatial envelope through which organized cloud systems of the MJO are likely to proceed (Salby and Hendon 1994).

The spatial structure of the MJO also has a strong seasonality (e.g., Knutson and Weickmann 1987). Following the seasonal migration of the Intertropical Convergence Zone (ITCZ) and warm SST across the equator, the locations of the MJO convection also migrate in latitude (Nakazawa 1986; Salby and Hendon 1994). In austral summer, the strongest MJO...
signals reside to the south of the equator in association with the Australian summer monsoon (e.g., Hendon and Liebmann 1990b; Wheeler et al. 2009). In boreal summer, the strongest signals are located to the north of the equator, a region under the influence of the Asian summer monsoon (e.g., Yasunari 1980; Lawrence and Webster 2002). This seasonal migration is evident in the western Pacific, but not in the Indian Ocean (Zhang and Dong 2004).

Climatological conditions preferred by the MJO also change in intraseasonal timescale. Bellenger and Duvel (2007) compared horizontal patterns of the monthly means of the standard deviation in the 20–90-day bandpass-filtered outgoing longwave radiation (OLR). Figure 1 is almost equivalent to the Fig. 1 of Bellenger and Duvel (2007), but for 30–90-day bandpass-filtered National Ocean and Atmospheric Administration (NOAA) Interpolated OLR data (Liebmann and Smith 1996) from 1979 to 2013. Using the forward and backward discrete Fourier transformations, the time series of OLR at each grid point was processed as following

\[ \tilde{X}(n) = \frac{1}{N} \sum_{k=k_1}^{k_2} \left( \sum_{n=1}^{N} X(n) e^{-2\pi i kn/N} \right) e^{2\pi i kn/N}, \]

where \( n \) is time in day, \( X(n) \) is the original daily time series, and \( \tilde{X}(n) \) is the complex demodulated time series, \( k \) is wavenumber, and \( k_1 \) and \( k_2 \) are wavenumbers corresponding to the periods of 90 and 30 days respectively. Here \( N \) is the total number of days, where \( N = 12784 \). The 35-year climatological monthly mean of the magnitude of \( \tilde{X}(n) \) is plotted. Because the spatial distributions of the monthly mean variability of the intraseasonal component are quite similar between months during austral summer (December–March) and during boreal summer (May–October), the monthly means of atmospheric variables may be considered as climatological background for each season. In contrast, the patterns of boreal spring (April) and fall (November) may be regarded as transition patterns between austral summer and boreal summer. We examine further details of these quick transitions in Section 3.

An MJO event that was observed during an international observation campaign in the Indian Ocean coordinated under the program of the Cooperative Indian Ocean Experiment on Intraseasonal Variability in the Year 2011 (CINDY2011; Yoneyama et al. 2013) occurred during the quick transition period in November when the seasonal change is too fast to be separable from the intraseasonal timescale. We refer to this event as MJO2, following Yoneyama et al. (2013) and Moum et al. (2014). CINDY2011
was launched to collect observations during October 2011–March 2012 to advance our knowledge of physical processes of the MJO, especially, in its convective initiation. (Yoneyama et al. 2013). CINDY2011 included the U.S. participation under Dynamics of the MJO (DYNAMO; Gottschalck et al. 2013). Convective initiation of MJO2 was around 21 November 2011 in the western Indian Ocean, and an envelope of low OLR shifted to the MC in early December, but it barely reached the western Pacific.

The present study was undertaken to get insights into the impacts of seasonal change of SST on the MJO through investigation of MJO2 as an example. Fu et al. (2015) and Wang et al. (2015) have already reported successful simulations of MJO2 by coupled-atmosphere-only global models and a regional atmospheric model, respectively. They showed that MJO2 representation was improved significantly by using daily SST values instead of a constant SST value throughout the simulation as the bottom boundary condition. However, discussions about the seasonal SST change were absent in their work. As will be shown in Section 3, relatively fast warming of the oceanic area of the southeastern MC compared to that of the southeastern Indian Ocean in the intraseasonal timescale occurred simultaneously with the eastward movement of MJO2. We discuss the behavior of MJO2 that appears to be partly synchronized with the seasonal eastward shift of the precipitation center.

Using a global CSRM, we first conduct an ensemble hindcast of MJO2 along with the first MJO event (MJO1), which was also observed during the CINDY2011/DYNAMO campaign. The ensemble hindcast is conducted to confirm that the model can reproduce MJO1 and MJO2 marginally if realistic evolution of SST is given. Though far from perfect, this control simulation constitutes the basis of the following sensitivity tests. Next, we examine the influence of seasonal SST variation by comparing a pair of sensitivity runs forced by different SST conditions: one uses temporally constant SST and the other includes climatological seasonal change of SST. Then, to test the impact of the atmosphere–ocean coupling, we perform another pair of sensitivity runs with the mixed-layer ocean model by changing the mixed layer depth. The model and data are described in Section 2. Results are presented in Section 3. Summary and discussions are given in Section 4.

2. Experimental setup and data

The global CSRM used in this study is the Nonhydrostatic Icosahedral Atmosphere Model (NICAM; Satoh et al. 2008, 2014). Governing equations of NICAM are fully compressible and nonhydrostatic. The equations are discretized horizontally on the icosahedral hexagonal–pentagonal mesh using the finite volume method (Tomita et al. 2001). An upwind-biased semi-Lagrangian transport scheme is used for water substances (Miura 2007). The conventional terrain-following coordinate is adopted in the vertical. The model setups of the control simulation are very similar to those of Miyakawa et al. (2014), except that a bucket land surface model is used in place of MATSIRO (Takata et al. 2003), which is a more complicated land surface model, and time evolution of SST is not predicted by a mixed-layer ocean model but is prescribed externally.

Horizontal grid spacing is set to approximately 14 km to save computational time, which is degraded from the mesh sizes of approximately 3.5 and 7 km of Miura et al. (2007). Although the 14 km mesh is too coarse to resolve turbulent motions inside and around clouds, it has been verified to be tolerable in reproducing month-long large-scale evolutions of the MJO (Miyakawa et al. 2014). The number of the vertical layers is 38. The layer thickness gradually increases with height and reaches the model top at the altitude of 38 km. The model time-step is 30 s. No cumulus parameterization scheme is used, and the condensation processes are computed by the NICAM single-moment water 6 cloud microphysics scheme that prognoses six categories of hydrometeors (Tomita 2008). The subgrid-scale turbulence model is a modified version of the Mellor–Yamada scheme (Noda et al. 2010; Mellor and Yamada 1982; Nakanishi and Niino 2004). The radiation model is a two-stream radiative transfer scheme using a correlated k-distribution method (mstrnX; Sekiguchi and Nakajima 2008). The surface fluxes are computed by a modified version of the Louis scheme (Uno et al. 1995; Louis 1979).

The size of the ensemble hindcast is five and the initial dates of the hindcast are 12–16 October 2011 at 00 UTC. Initial conditions of the atmosphere are derived by horizontal and vertical linear interpolations from the National Centers for Environmental Prediction (NCEP) Final Operational Model Global Tropospheric Analyses data. It is confirmed that an addition of one more member initialized at 00 UTC on 11 October 2011 produces only minor differences for the control case. The initial conditions of the atmosphere are the same between the control run and the sensitivity runs. The control run uses a realistic time variation of SST, which is made by spatial and temporal
interpolation of weekly data of the NOAA Optimum Interpolation SST (OISST) version 2 on a one-degree grid (Reynolds et al. 2002). The SST settings of the sensitivity runs will be explained in a later section as necessary.

The integration period of all simulations is 60 days. That is, the simulation periods are from 00 UTC on 12–16 October to 00 UTC on 11–15 December. Note that no artificial techniques to nudge the model atmosphere to a realistic atmosphere are applied during the temporal integrations. Because information after the initial date is used in the form of the realistic time change of SST as the bottom boundary condition, this study is by no means an example showing the capability of NICAM in predicting MJO1 and MJO2. The period that is analyzed is from 00 UTC on 16 October to 00 UTC on 11 December, which is common to the five members.

As observational references, we analyze daily products of the NOAA Interpolated OLR data, the NCEP/National Center for Atmospheric Research (NCAR) reanalysis data, one-degree precipitation estimate of Global Precipitation Climatology Project (GPCP; Huffman et al. 2001), and the NOAA OISST on a quarter-degree grid. With the exception of precipitation, climatological means are computed for the years between 1979 and 2013 excluding 2011, the year of CINDY2011/DYNAMO. The year 2011 is excluded because we compare the condition of 2011 with the climatological mean to confirm that 2011 is not largely different from an average year. The climatological mean of precipitation is generated from data between 1997 and 2013 because the GPCP daily products are available after October 1996.

3. Results

3.1 Quick transitions in November–December and April–May

Transitions of the spatial pattern of the intraseasonal variability that are observed in the periods from November to December and from April to May are remarkably quick and distinct (Fig. 2). In mid-November, the intraseasonal variability of OLR is weak in the southern MC and the South Pacific convergence zone (SPCZ). The spatial pattern of the variability appears to be more like that of boreal summer. The intraseasonal variability in the southern MC and the SPCZ becomes stronger in late November. The signal

Fig. 2. 10 day mean of the daily climatology of the standard deviation of the 30–90 day filtered OLR (W m$^{-2}$) (a) from mid-November to early December and (b) from mid-April to early May.
in those regions intensifies further and the spatial pattern of austral summer eventually establishes in early December. The temporal change from April to May is roughly in reverse to that from November to December. The variability in the southern MC and the SPCZ weakens and that around the Philippines strengthens. The spatial pattern becomes close to that of boreal summer by early May.

In the southern MC, the transition between the weak variability in boreal summer and the strong one in austral summer roughly coincides with the transition between mean downward and upward motion in the mid-troposphere. Figure 3 shows the 35 year (1979–2013) climatology of the 10 day mean of pressure velocity at the 500 hPa level ($\omega_{500}$) from the NCEP/NCAR reanalysis data (Kalnay et al. 1996). The southern MC is under downward motion in mid-November, but the region of upward motion shifts southward to cover the Banda and Arafura Seas and the northern Australia in early December. The temporal transition of $\omega_{500}$ is less distinct for April–May, but the region of upward motion shifts northward and the Timor Sea is under the downward motion after late April.

Figures 2 and 3 seem to suggest that the transition between boreal summer and austral summer is more prominent in November–December than in April–May. This speculation is confirmed by comparing area averaged precipitable water (PW) and $\omega_{500}$ of the NCEP/NCAR reanalysis between the Indian Ocean (80°E–100°E, 5°S–5°N), the southeastern MC (120°E–140°E, 10°S–0°), and the western Pacific (150°E–170°E, 5°S–5°N) (Fig. 4). From December to March, PW of the southeastern MC is larger than 45 kg m$^{-2}$ and is comparative with or larger than those in the Indian Ocean and the western Pacific. After April, PW of the southeastern MC decreases continuously to approximately 40 kg m$^{-2}$ until August. Moreover, the driest season continues until the end of October. During the driest season the area averaged $\omega_{500}$ is nearly zero or slightly positive in the southeastern MC. The recovery of PW of the southeastern MC takes place very quickly from November to early December to approximately 47 kg m$^{-2}$ nearly within 40 days.

Although PW is less variable in the western Pacific than in the southeastern MC, it appears to experience a weak dry season after August and strong
upward motion recovers in November. From October to November, PW is the largest and upward motion is the strongest in the Indian Ocean compared to the southeastern MC and western Pacific. Interestingly, PW and $\omega_{500}$ values of the three regions converge to similar values in December and relatively closer values of PW and $\omega_{500}$ between the three regions are maintained until March. This result implies that the instabilities of the atmospheric columns of the three regions are equal during the period. Considering that December–March are months in which the spatial pattern of austral summer is usually evident in the intraseasonal variability in OLR, the destabilization of the atmospheric column of the southeastern MC can be an important factor to realize the MJO in austral summer.

3.2 The control case

Figures 5a and 5b show time evolutions of the observed OLR and zonal wind at the 850 hPa level ($u_{850}$) of the NCEP/NCAR reanalysis. As it has been reported by Yoneyama et al. (2013) and Gottschalck et al. (2013), there are two distinct eastward moving signals of low OLR in the period of our interest. One is MJO1 in late October to early November and the other is MJO2 in late November to early December. Similar to the finding of Nakazawa (1988), both MJO1 and MJO2 are not single unitary systems but appear to be packets comprising several major internal systems. Convective activity of MJO1 initiates near 60°E approximately on 21 October. A stronger signal approximately on 26 October moves eastward and passes through Sumatra Island (~100°E) approximately by 1 November and reaches around 120°E before the convective activity ceases. Enhancement of westerly winds follows the strong signal with a lag of several days in the Indian Ocean, but the phase of active convection and strong westerly coincides in the MC region. These phase relationships are consistent with the findings of Hendon and Salby (1994). After MJO1 has passed, convection is suppressed substantially over the Indian Ocean in early November because of the meridional transport of dry air (Maloney 2009) and the subsidence behind the strong convection of MJO1 (Nasuno et al. 2015). Convective activity of MJO2 begins to develop over the Indian Ocean in mid-November, and the activity comprises two eastward-moving signals (Moum et al. 2014). The first signal organized around 21 November and the second one lags the first by several days. The penetration of the second signal into the MC region approximately on 1 December sets off active convection in the region between 100°E and 140°E in early December. Along with the eastward movement of MJO2, the region of westerly winds extends eastward and the boundary between westerly and easterly winds shifts from approximately 80°E in late November to approximately 130°E in early December.

Figures 5c and 5d are the ensemble means of OLR and $u_{850}$ of the five members of the control case. The simulation seems to capture some important signatures of MJO1 and MJO2, although it is far from being satisfactory. Convective activity associated with MJO1 initiates at the correct timing approximately...
on 21 October and shifts to the west of Sumatra Island approximately on 1 November. However, continuous eastward propagation is not obvious, and strong convective activity in the MC approximately on 6 November is almost missing except for a slight enhancement of westerly winds. Alternatively, convective activity remains strong compared to the observation in the Indian Ocean in early November. Despite large differences between the observation and simulation in the first 30 days, the convective activity of MJO2 initiates in mid-November and the enhancement of westerly winds follows. The first eastward-moving signal develops approximately on 16 November, which is about five days earlier than the observation, and the second one forms several days later. The second signal moves into the MC approximately on 26 November, again about five days earlier than the observation, and the convective activity in the region between 100°E and 140°E is continuously enhanced after that. Following MJO2, the westerly region extends eastward and the boundary between westerly and easterly winds shifts to approximately 130°E as in the observation. Convection in the Indian Ocean weakens in early December when the region of active convection is in the MC.

The lack of the strong convection of MJO1 approximately on 6 November in the MC region is similar to the simulation initialized on 7 October of Fu et al. (2015) and the earlier initiation of MJO2 by approximately 5 days is also common with their simulation initialized on 4 November. These defects are readily detectable in Fig. 3 of Fu et al. (2015), although they did not point these explicitly. In contrast, the 76 day simulation of Wang et al. (2015) appears to successfully reproduce the strong convection approximately on 6 November. However, the eastern boundary of their regional model was at 120°E and the evolution of the atmosphere in the MC was strongly constrained by the realistic evolution of the atmosphere that is provided to the computational domain as boundary conditions. Their successful simulation indicates that a proper simulation of a westward
moving signal in the western Pacific may be a key to reproduce the enhancement of convection in the MC in early November. Initiation of MJO2 delays by several days in Wang et al. (2015). The reason for these earlier or later initiations of MJO2 has not been clarified.

Representation of MJO2 in each ensemble member is more ambiguous compared to the ensemble mean (Fig. 6). The best member may be the one initialized on 15 October with a distinct eastward moving signal in OLR and $u_{850}$ of MJO2 (Figs. 6b, 7a). The worst member may be the one initialized on 14 October, in which the convective activity sticks to the Indian Ocean and the westerly region does not extend eastward in early December (Figs. 6c, 7b). In spite of the large spreads, the overall impressions of the ensemble means are similar even if the best or the worst member is omitted (Fig. 8). This confirms that MJO2 in the ensemble mean is not just an insignificant artifact emerging from a single excellent case. The cause for the existence of the MJO2-like signal in the ensemble mean is postulated from the result that most

Fig. 6. Same as Fig. 5c, except for each member of the ensemble hindcasts. The initial date is 00 UTC on (a) 16 October, (b) 15 October, (c) 14 October, (d) 13 October, and (e) 12 October.
of the five members share the following key features: (i) strong convection occurs in the western Indian Ocean (~60°E) approximately on 16 November, (ii) convective activity enhances in the central Indian Ocean (~80°E) after 21 November, except for the member initialized on 16 October, (iii) convective activity in the eastern Indian Ocean (~100°E) also strengthens approximately on 26 November, (iv) convective activity persists over the MC in early December, except for the member initialized on 14 October.

3.3 Seasonal variations

The fact that the MJO2-like signal appears in the ensemble mean (Figs. 5c, d) while its representation is unsatisfactory in each member (Fig. 6) raises a question that the eastward shift of convection from the Indian Ocean to the MC might not only be determined by the internal dynamics of the atmosphere but also partly by the seasonal change of the external conditions such as SST and land surface temperature. Before performing sensitivity experiments addressing this question, temporal changes of SST, precipitation, winds, and humidity of 2011 and those of the climatology are overviewed by comparing the means of the first 29 days (16 October–14 November) and the latter 26 days (15 November–10 December) of the analysis period.

In the western Pacific and the MC, the time evolution of SST in 2011 (Figs. 9a, b) is similar to the climatology (Figs. 9d, e). In the prior period, SST is higher in the northern hemisphere than in the southern hemisphere. An area of SST warmer than 29°C extends nearly to the latitude of 20°N in the western Pacific, whereas SST colder than 28°C prevails to the east of Australia. In the latter period, the season-
ality of solar radiation leads to the southward migration of the higher SST region. The SST distribution in the latter period is more or less symmetric about the equator. The southeastern MC becomes warmer, and for 2011, SST in the Banda and Arafura Seas is higher than 30°C.

In contrast, the Indian Ocean was under the influence of a moderate positive phase of the IOD in October 2011 (Gottschalck et al. 2013). SST anomalies from the climatology in the southwest of Sumatra and Java islands are negative and that of the western Indian Ocean is positive in the prior period. A region of SST higher than 29°C covers the central and western Indian Ocean near the equator. In the latter period, the positive IOD is terminated as the negative SST anomalies offshore of Sumatra and Java islands diminish. As a result, the warming of the eastern Indian Ocean and cooling of the western Indian Ocean are more emphasized than the climatological seasonal change (Figs. 9c, f). The higher SST region migrates southward, but changes are smaller in the latitudes between 10°S and the equator compared to the southeastern MC. The zonal SST difference between the MC and the Indian Ocean is enlarged due to the seasonal changes of SST in the southern side of the equator.

In the prior period, the distribution of precipitation in 2011 is also similar to the climatology for the western Pacific and the MC. However, precipitation heavier than the climatology is observed in the central and western Indian Ocean due to the influence of the positive phase of the IOD (Figs. 10a, d). The maximum precipitation is located along the equator, except for longitudes around 60°E, where the precipitation maximum is split into northern and southern parts at approximately 10°N and 10°S, respectively.

In the latter period, precipitation in the clima-
tological mean decreases near the equator and its pattern becomes more symmetric about the equator in the Indian Ocean (Fig. 10e). Precipitation amount decreases in the central Indian Ocean and increases in the MC and the western Pacific, indicating that the center of precipitation has shifted eastward. The precipitation amount in the Banda and Arafura Seas areas exceeds 2 mm day$^{-1}$, and the high precipitation areas around ITCZ and New Guinea are connected. Similar temporal changes of the precipitation pattern are present in 2011 (Fig. 10b), but precipitation is strongly enhanced on the southern side of the equator in the Indian Ocean and over the ITCZ in the western Pacific (Figs. 10c, f). In the MC, the pattern of the temporal changes is similar between 2011 and the climatology, and the amount of precipitation increases mainly in the oceanic regions such as the South China, Java, Banda, and Arafura Seas. Interestingly, the spatial pattern of Fig. 10f in the MC resembles that of the composited OLR anomaly in December–February of Wheeler and Hendon (2004). If the MJO prefers the location of climatological convection (Salby and Hendon 1994), the convective activity of the MJO is more likely to be enhanced climatologically over the MC in the latter period than in the prior period.

Although precipitation rates are obviously overestimated, the overall distribution of precipitation of the ensemble hindcast is analogous to the observational data (Fig. 11). Other defects may include higher precipitation rates along the east coast of Somali Peninsula and over the ocean east of Philippines and the sustained precipitation near the equator in the western Indian Ocean. Temporal changes from the prior period to the latter period (Fig. 11c) roughly traces those observed in 2011 (Fig. 10c). In the latter period, the precipitation pattern in the Indian Ocean becomes more symmetric about the equator compared to the prior period; however, the precipitation decrease in the central Indian Ocean is not obvious. In the MC, precipitation is enhanced over the South China, Java, Banda, and Arafura Seas, and in northern Australia. Precipitation of the ITCZ and the SPCZ also increases. The eastward shift of the center
of convection has also occurred in the ensemble hindcast.

It is reasonable to consider that the temporal variations of precipitation are related to moisture variations. Then, the climatological time change of specific humidity and horizontal winds at the 700 hPa level are examined (Fig. 12). In the prior period, latitudinal peaks of specific humidity are on the northern side of the equator in the MC and the western Pacific, whereas they are split into the northern and southern peaks in the Indian Ocean (Fig. 12a). The humidity maximum is shifted westward of the SST maximum to locations of the upward branch of the Walker circulation. One reason for the westward shift is that the large zonal SST difference between the MC and the Indian Ocean forces westerly winds between those two regions. Because larger temperature gradient is in close equivalence with larger pressure gradient in the tropics, convergence of zonal winds occurs where westerly wind decelerates as the zonal SST gradient decreases. Another reason is that moisture transported by easterly winds along the ITCZ accumulates in the MC because of the low level convergence caused due to the mountain regions of Malay Peninsula and Sumatra Island. Wind fields near the equator display a Matsuno–Gill pattern (Matsuno 1966; Gill 1980), those to the west of the maximum humidity display the Rossby-wave response, and those to the east display the Kelvin-wave response. Lower specific humidity along the equator in the western and central Indian Ocean is associated with drier air from higher latitudes transported around the outside of the Rossby-wave response. The southeastern MC is also less humid than the surrounding regions because of the drier air transported by southeasterly winds from the area off the east coast of Australia.
Overall impressions of the latter period (Fig. 12b) are not largely different from the prior period (Fig. 12a). However, humid regions move southward, following the southward migration of the higher SST regions, and the distribution of humidity becomes more symmetric about the equator. The largest difference near the equator is in the southeastern MC, where specific humidity increases by more than 1 g kg\(^{-1}\) over the Banda and Arafura Seas (Fig. 12c). This moistening, which helps establish a suitable environment for sustained deep convection, is associated with the development of anomalous westerly or northwesterly winds in regions that lead to a weakening of the dry air intrusion from the southeast. Strengthening of westerly winds is not a limited feature of the southeastern MC but prevails in the region between the equator and 10°S. Westerly winds extend eastward to nearly 160°E on the equator. It has been perceived that the MJO prefers mean westerly winds in the lower troposphere (e.g., Inness et al. 2003; Zhang and Dong 2004). If this is the case, the mean wind field in the MC and the western Pacific is more suitable for the MJO in the latter period than in the prior period.

The temporal changes of the specific humidity and zonal winds in 2011 (Fig. 13a) are basically similar to those of the climatology (Fig. 12a). Moistening in the southern Indian Ocean and to the northeast of New Guinea in the western Pacific is a distinct feature of 2011, which is potentially associated with the passage of MJO2 during the latter period. It is noted that the southeastern MC and the SPCZ are also more humid than the climatology. The ensemble hindcast captures the strong moistening to the northeast of New Guinea, but the moistening of the southeastern MC and the southern Indian Ocean is weak (Fig. 13b). Increase of specific humidity over Australia is similar to the climatology (Fig. 12c), but this tendency is opposite to that of the observational data in 2011 (Fig. 13a). Although the magnitude of the zonal wind is exaggerated compared to those of the observational data and the climatology, the enhancement of anomalous westerly winds in the southern part of the equator is reproduced in the ensemble mean. It is confirmed that the ensemble simulation reproduced two seasonal changes that are conceived to be important in developing environmental conditions preferred by the passage of the MJO convection through the southeastern MC. One is the lower tropospheric moistening of the southeastern MC, which is associated with the warming of the Banda Sea and Arafura Sea areas. The other is the establishment of the mean westerly winds near the equator extending to the western Pacific.

Figure 14a compares the time evolution of SSTs in the Indian Ocean, the southeastern MC, and the western Pacific. The definitions of those regions are the same as Fig. 4. SST in the southeastern MC was lower than those of the other regions by approximately 1 K until early November, but it increases rapidly in mid-November to approximately 302.5 K, which is almost equal to SSTs in the other regions. In the NCEP/NCAR reanalysis data (Fig. 14b), PW has two distinct peaks in the Indian Ocean in late October and in late November, corresponding to active convection of MJO1 and MJO2. In the southeastern MC, PW increases steadily from approximately 40 kg m\(^{-2}\) on 16 October to approximately 55 kg m\(^{-2}\) on 10 December. This steady increase in PW is reproduced in the ensemble mean of the control run although the temporal variability is highly diminished in the later period of the simulation (Fig. 14c). PWs in the three regions become comparable in early December as with the case of climatological time change in Fig. 4.

### 3.4 Sensitivity tests

It seems reasonable from the control simulation and the data analyses to hypothesize that the eastward movement of convection of MJO2 is, at least
in part, caused by the seasonal change of SST. To test this hypothesis, we performed a pair of ensemble runs, changing the setup of SST from the control case as follows. In the first case, SST is fixed to the weekly mean values given on 12 October 2011 throughout the simulation periods to test if the MJO2-like signal could survive without the seasonal variation of SST. Hereafter, we refer to this case as FIXED. For the second case, the seasonal variation of SST along with the characteristics of the positive phase of the IOD in October 2011 is included in the model by the following procedure. First, monthly climatology of SST is determined by averaging monthly data of OISST from 1982 to 2010. We assume that the monthly climatology is representative of the 15th day of each month. Next, we estimate SST anomalies of 2011 on 12 October 2011. Climatological SST values on 12 October are computed by temporal linear interpolation of the monthly means. We subtract them from the weekly mean values on 12 October 2011 to generate SST anomalies reflecting the influence of the positive IOD of 2011. Then, the SST anomalies are added to every monthly mean. The model uses the final monthly product of each month to represent SST values of the 15th day of that month. SST values of each time step are interpolated temporally from the 15th day values to both sides of the date. We refer to this case as CLIM. Both FIXED and CLIM comprise the five members. The initial dates and the simulation periods are the same as the control case.

Figure 15 shows the time evolutions of the ensemble mean OLR of FIXED and CLIM. In both cases, the overall behavior is similar to the control case until early November (Fig. 5b), but convection in the western Indian Ocean is enhanced more strongly due to continued influence of the positive phase of the IOD. In FIXED, the convective center is still in the Indian Ocean even after mid-November, and does not shift to the MC (Fig. 15a). A pair of weak eastward moving signals is vaguely detectable at approximately 80°E approximately on 16 and 21 November. Nevertheless, those signals cannot proceed to the MC, and
westward moving signals are more prominent during the period from late November to early December in the Indian Ocean. In CLIM, a packet of eastward moving convection is organized approximately at 90°E (Fig. 15b) during 16 and 21 November, which is displaced eastward compared to the initiation location of the control case approximately at 60°E (Fig. 5b). The packet of convection passes through the MC in late November and reaches the western Pacific in early December. Although the timing is earlier than observed and signals in the western Indian Ocean are rather weak, some of the characteristic features of MJO2 are reproduced in CLIM. Convective activity in the eastern Indian Ocean (~100°E) strengthens approximately on 26 November and shifts eastward to the eastern MC approximately on 1 December. Moreover, the strong convection persists over the eastern MC in early December while convective activity is strongly suppressed in the Indian Ocean.

The largest difference in the temporal change of SST between FIXED and CLIM is observed in the southeastern MC (Figs. 16a, b). Although SST in the southeastern MC is constantly lower than SSTs in the Indian Ocean and the western Pacific in FIXED, the SST increases quickly in the period from mid-October to mid-November to the similar level to that of the Indian Ocean and the western Pacific in CLIM. In association with the difference in SSTs, PW in the southeastern MC becomes comparable to that of the Indian Ocean approximately after 15 November in CLIM (Fig. 16d), but the PW is constantly lower than the Indian Ocean in FIXED (Fig. 16c). Timing of this catching-up of SST in the southeastern MC with the Indian Ocean in CLIM is earlier compared to the control run. This result is consistent with the faster increase of SST in the southeastern MC compared to the control run and the earlier enhancement of convection in the MC (Fig. 15b).

The contrast between FIXED and CLIM supports the hypothesis that the eastward shift of the convection center in the period between late November and early December in 2011 is partly explained by the climatological seasonal change of SST. It is also suggested that while formation of a large-scale eastward moving packet of convection that has some signatures of MJO2 is not highly sensitive to spatial and temporal details of SST, the timing and the location of convective activity of CLIM could have been greatly biased compared to the control case due to the lack of those details.

Convective activity of CLIM is strongly concentrated in the regions around 60 and 100°E in early November (Fig. 15b). This feature can be associated with the very weak MJO-2-like signal approximately at 80°E in mid-November, because the upward motion in the western and eastern Indian Oceans accompanies subsidence in the central Indian Ocean and heating and drying due to the subsidence stabilizes the stratification in that region. Therefore, it is probable that the MJO2-like signal would become more realistic if the convective activity in the western and eastern Indian Ocean could be weakened. To test this speculation, we conducted another pair of ensemble runs that involved a mixed-layer ocean model.

The mixed-layer ocean model predicts the temporal evolution of SST at each horizontal cell using an energy budget equation. SST tends to decrease in convectively active regions due to reduced insulation and enhanced surface heat fluxes, whereas the reverse is the case in convectively suppressed
regions. These SST responses are likely to prevent convective systems from being localized in one location and promote their developments in other places. This mechanism may be one of the reasons for the improved representation of the MJO in climate models that allow atmosphere–ocean coupling (e.g., DeMott et al. 2014).

In the following experiments, SST values are nudged to those of CLIM with an e-folding time of 7 days. A pair of sensitivity runs is conducted with different settings of the mixed-layer depth. One is with a mixed-layer of 15 m, which is the same as Miyakawa et al. (2014), and the other is 25 m: these are hereafter referred to as MIX15 and MIX25, respectively. The mixed-layer depths of 15 and 25 m are the values on both sides of 20 m, which correspond to the depth that the maximum intraseasonal variations of precipitation takes place in the climate model simulation of Maloney and Sobel (2004). The other experimental settings are the same as CLIM.

The strengthening of the convective activity of MJO1 near 60°E after 21 October is stronger in MIX15 (Fig. 17a) than in the control and CLIM. The signal of convective activity proceeds into the MC and reaches approximately 120°E approximately on 6 November. This feature is only present in MIX15. Strong convection of MJO1 is not sustained over the MC in the other cases. The better representation of MJO1 in mix15 is consistent with the monthly predictability of the MJO that is demonstrated by NICAM, using the very similar settings as MIX15 (Miyakawa et al. 2014). The mixed-layer ocean model seems to enhance intraseasonal variability of convection in NICAM as has been the case with climate models.

Although the concentration of convection around 60°E and 100°E is mitigated, MJO2 in MIX15 behaves very differently from the observation (Fig. 5a). An eastward-moving signal is generated approximately on 1 November shortly after MJO1 and moves very slowly eastward to pass 100°E approximately on 21 November. The eastward moving speed is approx-
imately 3 m s$^{-1}$, which is slower than the typical speed of the MJO of 5 m s$^{-1}$. Because convection is suppressed in the west of active convection, the eastward moving signals in the western Indian Ocean are almost missing approximately on 21 November in MIX15. Because MJO1 and MJO2 in MIX15 appears to be the constituents of one packet of eastward-moving signal, the overall impression of MIX15 is quite different from CLIM before 21 November. After that date, MIX15 and CLIM begin to share some common features of the eastward-moving signal that proceeds into the MC and convection becomes active in the regions between 100°E and 140°E.

The ensemble mean OLR of MIX25 (Fig. 17b) is more like that of CLIM than MIX15, but the missing convective activity near 80°E in mid-November is recovered. MJO1 develops in the western Indian Ocean after 21 October and decays at 100°E in early November. The MJO2-like signal is generated approximately on 16 November in the western Indian Ocean and moves eastward to the MC approximately on 21 November. Convection in the MC becomes active at that timing, although convection in the Indian Ocean does not weaken sufficiently as seen in the region between 80°E and 100°E in early December. The difference between MIX15 and MIX25 suggests that shallower ocean mixed-layer produces stronger atmosphere-ocean coupling that enhances the organization of convection in the temporal and spatial scales of the MJO.

The recovery to the state of higher SST in the southeastern MC is slower in MIX25 than in MIX15 due to the larger heat capacity of the mixed-layer (Figs. 18a, b). For both MIX15 and MIX25, convective activity is suppressed and larger insolation of shortwave radiation is expected in the MC in the early November. However, it takes longer time to warm the mixed-layer ocean in MIX25 because a larger amount of energy is required per degree of temperature increase. Coherently, the recovery to the state of higher PW in the southeastern MC is also slower in MIX25 than in MIX15 (Figs. 18c, d). These results are consistent with the slower shift of the convective center from the Indian Ocean to the MC in MIX25 compared to MIX15 (Figs. 17a, b).

The largest difference between CLIM, MIX15, and MIX25 from FIXED is the presence of the shift of the convective center from the Indian Ocean to the MC during the 56 day period of the analysis. The convection center migrates from the Indian Ocean to the MC in late November when an eastward moving signal develops in the Indian Ocean at the timing similar to that of MJO2 and moves into the MC. These sensitivity experiments support the assertion that the climatological seasonal change of SST can force the eastward shift of the convection center from the Indian Ocean to the MC. The MJO2-like signal occurs in the seasonal transition period and seems to be a trigger of the sustained convection in the MC and the rapid drying of the Indian Ocean.

4. Summary and discussions

This study was directed to investigate the possible impacts of the seasonal change of SST on the MJO. We selected the MJO2 of CINDY2011/DYNAMO as a good sample event to examine for this purpose, because MJO2 initiated in the Indian Ocean in mid-November and moved onto the Western Pacific in December. Therefore, the lifetime overlapped the season with the rapid change in the climatological
spatial distribution of intraseasonal variability in the tropics. Because the monthly mean variability of the intraseasonal component of OLR changes too swiftly from November to December to be separated from the intraseasonal timescale of the MJO, we hypothesized that MJO2 was, at least in part, caused by the seasonal change of SST, which chiefly determines the geographical preference of the intraseasonal variability of convection. We performed a series of sensitivity experiments using different SST settings in a global CSRM, NICAM.

Although there was some dissemblance, the ensemble hindcasts initialized during 12–16 October at 00 UTC reproduced not only MJO1 but also MJO2 when realistic time evolution of SST was given for the bottom boundary condition (Fig. 5). Although the timing was about 5 days earlier, two prominent eastward-moving signals observed inside MJO2 emerged in the ensemble mean. This signal of MJO2 in the ensemble mean was robust despite the fact that each ensemble member was unsatisfactory (Fig. 6). This interesting fact suggested that MJO2 in the model was not just an artifact due to a single excellent case and that it is reasonable to consider the behavior of MJO2 to be restricted considerably by the external conditions. The important restriction was the prescribed SST in this case.

The data analyses showed that the eastward movement of MJO2 convection coincided with the seasonal change of the precipitation center, which shifts its position zonally from the Indian Ocean to the MC as it migrates in latitude from the north to the south of the equator (Fig. 10). Relatively faster warming of the oceanic area of the southern MC than the southern Indian Ocean provides an explanation for this seasonal migration of convection (Fig. 9). Until mid-November, convection over the MC is suppressed climatologically due to the southeasterly winds off the east coast of Australia, which transports drier air from region of cooler SST (Fig. 12a). From November to December, the climatological conditions change drastically. The southeasterly winds weaken (Fig. 18).
and upward motion strengthens (Fig. 3), and as a result, humidity in the lower troposphere increases substantially in the southern MC. The climatological mean precipitation increases mainly in the oceanic regions of the MC. This condition with higher SST in the MC than in the Indian Ocean is considered here to be the key of this MJO favorable environment.

A pair of sensitivity tests elucidated the influence of the changing climatology of SST from October to December for the eastward migration of convection (Fig. 15). When the SST values were fixed to those of 12 October 2011 throughout the simulation, convective activity persistently existed in the Indian Ocean. In contrast, convective center shifted from the Indian Ocean to the MC in mid-November when the SST values included the climatological change. The contrast between the two sensitivity runs provides evidence that the MJO2-like signal in the simulation, even after one month from the initial date, is not just a coincidence, but is at least partially caused by the seasonal change of SST.

The initiation of the MJO2-like signal in the ensemble hindcast using the prescribed climatological SST was biased eastward by approximately 30° in longitude. Moreover, the convection in the central Indian Ocean was much weaker than the observation in mid-November. Using the ocean mixed-layer model, this bias was mitigated. In the ensemble hindcast using the mixed-layer depth of 15 m, MJO1 and MJO2 appeared to be merged into one big packet of clouds that moved eastward at a slower pace than the typical speed of the MJO (Fig. 15a). This suggests that the air-sea coupling enhances the intraseasonal variability of convection in NICAM as in climate models. The hindcast using the mixed-layer depth of 25 m was more like the case with the prescribed SST values (Fig. 15b). This is due to the greater heat capacity of the ocean in the simulation with the mixed layer of 25 m compared to that with 15 m.

The MJO has been widely considered as an internal mode of the atmosphere or an atmosphere–ocean coupled mode that is superimposed on the climatological mean states. It has sometimes been assumed that the climatological mean state can be defined because its temporal change is much slower than the MJO. Data analyses or theories of the MJO that define a climatological background from time series of an atmospheric or oceanic variable and remove it from the original variable are implicitly based on the above perspective. However, this study shows that such perspective is inadequate for a certain type of MJO events such as MJO2 during the CINDY2011/ DYNAMO. In the period from October to December 2011, a climatological seasonal change is rapid enough to be comparable with the MJO; therefore, we cannot treat it as the background.

The first-order approximation for the distribution of convection in the simulation period is that convection in the Indian Ocean is stronger in the first half of the period and is weaker in the latter half. In contrast, convection in the MC is weaker in the first half and is stronger in the latter half. In the first half, SST in the southeastern MC is relatively cooler than the surrounding regions and convection is suppressed. In the latter half, SST in the southeastern MC becomes warmer than the Indian Ocean. Following the temporal change of SST, an atmospheric column of the southeastern MC becomes positively buoyant to the extent that is comparable with those of the Indian Ocean and the western Pacific. The eastward shift of the precipitation center indicates that a large-scale region of positive buoyancy also migrates from the Indian Ocean to the MC. Here an important point is that the shift cannot be discontinuous in the real atmosphere. Following this statement, it is conceivable that the atmosphere requires a weather system to enable rapid but continuous transition of the buoyancy center from the Indian Ocean to the MC.

In 2011, MJO2 seems to play the role. The air in the MC becomes more positively buoyant due to the rapid increase of SST than in the Indian Ocean, and a large-scale circulation that has an upward branch in the MC is strengthened as expected from the increased precipitation in that region. Thus, although NICAM can reproduce MJO2 with some reality, it does not necessarily mean that NICAM can simulate the internal dynamics of the atmosphere accurately for longer than one month. Instead it would be reasonable to consider that MJO2 readily emerges as a system to diminish instability generated by the seasonal evolution of SST. Our results lead to a hypothesis that a certain type of the MJO can be interpreted as a transition process, which strengthens large-scale upward motion in the MC and the western Pacific as a response to the transition of the region of large-scale positive buoyancy production following the warmer SST. This perspective, which regards a type of the MJO to be a weather system required to realize a regime shift caused by the temporal change of the bottom boundary condition, is consistent with the results of Hendon and Liebmann (1990a) that the onset of the Australian summer monsoon is associated with the traverse of the MJO. Further investigations will be needed to confirm whether this perspective is
applicable for other MJO events in the same season in other years.

The representation of MJO2 in each member of the NICAM simulations is far from satisfactory (Fig. 6), and improvements in our model are needed to enable further discussions. For example, changing the resolution of the SST data from weekly to daily might amend the earlier initiation of MJO2. Moreover, if there exists a wave-like atmospheric disturbance that is associated with the initiation of MJOs, better representation of its propagation speed could improve the timing of MJO2. Taking this into account, the increase of the number of vertical levels will be a possible revision to make for more accurate calculation of their propagation speed, because insufficient vertical resolution can lead to inaccurate representation of the equivalent depths. Needless to say, improvements on the physics packages of NICAM are important in addition to the above changes.

As it was demonstrated for MJO1 in this study, the simulations are not highly sensitive to the evolution of the SST for the simulations under one month. However, the pair of eastward-moving signals inside MJO2 was not captured in the climatological sensitivity runs that excluded the details of SST. This indicates that not only the correct initial conditions but also the realistic evolution of SST is necessary for the realistic simulation of MJO2. It is presumed that coupling of a good ocean model with NICAM is required for the prediction of the MJO for longer than one month.

Acknowledgment

This study is supported by Grant-in-Aid for Scientific Research (B-25287119) of the Japan Society for the Promotion of Science, Japan. We used the Earth Simulator of Japan Agency for Marine-Earth Science and Technology for all of the simulations in this study.

References

Bellenger, H., and J. P. Duvel, 2007: Intraseasonal convective perturbations related to the seasonal march of the Indo-Pacific monsoons. J. Climate, 20, 2853–2863.

DeMott, C. A., C. Stan, D. A. Randall, and M. D. Branson, 2014: Intraseasonal variability in coupled GCMs: The roles of ocean feedbacks and model physics. J. Climate, 27, 4970–4995.

Fu, X., W. Wang, J.-Y. Lee, B. Wang, K. Kikuchi, J. Xu, J. Li, and S. Weaver, 2015: Distinctive roles of air–sea coupling on different MJO events: A new perspective revealed from the DYNAMO/CINDY field campaign. Mon. Wea. Rev., 143, 794–812.

Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. Quart. J. Roy. Meteor. Soc., 106, 447–462.

Gottschalck, J., P. E. Roundy, C. J. Schreck III, A. Vintzileos, and C. Zhang, 2013: Large-scale atmospheric and oceanic conditions during the 2011–12 DYNAMO field campaign. Mon. Wea. Rev., 141, 4173–4196.

Hendon, H. H., and B. Liebmann, 1990a: A composite study of onset of the Australian summer monsoon. J. Atmos. Sci., 47, 2227–2240.

Hendon, H. H., and B. Liebmann, 1990b: The intraseasonal (30–50 day) oscillation of the Australian summer monsoon. J. Atmos. Sci., 47, 2909–2924.

Hendon, H. H., and M. L. Salby, 1994: The life cycle of the Madden–Julian oscillation. J. Atmos. Sci., 51, 2225–2237.

Hendon, H. H., C. Zhang, and J. D. Glick, 1999: Interannual variation of the Madden–Julian oscillation during austral summer. J. Climate, 12, 2538–2550.

Huffman, G. J., R. F. Adler, M. M. Morrissey, D. T. Bolvin, S. Curtis, R. Joyce, B. McGavock, and J. Susskind, 2001: Global precipitation at one-degree daily resolution from multisatellite observations. J. Hydrometeor., 2, 36–50.

Inness, P. M., J. M. Slingo, E. Gualiyardi, and J. Cole, 2003: Simulation of the Madden–Julian oscillation in a coupled general circulation model. Part II: The role of the basic state. J. Climate, 16, 365–382.

Izumo, T., S. Masson, J. Vialard, C. de Boyer Montegut, S. K. Behera, G. Madec, K. Takahashi, and T. Yamagata, 2010: Low and high frequency Madden-Julian oscillations in austral summer: Interannual variations. Climate Dyn., 35, 669–683.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y. Zhu, M. Chelliah, W. Ebisuzaki, W. Higgins, J. Janowiak, K. C. Mo, C. Ropelewski, J. Wang, A. Leetmaa, R. Reynolds, R. Jenne, and D. Joseph, 1996: The NCEP/NCAR 40-year reanalysis project. Bull. Amer. Meteor. Soc., 77, 437–471.

Knutson, K. R., and K. M. Weickmann, 1987: 30–60 Day atmospheric oscillations: Composite life cycles of convection and circulation anomalies. Mon. Wea. Rev., 115, 1407–1436.

Lawrence, D., and P. J. Webster, 2002: The boreal summer intraseasonal oscillation: Relationship between northward and eastward movement of convection. J. Atmos. Sci., 59, 1593–1606.

Liebmann, B., and C. A. Smith, 1996: Description of a complete (interpolated) outgoing longwave radiation dataset. Bull. Amer. Meteor. Soc., 77, 1275–1277.

Louis, J.-F., 1979: A parametric model of vertical eddy fluxes in the atmosphere. Bound.-Layer Meteor., 17, 187–202.

Madden, R. A., and P. R. Julian, 1971: Detection of a 40–50 day oscillation in the zonal wind in the tropical
Madden, R., and P. Julian, 1972: Description of global-scale circulation cells in the tropics with a 40–50 day period. J. Atmos. Sci., 29, 1109–1123.

Maloney, E. D., 2009: The moist static energy budget of a composite tropical intraseasonal oscillation in a climate model. J. Climate, 22, 711–729.

Maloney, E. D., and A. H. Sobel, 2004: Surface fluxes and ocean coupling in the tropical intraseasonal oscillation. J. Climate, 17, 4368–4386.

Matsuno, T., 1966: Quasi-geostrophic motions in the equatorial area. J. Meteor. Soc. Japan, 44, 25–43.

Mellor, G. L., and T. Yamada, 1982: Development of a turbulence closure model for geophysical fluid problems. Rev. Geophys. Space Phys., 20, 851–875.

Miura, H., 2007: An upwind-biased conservative advection scheme for spherical hexagonal–pentagonal grids. Mon. Wea. Rev., 135, 4038–4044.

Miura, H., M. Satoh, T. Nasuno, A. T. Noda, and K. Oouchi, 2007: A Madden-Julian Oscillation event realistically simulated by a global cloud-resolving model. Science 318, 1763–1765.

Miura, H., M. Satoh, and M. Katsumata, 2009: Spontaneous onset of a Madden-Julian oscillation event in a cloud-system-resolving simulation. Geophys. Res. Lett., 36, L13802, doi:10.1029/2009GL039056.

Miyakawa, T., M. Satoh, H. Miura, H. Tomita, H. Yashiro, A. T. Noda, Y. Yamada, C. Kodama, K. Miomoto, and K. Yoneyama, 2014: Madden–Julian Oscillation prediction skill of a new-generation global model demonstrated using a supercomputer. Nature Commun. 5, 3769, doi:10.1038/ncomms4769.

Moum J. N., S. P. de Szeoeke, W. D. Smyth, J. B. Edson, H. L. DeWitt, A. J. Moulin, E. J. Thompson, C. J. Zappa, S. A. Rutledge, R. H. Johnson, and C. W. Fairall, 2014: Air–sea interactions from westerly wind bursts during the November 2011 MJO in the Indian Ocean. Bull. Amer. Meteor. Soc. 95, 1185–1199.

Nakanishi, M., and H. Niino, 2004: An improved Mellor–Yamada level-3 model with condensation physics: Its design and verification. Bound.-Layer Meteor., 112, 1–31.

Nakazawa, T., 1986: Intraseasonal variations of OLR in the tropics during the FGGE year. J. Meteor. Soc. Japan, 64, 17–34.

Nakazawa, T., 1988: Tropical super clusters within intraseasonal variations over the western Pacific. J. Meteor. Soc. Japan, 66, 823–839.

Nasuno, T., T. Li, and K. Kikuchi, 2015: Moistening processes before the convective initiation of Madden–Julian oscillation events during the CINDY/DYNAMO period. Mon. Wea. Rev., 143, 622–643.

Noda, A. T., K. Oouchi, M. Satoh, H. Tomita, S. Iga, and Y. Tsushima, 2010: Importance of the subgrid-scale turbulent moist process: Cloud distribution in global cloud-resolving simulations. Atmos. Res., 96, 208–217.

Reynolds, R. W., N. A. Rayner, T. M. Smith, D. C. Stokes, and W. Wang, 2002: An improved in situ and satellite SST analysis for climate. J. Climate, 15, 1609–1625.

Saji, N. H., B. N. Goswami, P. N. Vinayachandran, and T. Yamagata, 1999: A dipole mode in the tropical Indian Ocean. Nature, 401, 360–363.

Salby, M. L., and H. H. Hendon, 1994: Intraseasonal behavior of clouds, temperature, and motion in the tropics. J. Atmos. Sci., 51, 2207–2224.

Satoh, M., T. Matsuno, H. Tomita, H. Miura, T. Nasuno, and S. Iga, 2008: Nonhydrostatic icosahedral atmospheric model (NICAM) for global cloud resolving simulations. J. Comput. Phys., 227, 3486–3514.

Satoh, M., H. Tomita, H. Yashiro, H. Miura, C. Kodama, T. Seiki, A. T. Noda, Y. Yamada, D. Goto, M. Sawada, T. Miyoshi, Y. Niwa, M. Harra, T. Ohno, S. Iga, T. Arakawa, T. Inoue, and H. Kubokawa, 2014: The non-hydrostatic icosahedral atmospheric model: Description and development. Prog. Earth. Planet. Sci., 1, 18, doi:10.1186/s40645-014-0018-1.

Sekiguchi, M., and T. Nakajima, 2008: A k-distribution-based radiation code and its computational optimization for an atmospheric general circulation model. J. Quant. Spectrosc. Radiat. Transfer, 109, 2779–2793.

Takata, K., S. Emori, and T. Watanabe, 2003: Development of minimal advanced treatments of surface interaction and runoff. Global. Planet. Change, 38, 209–222.

Tomita, H., 2008: New microphysics with five and six categories by diagnostic generation of cloud ice. J. Meteor. Soc. Japan, 86A, 121–142.

Tomita, H., M. Tsugawa, M. Satoh, and K. Goto, 2001: Shallow water model on a modified icosahedral geodesic grid by using spring dynamics. J. Comput. Phys., 174, 579–613.

Uno, I., X.-M. Cai, D. G. Steyn, and S. Emori, 1995: A simple extension of the Louis method for rough surface layer modelling. Bound.-Layer Meteor., 76, 395–409.

Vitart, F., 2014: Evolution of ECMWF sub-seasonal forecast skill scores. Quart. J. Roy. Meteor. Soc., 140, 1889–1899.

Wang, S., A. H. Sobel, F. Zhang, Y. Q. Sun, Y. Yue, and L. Zhou, 2015: Regional simulation of the October and November MJO events observed during the CINDY/DYNAMO field campaign at gray zone resolution. J. Climate, 28, 2097–2119.

Wheeler, M. C., and H. H. Hendon, 2004: An all-season real-time multivariate MJO index: Development of an index for monitoring and prediction. Mon. Wea. Rev., 132, 1917–1932.

Wheeler, M. C., H. H. Hendon, S. Cleland, H. Meinke, and A. Donald, 2009: Impacts of the Madden–Julian oscillation on Australian rainfall and circulation. J. Climate, 22, 1482–1498.
Wilson, E. A., A. L. Gordon, and D. Kim, 2013: Observations of the Madden Julian Oscillation during Indian Ocean dipole events. *J. Geophys. Res. Atmos.*, 118, 2588–2599.

Yasunari, T., 1980: A quasi-stationary appearance of 30 to 40 day period in the cloudiness fluctuations during the summer monsoon over India. *J. Meteor. Soc. Japan*, 58, 225–229.

Yoneyama, K., Y. Masumoto, Y. Kuroda, M. Katsumata, K. Mizuno, Y. N. Takayabu, M. Yoshizaki, A. Shareef, Y. Fujiyoshi, M. J. McPhaden, V. S. N. Murty, R. Shirooka, K. Yasunaga, H. Yamada, N. Sato, T. Ushiyama, Q. Moteki, A. Seiki, M. Fujita, K. Ando, H. Hase, I. Ueki, T. Horii, C. Yokoyama, and T. Miya-kawa, 2008: MISMO field experiment in the equatorial Indian Ocean. *Bull. Amer. Meteor. Soc.*, 89, 1889–1903.

Yoneyama, K., C. Zhang, and C. N. Long, 2013: Tracking pulses of the Madden-Julian oscillation. *Bull. Amer. Meteor. Soc.*, 94, 1871–1891.

Zhang, C., 2005: Madden-Julian Oscillation. *Rev. Geophys.*, 43, RG2003, doi:10.1029/2004RG000158.

Zhang, C., 2013: Madden-Julian oscillation: Bridging weather and climate. *Bull. Amer. Meteor. Soc.*, 94, 1849–1870.

Zhang, C., and J. Gottschalck, 2002: SST anomalies of ENSO and the Madden-Julian oscillation in the equatorial Pacific. *J. Climate*, 15, 2429–2445.

Zhang, C., and M. Dong, 2004: Seasonality in the Madden–Julian oscillation. *J. Climate*, 17, 3169–3180.