Cloud-permitting simulation of the October and November MJO events observed during CINDY/DYNAMO field campaign: Comparison with observations and budget of moisture and moist static energy

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Abstract

This study investigates the October and November MJO events observed during the CINDY/Dynamo (Dynamics of MJO) field campaign through cloud-permitting numerical simulations. The simulations are compared to multiple observational datasets. The control simulation at 9 km horizontal resolution captures the slow eastward progression of both the October and November MJO events in surface precipitation, outgoing long wave radiation, zonal wind, humidity, and large scale vertical motion. The vertical motion shows weak ascent in the leading edge of the MJO envelope, followed by deep ascent during the peak precipitation stage, and trailed by a broad second-baroclinic-mode structure with ascent in the upper troposphere and descent in the lower troposphere. Both the simulation and the observations also show slow northward propagation components and tropical cyclone-like vortices after the MJO passage. Comparison with synthesized observations from the northern sounding array shows that the model simulates the passage of the two MJO events over the sounding array region well.

Analysis of the moist static energy (MSE) budget shows that both advection and diabatic processes (i.e., surface fluxes and radiation) contribute to the development of the positive MSE anomaly in the active phase, but their contributions differ in by how much they lead the precipitation peak. In comparison to the observational data sets used here, the model simulation may have a stronger surface feedback and a weaker radiative feedback. The contribution from the total advection on the mesoscale to the moisture and MSE budget are small, although the individual terms (horizontal or vertical) can be significant. Gross moist stability in the simulations (similar to that derived from the observation datasets earlier by Sobel, Wang and Kim, 2014) shows a markedly increase from near zero values to ~ 0.8 during the active phase.
1. Introduction

The Madden-Julian Oscillation (MJO; Madden and Julian 1971, 1972) is an intra-seasonal weather phenomenon in the tropics. Because of its influence on global weather and climate (Zhang 2005), understanding, simulation, and prediction of the MJO has great scientific and societal value.

Modeling and prediction of MJO initiation in the Indian Ocean remains a long standing challenge. The field campaign Cooperative Indian Ocean Experiment on Intraseasonal Variability (CINDY)/Dynamics of the MJO (DYNAMO)/ARM MJO Investigation Experiment (AMIE), hereafter “DYNAMO”, was designed specifically to address this issue. The DYNAMO observational network captured four MJO events from October 2011 to March 2012 (Zhang et al. 2013, Yoneyama et al. 2014; Gottschalck et al. 2013).

New findings emerging from DYNAMO have been reported in a number of studies that document various aspects of these MJO events in the Indian Ocean, including the large scale structure of dynamic variables (temperature, zonal winds, humidity, and vertical motion) derived from the sounding network (Johnson and Ciesielski, 2013, Ciesielski et al. 2013), the cloud population observed from the ground based precipitation radars (e.g., Zuluga and Houze 2013, Powell and Houze 2013), the air sea processes regulating the atmosphere-ocean interaction (Moum et al. 2013), and the budget of moist static energy in the northern sounding array (Sobel et al. 2014). Attempts have been made in these observational studies to infer the roles of different components of the coupled atmosphere-ocean system for the MJO initiation and propagation. Yet the precise mechanism responsible for the initiation and propagation of these MJO events remains elusive.
A variety of theoretical models has been proposed in the past, emphasizing various
different processes: frictional boundary layer moisture convergence (Wang 1988, Wang and Rui
1992), surface enthalpy fluxes (Emanuel 1987, Neelin et al. 1987), radiative feedback (Hu and
Randall 1994, Raymond 2001, Bony and Emanuel 2005), a combination of both surface enthalpy
fluxes and radiative feedback as sources of moist static energy (Sobel et al. 2008, 2010),
moisture modes coupling temporal and spatial variation of moisture with dry dynamics (e.g.,
Sobel et al. 2001; Sobel and Maloney, 2012, 2013; Raymond and Fuchs 2002, 2007; Majda and
Stechmann 2009). Although none of these has been accepted as fully satisfactory by the research
community as a whole, the recent development of these MJO theories has converged attention on
the prominent role of free-tropospheric moisture. The idea that free-tropospheric moisture
variations are important to the MJO has been reinforced by many studies demonstrating that
numerical models in which deep convection is more sensitive to free-tropospheric moisture
produce better MJO simulations (e.g., Thayer-Calder and Randall 2009).

Dissecting the budgets of relevant moist variables, therefore, provides a useful
quantitative framework to diagnose the MJO. Moist static energy (MSE) or moist entropy, being
approximately conserved under phase change of water, are attractive. Working with the most
nearly conserved variables is often illuminating as a general rule; it is particularly appropriate if
the MJO is a diabatic mode which requires sources of MSE for its development and maintenance
(e.g., Sobel et al. 2008, 2010). On the other hand, working with the MSE (or entropy) budget can
be difficult. The largest terms in the temperature and moisture budgets cancel when they are
added to yield the MSE budget, reducing the amplitude of the apparent signal relative to the
noise. The MSE budgets computed from reanalysis datasets contain non-negligible residuals (e.g.,
Mapes et al. 2012, Kim et al 2014), which have been attributed to errors in the numerical models
and analysis increments in the data assimilation systems (e.g. Mapes and Bacmeister 2012). Yet another possibility is that the assumption - made by Yanai et al. (1973) and virtually all subsequent studies to our knowledge - that subgrid horizontal transports are small may not be valid. One of the aims of this study is to assess the role of mesoscale variability in the MSE budget of simulated MJO events in a high resolution regional modeling system.

MJO simulations have improved relatively slowly compared to other large scale flow features in the comprehensive climate models used for climate assessment (e.g., Lin et al. 2006; Hung et al. 2013). This may be partly because parameterization changes which would improve the MJO simulation tend to degrade some aspects of the mean climate, and are seen as undesirable for the purposes of assessments (Kim et al. 2011); nonetheless, it indicates a structural problem in the models. On the other hand, a number of recent process oriented studies have been performed using high resolution regional models (Ray et al. 2009; Holloway et al. 2013; Khouider and Han 2013; Hogos et al. 2012). The focused view offered by high resolution numerical modeling is attractive because convection and its coupling with large scale dynamics are better expressed than in lower resolution models in which convection is highly parameterized. An additional benefit of using a regional model is that error outside the region of interests may be minimized by prescribing boundary conditions directly from the analysis/reanalysis dataset, further improving the MJO simulation skill (e.g., Ray et al. 2010).

While this is not prediction skill - since it requires knowledge of the boundary conditions – it allows detailed analysis of a simulated MJO which is internally consistent (to the extent that the model budgets close), contains information on finer scales than a global climate model would, and, to the extent the simulations are successful, can bear close resemblance to the observed MJO.
In this study, a cloud-permitting regional modeling system, combined with several observational datasets, is used to (1) simulate the MJO events which occurred during DYNAMO and further document their multiscale variability, (2) validate the simulation of the mean state and intraseasonal variability using the observational datasets, (3) construct the budgets of moisture and moist static energy of the simulated MJO events. Observational validation in this study will serve as a stepping stone to future numerical experiments exploring the dynamics of these MJO events.

The rest of this paper is structured as follows. Section 2 has the details of the numerical model configuration, experimental design, and observational datasets. Section 3 contains the description of the observed and simulated MJO events. Section 4 describes the moisture and moist static energy budgets. Results are summarized in section 5.

2. Numerical experiments and observation dataset

2.1 Numerical model and experiment design

The Weather Research and Forecasting model version 3.4.1 (Skamarock et al. 2008) is used in this study. The ERA-Interim reanalysis (Dee et al. 2011) is adopted to construct the initial, bottom and lateral boundary conditions for the regional simulation. The lateral boundary consists of a narrow transition zone of 5 grid points, where the tendencies at the outmost grids are prescribed from ERA-Interim, and gradually merged with the tendencies generated by the model.

At the lower boundary, SST is updated every 6 hours over ocean surface using the ERA-I SST. Additional sensitivity experiments are also conducted in which time-averaged SST is used. Surface temperature over land is allowed to vary using the Unified Noah land-surface physics
scheme (Chen and Dudhia 2001). The surface skin temperature as a separate variable is
diagnosed using the surface skin temperature scheme (Zeng et al. 2005). This creates diurnally
varying surface temperature over both ocean and land. The RRTMG long wave radiation scheme
(Iacono et al. 2008) and the updated Goddard short wave scheme (Matsui et al. 2007) are used to
parameterize radiative processes. Vertical turbulent eddy mixing in the boundary and the free
troposphere is parameterized using the YSU PBL scheme (Hong et al. 2006). Surface fluxes are
treated using the Monin-Obukov scheme. The WDM double momentum microphysics scheme
(Lim and Hong 2010) from WRF V3.5.1 is adopted with additional modification on the limit of
the shape parameters and terminal speed of snow, based on preliminary tests and our experience
in cloud-resolving simulations. No convective parameterization scheme is used.

The horizontal and vertical advection schemes are 5th order and 3rd order accurate,
respectively. Moisture and condensate are advected using a positive definite scheme. The
implicit damping scheme is used to suppress unphysical reflection of vertically propagating
gravity waves in the top 5 km (Klemp et al. 2008). The 2nd order horizontal turbulent eddy
mixing is parameterized using the Smagorinsky first order closure, and evaluated in the physical
space.

The computational domain covers the equatorial Indian Ocean, from 20°S to 20°N, 48°E
to 120°E, as shown in Figure 1a. The vertical is discretized in to 45 vertical levels, with 9 levels
in the 1 km and a nominal top at 20 hPa. Horizontal grid spacing is chosen to be 9 km. This is
not adequate for individual convective cells, but can partly resolve organized cloud systems and
mesoscale circulations, as well as their upscale impact and coupling with large scale dynamics.

The model simulations start from 1 Oct, 2011. For the first 3 days, horizontal winds are
relaxed to ERA-Interim using spectral nudging with a zonal wavenumber 0 - 4 (>2000 km;
wavenumber is defined relative to the computational domain) and a meridional wavenumber 0 - 2 (>2000 km). No nudging is used for other variables. This nudging does not introduce additional sources/sinks for moisture, but tightly constrains the mean flow and horizontal convergence over the largest scales in the domain. This 3-day spectral nudging also allows the mesoscale to saturate in spectral space. After Oct 4, the model is integrated to Dec 15, 2011 without any further nudging. We focus on the free run period from Oct 4 until the end of the simulation.

2.2 Observational datasets

A number of observational datasets are used to study the two MJO events and to validate our simulations. The large scale dataset derived from the DYNAMO northern sounding array (Johnson and Ciesielski 2013, Ciesielski et al. 2013) provides time evolution of the vertical structures within the area covered by the array. Large scale horizontal flow fields are extracted from the ECMWF-Interim reanalysis dataset. Surface rainfall is taken from the 3-hourly 0.25° TRMM 3B42 rainfall product version 7A. The 8-km CMORPH (CPC MORPHing technique) rainfall dataset (Joyce et al. 2004), instead of 0.25 degree TRMM data, is used to construct the wavenumber-frequency diagram of rainfall, since high frequency waves are better represented in CMORPH. Radiative fluxes are composited from the 1x1-degree daily CERES (Clouds and the Earth’s Radiant Energy System, Wielicki et al. 1996, Lebo et al. 2012) SYN1deg data. Total precipitable water vapor estimated from microwave satellite observations - SSM/I (Special Sensor Microwave Imager) and TMI (Tropical Rainfall Measuring Mission's Microwave Imager) are also used for comparison with the simulations.
3. Comparison of the WRF simulation with observations

3.1 The spatial and temporal evolution of the MJO events

The time mean zonal wind in the equatorial northern Indian Ocean is westerly in the lower troposphere and strong easterly at the upper troposphere for October and November of 2011, as shown in Figure 1. The coherent MJO events observed during the DYNAMO were strongest over the NSA, and weaker south of the equator.

Figure 1 shows the Hovmoeller diagram of daily rainfall averaged between 5°S and 5°N from TRMM and the WRF simulation. Both model and observations clearly show two MJO events, starting from ~ 60°E and propagating eastward. The October event moves with a speed of ~ 5 m/s while the November event moves slightly faster. The eastward propagation of surface precipitation is greatly disrupted near the maritime continent (~ 100°E) in both the model and observation: the October event has little propagation across the maritime continent in rainfall (more so in the simulation), while the November event maintains coherent propagation in both model and TRMM after passage over Sumatra (east of 100°E). MJO propagation across the maritime continent is poorly simulated in most climate models (e.g., Hung et al. 2013). Despite its occurrence 10 days after the model initialization date, the simulation correctly captures the initialization of the first event on around Oct 16 at 60°E, and the November MJO event with a delay of ~ 2-3 days relative to observations.

The slow eastward progression of both MJO events in the zonal wind at 850 hPa (U850) is further shown in Figure 3. Eastward propagating westerlies associated with the October MJO events are apparent in both observations (Fig. 3b) and the WRF simulation (Fig. 3a). Rainfall associated with the October MJO event does not collocate with the westerly maximum, but occurs mostly in the leading edge of westerly regime bordering easterly (where there is
maximum low-level zonal convergence), similar to what is seen in the observations (Fig. 3a-b).

This low level convergence of zonal winds also corresponds well to the MJO precipitation for
the November event. The U850 and rain phase relation is consistent with the classic MJO
conceptual model in Madden and Julian (1972), but different from what occurs in the Western
Pacific (Houze et al. 2000), where rainfall tends to collocate with the westerly maximum. The
ERA-I also shows a strong westerly anomaly (~ 10m/s) around Nov 25-28 that seems to
propagate westward; a similar strong westerly anomaly is seen in the simulation Fig 3a without
westward propagation. The largest U850 discrepancy between model and ERA-I is a strong
westerly in the simulation after the November event which is not found in the observations. As
this is almost two months after initialization, we expect that any memory of the initial conditions
has been lost by this time and that the lower and lateral boundary conditions are the only
significant external influences on the solution.

Figure 4 shows the vertical structure of the 3-day averaged large scale omega, along with
precipitation and net column radiative cooling, all averaged between the equator and 5°N for the
late October MJO events. Observational validation of large scale vertical motion will be
discussed in a later section. During Oct 12-14, convection in the whole Indian Ocean is
suppressed with a weak precipitation maximum (~ 15 mm/day) near 66°E; and a region of
relatively weak but broad lower tropospheric ascent located between 65°E and 85°E. During Oct
15-17 when the MJO event is already underway, vertical motion shows a westward tilt of the
ascent region and a top-heavy omega with a distinct second baroclinic mode structure (i.e., ascent
in the upper troposphere and descent in the lower troposphere) between 60°E to 65°E, while the
leading edge of ascent (70°E -77°E) is bottom heavy. During Oct 21 - 23, a top-heavy omega
with a first baroclinic structure (ascent over the whole troposphere with a strong peak in the
upper troposphere) is collocated with the precipitation maximum between 67°E and 73°E; eastward of this region mostly bottom-heavy shallow ascent occurs, while westward of this region the second baroclinic mode structure in omega expands further. Overall, this structure - bottom heavy ascent in the leading edge of the precipitation center, deep ascent collocated with the maximum precipitation, and ascent/descent trailing the maximum precipitation - is maintained during the entire eastward progression of the October MJO event. The ascent/descent dipole in the vertical is coincident with a large region of strong relative radiative warming, presumably associated with stratiform cloud (Powell and Houze 2013); precipitation is not large over as large a region. The longitudinal extent of this structure grows after the passage of the precipitation maximum. By the period of Oct 30 - Nov 1, an area of more than 20 degrees in longitude (from 60°E to 84°E) is dominated by the ascent/descent dipole. Its western edge, unlike the precipitation peak, appears to expand even further westward...

This structure of vertical motion is a reminiscent of the cross-scale self-similarity in the cloud fields shared by mesoscale convective systems, 2-day waves, synoptic convectively-coupled Kelvin waves, and MJOs (Mapes et al. 2006). All have shallow clouds at the leading edge, followed by deep convective clouds, and trailed by stratiform processes. Similar to this self-similar progression of cloud fields, the WRF model simulates a sequence of evolution in the vertical structure of vertical motion. The dynamical implications of this structure will be further explored in the later section on the gross moist stability. The occurrence of shallow vertical motion and associated heating is particularly interesting. Its role has been discussed from different perspectives: (1) a shallow, bottom heavy vertical motion indicates that moist static energy is imported into the region, hence contributing to build up of moist static energy (e.g., Wang and Sobel 2012); (2) the circulation response to a shallow heating cannot efficiently
disperse energy away horizontally due to a slow phase speed of the shallow mode (Wu 2003); (3)
more parameterized shallow heating is beneficial for improving the MJO simulations in climate
models (e.g., Zhang and Song 2009). On the other hand, cloud-radiative feedback associated
with the borad stratiform processes may have play significant role.

The November MJO event (Figure 5) shows a similar time-longitude structure in the
vertical motion, but evolves at a slightly faster pace. The WRF model simulation also shows that
precipitation and omega associated with this MJO event do not propagate continuously, but with
a more stepwise eastward progression, e.g., from 21-23 Nov to 27-29 Nov.

High frequency convective events are apparent in 3-hourly surface precipitation data.

Figure 2 c and d show clearly fast westward-propagating signals in both CMORPH and WRF.
Some of these westward waves are prominent during the active phase of the two MJO events in
the open Indian Ocean, while others are geographically linked to large islands over the maritime
continent, e.g., Sumatra. These fast signals are in phase with diurnal variations, and closely
related to the so-called two-day waves (Zhou and Kang 2013, Tulich et al. 2011), which have
been observed by the precipitation Radar deployed at the Gan radar supersite (e.g., Zuluaga and
Houze 2013).

The high frequency variability in the time-longitude diagram of 3-hourly precipitation is
also apparent in spectral space. Figure 6 shows the wavenumber-frequency diagram, a regional
equivalent of the global Wheeler-Kiladis diagram (Wheeler and Kiladas 1999), for the 3-hourly
surface rainfall from CMORPH and WRF. A 16-day time segment with a 8-day overlap is used
for the Fourier transform in the longitude range 50°E to 95°E, and further averaged over the
latitude band from 5°S to 5°N. The background diagram is obtained by applying multiple passes
of a 1-2-1 filter on the wavenumber space – 40 passes in the present study. Because of the relatively short longitude and temporal windows, low-frequency and low-wavenumber variability are not resolved well. The primary features standing above the background red spectrum are Kelvin waves and the westward inertia-gravity waves, bounded by the dispersion curves for the theoretical n=1 inertial-gravity waves (WIGs) with equivalent depths between 12 and 50 meters, corresponding to a phase speed of 10 – 22 m/s for WIGs with a horizontal wavelength in the range between 1000 to 1500 km. This high frequency wave activity is similar to the quasi 2-day waves observed during TOGA-COARE (Takayabu et al. 1996).

3.2 Northward propagation of the two MJO events and tropical cyclones

In addition to the apparent eastward propagation, both DYNAMO MJO events also have a slow northward propagation component. Figure 7 is a latitude-time diagram of daily precipitation and 850 hPa relative vorticity, both averaged between 55°E and 90°E. Large-scale precipitation appears to migrate northward slowly, at a speed of ~ 1-2 m/s. Averaging over a narrow longitude band (e.g., 73-80°E) shows similar northward propagation, but may not capture the tropical cyclones. For both model and observations, the precipitation signature of the October MJO event can be tracked to Oct 15 near 5°S. South of that latitude, persistent precipitation can be seen in TRMM, indicating the presence of the ITCZ (Intertropical Convergence Zone) throughout the period. ITCZ precipitation is less persistent in the simulation.

Precipitation generally coincides with the local maximum of low level vortical motion during the active MJO phases, although the reverse is not true. The vortical signature of the MJO events is expected due to rotational dynamic entities such as Rossby gyres. Detailed synoptic analysis of these MJO events will be needed to better understand the nature of the rotational
dynamics. Several vorticity plumes can be found north of 5°N and extending to the extratropics, e.g., in late October, early November, and late November, in both observations and the WRF simulation. These rotational entities with local precipitation maxima are signatures of tropical depressions or named tropical cyclones. We will refer to these features as tropical cyclones, while not rigorously categorizing them into the intensity-based specific TC types, since the intensities simulated do not correspond closely to those in the observations; the simulation tends to produce weaker TCs than those observed at the horizontal grid spacing dx = 9km.

The left column of Figure 8 shows maps of total SSM/I precipitable water, TRMM rainfall, and ERA-I 850 hPa wind vectors at three dates close to those on which the vorticity plumes are best developed: Oct 27, Nov 3, and Nov 27. The same variables from WRF are shown in the right column of Figure 8. On Oct 27, while none of the best track data report even a tropical depression, strong low level rotation and local precipitation maximum centered at (62E, 10N) in the Arabian Sea indicate TC-like structure. This is the remnant of the MJO Rossby gyres that migrated northward, bringing significant rainfall, but without transforming into a full-blown TC.

The November 3 event shown in the mid row of Figure 8 later developed to Tropical Storm Four from Nov 7 – Nov 9 in the western Arabian Sea, as reported in the JTWC (Joint Typhoon Warning Center) best track dataset. The WRF model also shows a TC-like structure at this day but simulates much less precipitation than is found in SSM/I. Nevertheless, the system shows a concentrated moisture anomaly along with precipitation and relative vorticity maxima.

The late November TC event, reported as Tropical Storm Five (TC05A) in the JTWC best track dataset, is sampled in the bottom row of Figure 8. This event was well forecast by several numerical models (e.g. Fu et al. 2013). On November 27, the center of TC05A had
reached around 12°N in the eastern Arabian Sea. At this time the WRF shows a rotational structure and axisymmetric rainfall distribution associated with the Rossby gyre, but it is larger and the center is located ~5°N, far south of that in the observations. Three days later, the WRF model does show the axisymmetric TC structure having moved to ~12°N off the West coast of India. The delay of this TC event in the WRF simulation may be due to the delay of the simulated November MJO event, as discussed before and also shown in the time series of precipitation (Figure 9 below).

Overall, the tropical cyclone events observed during the DYNAMO period appear to be closely related to the northward propagation of the two MJO events. Many days into the simulation, the WRF model is still able to simulate the formation of TC like events, although their intensities and tracks do not match observation exactly. These results suggest that improved simulation of the MJO could in principle extend the lead time at which skillful TC forecasts can be made.

3.3 Comparison with the northern sounding array data

Johnson and Ciesielski (2013) show that the October and November MJO events maintain coherence in the northern sounding array region, while occasional passage of synoptic-scale disturbances disrupts the MJO signals over the southern sounding array region. The present WRF setup does not simulate well each individual synoptic event in the southern sounding array, but does much better for the large scale MJO events. In the following, we compare the temporal evolution and vertical structure of various quantities from the model simulation in the NSA region against the sounding array observations.
Time series of OLR averaged over the region is shown in Figure 9. CERES top-of-atmosphere (TOA) outgoing long wave (OLR) shows a significant reduction from ~290 W/m² to 180 W/m², i.e. by ~100 W/m², during the October MJO event, and by ~150 W/m² during late November. Reflected TOA shortwave radiation shows a significant increase from 50 W/m² in the suppressed phase to more than 200 W/m² in the active phase. This dramatic reduction in both OLR and downward shortwave is a marked feature of the MJO events in the Indian Ocean. The radiation anomalies may lead to radiative-convective instability, as several authors have postulated may be important to the MJO (Lee et al. 2001, Raymond 2001, Bony and Emanuel 2005, Sobel et al. 2008). The simulated OLR anomaly during late October and November is about 10 W/m² less than CERES at the peak of the MJO convective phases. The reflected short wave radiation at top of the atmosphere (Figure 9c) is 10 W/m² less than CERES. At TOA the long wave and short wave compensate each other to a great degree in both CERES and the simulation. Net column radiative heating will be further discussed later in the context of the atmospheric moist static energy budget.

The vertical structures of the large scale vertical motion, zonal winds, temperature, and humidity fields will be discussed below. Figure 10a shows the simulated large-scale vertical motion, $W$ (cm/s) averaged over NSA. The large-scale $W$ derived from the NSA horizontal winds using mass conservation (Johnson et al. 2013, Figure 10b) shows multiple episodes of ascent in the October MJO events, and two strong such episodes in the November events. The WRF model shows similar behavior. Vertical motion in both the model and NSA sounding data has a first baroclinic mode structure during the MJO active phase followed by the development of a top-heavy second baroclinic mode structure, though somewhat more so in the simulation than in the
observations. This time progression also resembles the time-longitude diagram in Figures 3 and 4.

The simulation differs from the sounding array data in several aspects: as with precipitation, the maximum in $W$ is larger in the simulation than in the observations for the October event, but weaker for the November event. The time-averaged vertical profile of $W$ from the NSA is top-heavy, similar to $W$ from the TOGA-COARE in the western Pacific, with a peak value of $\sim 1.2$ cm/s. The WRF-simulated mean values of $W$ agree well with the observations, (Figure 10c), but with an upper-tropospheric peak at slightly lower altitude ($\sim 400$ hPa), and slightly weaker amplitude. The standard deviation (Figure 10d) also shows similar vertical structure with a local maximum in the upper troposphere.

Low-level westerly burst is one of the defining characteristics of the MJO (Lin and Johnson 1996). Figure 11 shows the model simulated zonal wind (Fig. 11a) and the zonal wind from the NSA observations (Fig. 11b). During the first half of October, the wind in the lower troposphere over the NSA is dominantly westerly, becomes easterly prior to the October MJO event, and turns back to westerly during the MJO active phase (late October and early November). After the rainfall peak in the last 2 weeks of October, the westerlies continue to strengthen until the middle of November. The simulated transition from easterly to westerly in the lower troposphere starts from near the surface around Oct 20th and then deepens, reaching 400 hPa in the first week of November, while the observations show a similar but less distinct gradual deepening of the westerlies over a $\sim 5$-day period near Nov 1st. During this period of westerly development, the upper tropospheric easterlies also strengthen. After the passage of the October MJO event, easterlies prevail in both model and observations until $\sim$ Nov 20th, when the November MJO event arrives at the NSA. The lower-tropospheric westerly wind burst is nearly
10 m/s in observation, and greater than 10 m/s in the model. The NSA region remains westerly until the middle of December. The averaged zonal wind profile features a peak of 10 m/s easterly wind at ~ 200 hPa, with weak westerlies in the lower troposphere, as seen in both model and observations (Fig. 11c). The standard deviation (Fig. 11d) has a peak at ~ 150 hPa in both model and observations, but the model’s standard deviation is larger from 900 hPa to 400 hPa.

**Figure 12** shows the vertical structure of temperature anomalies, computed by subtracting the time mean of temperature profiles spatially averaged over NSA. A positive temperature anomaly can be seen through the depth of the troposphere from Oct 10 to Oct 15, and persists in the upper troposphere during the MJO active phase (the last two week of October). A negative temperature anomaly is first found in the lower troposphere during the last two weeks of October while the lower-tropospheric westerlies are still weak. Afterwards, the negative temperature anomalies continue to extend to the upper troposphere, and show a tilted structure in the time-height plot with a minimum in the 200 – 400 hPa layer. The tilted positive temperature anomaly prior to the MJO rainfall peak and the negative temperature anomaly after it, as described for the October event, are also seen for the simulated November MJO event, though evolving at a faster pace. The same structural evolution is seen in the sounding array observations, but somewhat less coherently. There is also significant temperature variability near the tropopause (100 – 150 hPa), which is likely due to vertical propagation of Kelvin waves or inertia-gravity waves. The simulated temperature anomalies generally agree well with observation in vertical structure and in the vertical structure of standard deviation (Fig. 12 d).

**Figure 13** shows relative humidity with respect to ice, computed from the profiles of temperature and water vapor averaged over the NSA. The dryness of the troposphere during the suppressed MJO phase is evident in both WRF simulation and NSA observations. Gradual lower
tropospheric moistening occurs prior to the MJO events (Oct 10\textsuperscript{th} – 20\textsuperscript{th}, Nov 15\textsuperscript{th} – 20\textsuperscript{th}) when
the low-level wind is easterly (Fig. 11) and the temperature anomaly is positive (Fig. 12). The
subsequent drying after the active phases is less dramatic in the WRF simulation than in
observations. The time mean relative humidity also shows a local maximum near the tropopause,
between 100 and 200 hPa (Fig. 13 c), ~80\% in NSA and 60\% in WRF. This local peak also
persists during the suppressed MJO phase. This feature may be related to the anomalous cirrus
activity found in MJO composites of satellite observations (e.g., Virts and Wallace 2010, Del
Genio et al. 2012). A local maximum is also seen in the standard deviation of relative humidity
in NSA, but it is much weaker in WRF. The reduced variation in relative humidity between 100
– 200 hPa in WRF indicates that WRF has difficulty in capturing variability associated cirrus
near the tropopause.

In short, comparison with observations in the northern sounding array indicates that the
WRF simulation captures the passage of two MJO events over the NSA region with high fidelity.
A prominent feature from this analysis is that the lower troposphere leads the upper troposphere
in nearly all variables (vertical velocity, moisture, temperature anomaly, and zonal wind),
indicating a westward tilt spatial structure during the two MJO life cycles. This spatial tilt agrees
with what is seen in MJO composites (e.g., Kiladis et al. 2005) in some variables (zonal wind,
temperature, humidity), suggesting they are not case specific but are general enough to be
considered as a common features of MJO events. In the following, we continue to analyze these
MJO events in the NSA region, with a focus on quantification of moisture and moist static
budget.

3.4 The role of daily SST
In this section we assess the role of the imposed time variations in SST on the simulated MJO events. It may be argued that since the MJO is to some extent a coupled atmosphere-ocean phenomenon, imposing daily SST is inappropriate and a coupled modeling approach is preferred (e.g., Seo et al. 2014). However, our simulations already incorporate time-varying lateral boundary conditions which are influenced by atmospheric observations from outside the domain; that exterior atmosphere is coupled to the atmosphere within the domain as much as the SST within the domain is. These are not forecast model runs, as those assessed in Ling et al. (2013), but simulations designed to produce a set of three-dimensional atmospheric fields which are both internally consistent (given the model dynamics, physics and numeric) and as consistent as possible with the observed atmospheric evolution during the simulation period. The observed daily SST, like the lateral boundary conditions, is a mathematically well-posed condition to help achieve consistency with observations.

It is nonetheless of interest to know how strongly the daily SST influences the simulated MJO events. We investigate this by conducting a numerical experiment in which daily SST is replaced by time independent SST averaged from October to the end of December. Figure 14b shows the 850 hPa zonal winds and surface precipitation from this experiment. The October MJO event is reasonably well reproduced in these two variables with time independent SST, although the MJO signature is weaker and its eastward propagation is faster. For the November MJO event, the MJO can barely maintain its strength. Thus this experiment seems to indicate that daily SST plays different roles in the two MJO events: it is more important in the November event than the October one.

However, the November MJO event occurs later in the simulation, so it is possible that the inferior simulation of it compared to the October one could be a consequence of the drift of
the model solution away from the observed atmospheric state, rather than any inherent difference
between the two MJO events. To address this issue, we perform another pair of simulations
initialized at 00 UTC on November 10 with horizontal winds nudged toward reanalysis in
their lowest spectral components until 00 UTC November 13. These two otherwise identical
simulations differ in one aspect: one uses daily SST, and the other uses monthly mean SST.
Results from these two simulations (Figure 14 c and d) differ substantially: the November MJO
event is well reproduced in the daily SST event, but poorly simulated with monthly mean SST.
This is similar to the results from the two experiments initialized from Oct 1st (Figure 14 a and
b). Together, these two pair experiments suggest that daily SST plays an important role for the
November MJO event, but much less so for the October event. A deeper understanding of this
difference between the two events will require further research.

4. Budget of water vapor and moist static energy

The moist static energy (MSE) budget of the MJO has been explored in both observations and
simulations (e.g., Maloney 2009; Raymond and Fuchs, 2009; Kiranmayi and Maloney, 2011; Wu
and Deng 2013, Anderson and Kuang 2012). Sobel et al. (2014) recently constructed the MSE
budget of the DYNAMO MJO events studied here using both observation and the ERA-I
reanalysis data, and showed that both the large scale horizontal and vertical advection term had a
negative contribution to the MSE budget during the active phases.

Most of these studies have used data with grid spacings of about 1 degree. However, the
budgets on these scales are not always closed, and the residual from the MSE budget is often of
the same order as the resolved terms. The possible contribution from mesoscale fluxes has not
been explicitly quantified in the MSE budget, due to the lack of observations adequate to the
purpose. We can study this within our simulation due to the relatively high resolution and internal consistency; the results constitute at least a plausible hypothesis about the role of mesoscale fluxes in observations. We will show that the relative magnitudes of individual advection terms (i.e., horizontal and vertical) may change significantly as horizontal resolution is degraded; however, the total nonlinear advection of MSE by mesoscale disturbances is robust to change in resolution.

4.1 Budget of water vapor

We first consider the budget of column integrated water vapor, which is written as

\[
\left< \frac{\partial q}{\partial t} \right> + HADV_q + VADV_q = E - P \tag{1}
\]

where \(\left< \right>\) denotes a mass weighted vertical integral; \(HADV_q\) and \(VADV_q\) are horizontal and vertical advection at resolved scales by the model; \(E\) is surface evaporation; \(P\), precipitation.

Note that we have neglected the diffusion process which smooths out the moisture field, therefore it is not a source/sink in a global domain but it may still be a net source/sink at local area. As shown below, the r.h.s and l.h.s of equation (1) agree quite well, indicating that this omission in (1) is justified.

The two nonlinear advection terms at the resolved scales have contributions from both mesoscale and large scale, and may be further written as:

\[
HADV = - \left< \frac{\partial q}{\partial t} \right> - \left< v_n \nabla q_v \right> - \left< \omega \frac{\partial q}{\partial p} \right>, \quad
VADV = - \left< \frac{\partial q}{\partial t} \right> - \left< \omega \frac{\partial q}{\partial p} \right> - \left< \omega' \frac{\partial q'}{\partial p} \right>.
\]
Here, the spatial scale separation of a quantity $A$ is written as $A = \overline{A} + A'$, where overbars denote spatial average, and primes ' denote eddies. These two nonlinear advection terms are commonly evaluated using data at a horizontal resolution on the order of 1 degree when reanalysis data is used, thus contribution from mesoscale disturbances cannot be assessed. To facilitate interpretation in terms of large scale dynamics, spatial and temporal averages of these budget terms are often computed over finite areas covering a few hundreds of kilometers, similar to or smaller than the DYNAMO sounding network. Taking advantage of the high resolution modeling data, we now evaluate the mesoscale contribution by performing an “averaging” operation before computing each budget term, as is typically implicitly done for the large scale MSE budget using reanalysis datasets. We use a range of horizontal scales for this averaging operation: $L = 9, 18, 63, 135$, and $225$ km. For $L = 9$ km, we compute each budget term directly on each model grid cell, while we average all the variables over multiple adjacent grid cells for other values of $L$. Skamarock (2004) demonstrated that 6-7 times the model horizontal grid spacing (63 km), is the effective scale above which numerical diffusion is considered to be unimportant.

Figure 15a shows the moisture budget over NSA with $L = 9$ km at 3-hourly temporal sampling. The dominant terms are vertical advection and rain, as expected, but the horizontal advection is also remarkably large – it is more than 15 mm/day during the active phase. Sobel et al. (2014) and Kern et al. (2013) demonstrated that large scale advection is primarily responsible for the strong horizontal advection of observed MJO events. As shown below, the simulated large horizontal advection can also be attributed to mesoscale fluxes and cancel out by vertical advection. Surface evaporation is smaller than any of these 3 terms and is not shown here. The role of shallow and deep convection in the moistening may also be deduced by comparing Fig.
15 with the vertical structure of large-scale vertical motion. Moistening of the lower troposphere around Oct 10-20 (Fig. 13), when rainfall does not exceed 15 mm/day, can be attributed to vertical advection associated with the shallow circulation (Fig. 4 a-c), while vertical moisture transport due to deep convection and stratiform processes (Fig. 4 d-f) dominates the moisture budget during the heavily raining period Oct 20-30. The agreement between the directly computed local change of column integrated water vapor on the left hand side of equation (1) and the total tendency terms on the right hand side is good despite the relatively coarse temporal resolution in this calculation. The water budget over NSA at the effective resolution ($L = 63$ km) in Figure 15 b shows similar temporal evolution of each term, but vertical advection does not exceed rain as it did for $L = 9$ km, and horizontal advection is also notably smaller, less than 10 mm/day over the MJO active phases, which is similar to what is computed from the sounding array budget (Johnson and Ciesielski 2013).

The two nonlinear advection terms computed with different averaging $L$ are shown in Figures 15 b and c. The magnitudes of both horizontal and vertical advection evidently decrease with $L$. There is also a sign of convergence with increasingly larger $L$. This suggests that mesoscale advection/fluxes of moisture (the difference of advection terms computed with large $L$ compared to those computed with small $L$), cancel to a great degree. Indeed, total advection, $HADV + VADV$, as shown in Figure 15 e is largely independent of $L$, further demonstrating this cancellation of horizontal and vertical mesoscale fluxes, and their relatively small contribution to the large scale moisture budget.

### 4.2 Budget of moist static energy

The column-integrated budget for MSE may be written

$$\frac{\partial e}{\partial t} = HADV_e + VADV_h + R + SH + LH + D_h$$

(2)
where \( e \) and \( h \) represent frozen moist enthalpy \( (e = c_p T + L_v q_v - L_v q_i, \text{ water vapor, } q_v, \text{ ice}) \), and frozen moist static energy \( (h = e + g z) \), respectively. \( R \) refers to column net radiation; \( SH \), surface sensible flux; \( LH \), surface latent flux; and \( D_h \) subgrid diffusion and numerical diffusion. The left hand side is the local tendency of the column-integrated frozen moist enthalpy, which will also be referred as the storage term. \( VADV_h \) is the vertical advection of moist static energy; \( HADV_e \) is the vertical integral of the horizontal advection of moist enthalpy. Together, these two terms will be referred as advection terms in the MSE budget, while \( R + SH + LH \) will be referred to as the diabatic terms. (The “diabatic heating” which would appear in a potential temperature or dry static energy equation due to condensation of water vapor is not present in (2) because MSE is conserved under phase change of water.) The conversion from kinetic energy to potential energy, although small, is accounted in \( HADV_e \), following equation 10.3 in Neelin (2007). The diffusion term is typically neglected. Validity of the MSE equation (2) seems to be questionable for non-hydrostatic processes simulated in the WRF model, since the MSE equation is strictly valid in the hydrostatic limit. However, it is still appropriate at least for large \( L \)’s, at which scale the non-hydrostatic effect is small.

**Figure 16** shows the column-integrated MSE budget, including vertical and horizontal advection, surface fluxes, column net radiative fluxes, and the storage term, all averaged over the region of the NSA. The surface enthalpy fluxes and radiation terms are computed as the averages of direct model output. This operation is independent of the horizontal average scale \( L \). **Figure 16 a** shows these diabatic terms and their sum. Both terms increase during the onset of the MJO phases. This result agrees with observations for the November MJO event in that both surface fluxes and radiation contribute to the buildup of the MSE anomalies (Sobel et al. 2014). However, the large scale surface fluxes from the OAFlux product remain nearly constant during
the October MJO event, while the WRF model shows a dramatic increase of surface fluxes in the
middle of October.

The advection terms in the MSE budget are shown in Figure 16 b, evaluated at $L = 63$
km. The horizontal advection $HADV_e$ is close to 200 W/m$^2$ (Figure 16 b) in the active phase,
larger than that derived from either the sounding network or the ERA-I dataset (on the order of
100 W/m$^2$ during the MJO active phases, see Sobel et al. 2014). Vertical advection is positive ~
10 days before rainfall peaks due to shallow circulation, becomes negative after the MJO onset,
and reaches the minimum after the rainfall peaks. This negative $VADV_h$ indicates energy export
by the top-heavy large-scale omega profile. The sum of $HADV_e$ and $VADV_h$ (Figure 16 c) shows
a broader period of positive values before the rainfall peaks, contributed by advection terms 5-10
days before the rainfall peak, and by the diabatic terms within a few days before rainfall peak,
becoming negative as the MJO active phase moves out of the NSA region. The different
contribution of advection and diabatic terms at different lead times relative to the rainfall peaks
is also observed from the budgets derived from the sounding array (Sobel et al. 2014).

There is a strong cancellation between the diabatic terms (blue) and the advection terms
(red). The sum of all the tendency terms excluding diffusion (Figure 16 c) is in some qualitative
respects similar to the advection terms alone, but more positive immediately before rainfall peaks
due to the positive contribution of the diabatic terms. The agreement between the local change of
column integrated MSE and the sum of the all tendency terms (excluding diffusion) is overall
quite good. In fact, difference in the time mean of the l.h.s. and r.h.s. excluding diffusion of
equation (2) is only 1-2 W/m$^2$, as also discussed below.

The time averaged value of the MSE budget terms may also be compared to that derived
from observations. The time-mean vertical and horizontal advection of MSE derived from the
sounding array are -23 W/m$^2$ and -26 W/m$^2$, respectively, while those from WRF are 50 W/m$^2$
and -100 W/m$^2$, respectively. The sum of the advection terms, on the other hand, is ~-50 W/m$^2$
from WRF, which agrees well with that from the sounding array (-49 W/m$^2$). This agreement
appears to be due to the cancelation between the vertical and horizontal terms. In the simulation,
the time averaged surface flux is 147 W/m$^2$, and column radiative flux is -100 W/m$^2$, while the
net surface flux is 111 W/m$^2$ from the OAFlux product, and -72 W/m$^2$ from CERES.

The horizontal and vertical advection terms of the MSE budget computed with different
values of $L$ are shown in Figure 17. A large, systematic change in both terms on the order of
several hundreds of W/m$^2$, and even a change of sign in $VADV$, is found as $L$ increases from 9 to
225 km. The sum of these two (Fig. 17 c) stands in contrast to the individual advection terms in
that it has much smaller dependence on $L$. In fact, the sum of $HADV_v$ and $VADV_h$ for $L \geq 63$ km
remains nearly unchanged with respect to further increase of $L$. However, $HADV_v + VADV_h$ at $L= 9$ km exceeds that for $L = 63$ km by a ~ 40 W/m$^2$, which is significant compared to the storage
term. In other words, the MSE budget is not closed without accounting for grid-scale diffusion if
the grid variables at 9 km are used.

We attribute this disagreement to both physical and numerical diffusion processes.
Unfortunately the explicit subgridscale diffusion tendency was not saved, thus we are unable to
further quantify it. Because contribution of diffusion to moisture is small, as seen in the moisture
budget, the diffusion on MSE must come from other terms of MSE, including temperature and
geopotential height. This may seemingly contradict the fact that horizontal temperature gradient
should remain weak via gravity wave adjustment, and the diffusion on temperature is generally
small at large scales. However, weak diffusion may not be true at the scale of O (10km) on
which convective heating and non-hydrostatic pressure gradient may drive significant horizontal
temperature variations across mesoscale convective systems.

Similar to what was shown above in the water vapor budget, Figure 17 also suggests
convergence of the two advection terms as \( L \) increases. To further illustrate this, Figure 18
shows the time mean advection terms versus \( L \) as the latter parameter varies from 9 to 225 km.
Both \( HADV_e \) and \( VADV_h \) show their extremes at \( L = 9 \) km, drop off quickly with increasing \( L \) to
50 km, and level off as \( L \) increases greater than 90 km. The cancelation between the two is much
less for \( L \) greater than 90 km, conforming that focusing on the large scale horizontal and vertical
advection separately is still a valid approach at large scales.

The time mean of the residual (the sum of all the rest terms in the MSE budget excluding
diffusion) is also shown as the gray line Fig 18. The difference between \( HADV_e + VADV_h \) and
the residual for \( L \) less than effective resolution is interpreted as diffusion, as discussed above. It
is the largest (~ 40 W/m\(^2\)) at \( L = 9 \) km, decreases to ~ 20 W/m\(^2\) at \( L = 18 \) km, and becomes small
at \( L = 63 \) km. Further increase of \( L \) beyond 63 km generally leads only to slight differences
between \( HADV_e + VADV_h \) and the residual. This is the contribution of resolved mesoscale eddies,
on the order of less than 10 W/m\(^2\).

It has been argued that horizontal advection may have played a diffusive role in the MSE
budget over the large scale in other contexts, e.g., in the maintenance of the time-mean east
Pacific intertropical convergence zone (Sobel and Neelin 2006; Peters et al. 2008). This seems to
be a useful way to think about the role of the mesoscale fluxes in our simulation of the MJO as
well. Because of the apparent resolution dependence of the diffusion term in the MSE budget, it
is likely that as the model’s effective resolution increases, the diffusion would be explicitly
resolved; hence its contribution would appear in the horizontal advection term. This remains to be explored in cloud system resolving simulations of the MJOs.

The agreement of all tendency terms for \( L \geq 63 \) km suggest that mesoscale fluxes of MSE (the difference of advection terms between large \( L \) and small \( L \), if expressed in the flux form by using the continuity equation as,

\[
< \nabla v_h' q_v' > + < \frac{\partial \omega}{\partial p} > \approx 0 ,
\]

are non-divergent. The above results may seemingly invalidate the recent effort that treats separately about horizontal advection and vertical advection (rather than the total advection term) in the MSE budget. However, near invariance of both advection terms at \( L \geq 63 \) km, for both MSE budget and the moisture budget, suggests that the large scale advection is still appropriate, while one just needs to keep in mind that this is due to the fact that mesoscale fluxes of moisture and moist static energy are largely “non-divergent”, and cancel out in the vertical integral on large horizontal scales to a great degree.

4.3 Gross moist stability

The import/export of MSE by dynamic processes (advection terms) can be expressed in terms of the gross moist stability (Neelin and Held 1987, Raymond et al. 2009), which plays a key role in the dynamic instability of moisture modes. Raymond and Fuchs (2009) suggests that negative NGMS (or column-integrated moist static energy import in the presence of moisture convergence) is a general feature in the Indian Ocean. Several authors have analyzed the NGMS in climate simulations, and found that NGMS is useful diagnostic differentiating models that have better skills for MJO simulations (e.g., Benedict et al. 2014).
NGMS is normally difficult to estimate from observations, because of the difficulty of deriving large scale vertical motion with sufficient accuracy from observational datasets. Sobel et al. (2014) derived the temporal evolution of normalized gross moist stability (NGMS) within the MJO life cycles from the sounding network observational dataset (Johnson and Ciesielski, 2013). They showed that NGMS starts from small positive value, and even negative values, during the onset stage, and increases afterward to relatively large positive values.

Figure 19 shows time series of NGMS, defined here as

$$\frac{\langle \mathbf{v} \cdot \nabla e \rangle + \langle -\omega \frac{\partial \tilde{h}}{\partial p} \rangle}{\langle -\omega \frac{\partial s}{\partial p} \rangle}$$

from the WRF simulation. The denominator and numerator are parts of the moist and dry static energy budget, respectively; therefore, they may also be computed via the residual terms in the budget equation. The estimate of NGMS indirectly from the residual terms will be referred to here as the indirect estimate of NGMS, while the NGMS computed using equation (3) will be referred to as direct NGMS. It can be shown by rearranging the dry and moist static energy budgets that NGMS is directly related to precipitation in the time mean: small or even negative NGMS is associated with large precipitation (Sobel 2007, Raymond et al. 2009, Wang and Sobel 2011, 2012; Anber et al. 2013). The following procedure is used to estimate NGMS. All variables are averaged to the effective model resolution (63 km). Spatial and temporal averaging are applied to the denominator and numerator before we evaluate NGMS. Specifically they are averaged over the NSA region and further smoothed by a 5 day running average. The estimate of NGMS is meaningless if the denominator gets close to zero and/or changes sign; for this reason,
we do not show NMGS when it is smaller than the average value of the denominator, as Sobel et al. (2014). The NMGS during the suppressed phase is omitted as a consequence.

As in the observation analysis in Sobel et al. (2014), NGMS in the NSA region from the WRF simulation is negative (-0.4) before the precipitation peak, and increases gradually to 0.8 at later stage (Figure 19a). The temporal evolution of NGMS reflects the transition of the large scale vertical motion from shallow, bottom heavy, to deep convection regimes, and top heavy omega profiles, as also shown in Figures 4 and 5 for both MJO events. The agreement between direct and indirect estimated NGMS lend us some confidence in our estimate of NGMS.

Figure 19 b and c show the longitudinal variation of the time series of both directly and indirectly estimated NGMS. The denominators and numerators are estimated at the spatial resolution of 5 degrees within the latitude belt from the equator to 5N. Values of NGMS less than -0.6 are seen at nearly all longitudes, followed by gradual increases of NGMS to positive values, except around Oct 10 at 60 to 70 E. Nevertheless, the local maximum of NGMS in this area does not persist, and the NGMS drops to small values Oct 11 before increasing gradually again, as in other longitudes. The two MJO events also differ quantitatively: the November one has larger maximum NGMS, and propagates faster than the October one.

In sum, the WRF simulation adds further evidence to the observationally derived temporal and spatial evolution of NGMS, as shown in Sobel et al. (2014). This temporal variation of the NGMS as an integrated part of the MJO life cycle has yet to be incorporated in any theoretical models, to our knowledge.

5. Summary and conclusions
Several MJO events in the equatorial Indian Ocean were observed during the international CINDY/DYNAMO field campaign from October 2011 to March 2012. This study focuses on the October and November MJO events as simulated by a regional model at horizontal resolution of 9 km. Results from the simulation are compared quantitatively to multiple observational datasets. The main results are summarized as follows:

(1) The simulation captures both the October and November MJO events. Slow eastward progression of the MJO active phase at a speed of 5-7 m/s is well simulated. Large-scale vertical motion in the simulation has weak ascent in the leading edge of the MJO envelope, followed by deep ascent with first baroclinic mode structure during the peak precipitation stage, and trailed by second baroclinic mode structure with ascent in the upper troposphere and descent in the lower troposphere. The trailing ascent/descent region is associated with significant net column radiation anomalies, and broadens significantly due to its westward expansion during the eastward propagation of the MJO events.

(2) Along with the eastward propagation, both the simulated and observed MJO events also have a slow northward propagation component (~ 1m/s). Two tropical cyclone events closely related to the MJO events are observed. The WRF model simulates similar TC structures to those in observations, but with weaker amplitudes. Significant high-frequency westward-propagating convective activity is found within the MJO envelope in the 3-hourly rainfall field. A wavenumber-frequency diagram of satellite-retrieved rainfall indicates that these are westward inertia gravity waves with phase speeds of 9-22 m/s. The model captures similar wave signals in the frequency-wavenumber domain.

(3) Comparison with the observations from the northern sounding array (Johnson and Ciesielski 2013) indicates that the model simulates the passage of the two MJO events over the northern
sounding array region well: rainfall peaks over 30 mm/day during late October and November; OLR drops by more than 100 W/m$^2$, and reflected short wave increases by more than 150 W/m$^2$. Simulated large-scale vertical velocity over the NSA follows the shallow convection, deep ascent, and ascent/descent sequence that is also found in the vertical motion deduced from observations. The time mean vertical velocity is $1.1 \times 10^{-2}$ m/s, close in amplitude to the NSA derived value ($\sim 1.2 \times 10^{-2}$ m/s). Lower tropospheric westerlies gradually deepen during the October MJO event while the upper tropospheric easterlies strengthen. Temperature anomalies develop first at low level and later in the upper troposphere, where they attain maxima on the order of 1-2 K. This time sequence indicates a westward tilt structure in the longitude-height plane, which appears to be a general MJO feature.

(4) In numerical experiments with daily SST replaced by time independent SST, the October MJO event is well reproduced in both 850 hPa zonal winds and surface precipitation, but the simulation of the November one is much degraded. This suggests that daily SST plays an important role for the November MJO event, but much less so for the October event.

(5) Analysis of the moist static energy budget shows that both advection and diabatic processes contribute to the MSE buildup, but their contributions differ in the lead time prior to the precipitation peak in a manner broadly consistent with earlier studies. It is shown that the contribution from mesoscale fluxes/advection terms to the moisture and MSE budget are small, although the individual terms can be significant. As more of the mesoscale contribution is included by decreasing the horizontal averaging scale on which budgets are computed, vertical advection is more positive, and horizontal advection is more negative. The MSE budget is well balanced if all variables are averaged at the effective model
resolution (~7 times of the grid spacing). However, a deficit of 40 W/m² is found if all
variables at model grid spacing are used for the MSE budget. This MSE deficit is interpreted
as diffusion, which may be explicitly resolved if model grid spacing is further increased.

(6) While the free-running WRF simulation captures many aspects of the MJO events
qualitatively, quantitative model biases are also apparent, notably in the following: (a) the
imulated November MJO event is delayed by 3 days, and also weaker than observations; (b)
the model shows high surface flux during the first MJO events, while it shows no increase in
the OAFlux product; (c) while the MJO precipitation is greater in the model than in
observations, OLR and column net radiation are weaker, suggesting that model has a weaker
radiative feedback; (d) The ITCZ in the southern hemisphere (south of 5S) is not simulated in
the model; (e) vertical advection of moist static energy is positive in the time mean (~ 50
W/m²), in contrast to negative values (~ -22 W/m²) derived from the sounding array
observation. Horizontal advection of MSE is ~ -100 W/m², which is also less than
observational value (~-26 W/m²).

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Figure 1. (a) 850 and (b) 150 hPa zonal winds averaged from October to December. The WRF domain is indicated by block box.
Figure 2. Daily surface precipitation (mm/day) from (a) WRF and (b) TRMM averaged over the latitudes 0 - 5°N. Bottom panels: as top panels but for 3-hourly precipitation.
Figure 3. Time-longitude diagram of zonal wind at 850 hPa (shading) and daily surface precipitation (contour, 15 mm/day) averaged over the latitudes 0 - 5° N for (a) WRF and (b) ERA-Interim and TRMM. A 3-day moving average is applied to daily precipitation.
Figure 4 Longitude-pressure diagram of 3-day mean pressure velocity (hPa/hour, red shading: descent; blue shading: ascent) from Oct 12 to Nov 1, for the October MJO event. Surface rainfall (mm/day, gray curve with the vertical axis on the right), and column radiation (W/m², black curve with the vertical axis on the rightmost) is also shown. Omega, precipitation, column
radiation is first averaged over the latitudes 0 – 5°N; a 250 km running average along longitude is further applied.

**Figure 5** As in Figure 4 but from Nov 11 to Dec 2 for the November MJO event.
Figure 6. Spectrum of surface rainfall in the Indian Ocean (50°E to 90°E) averaged between 5°S to 5°N for (a) CMORPH (b) WRF. The temporal window is 16 days, and the temporal resolution is every 3 hours. The solid curve corresponds to n=1 inertial gravity waves and Kelvin waves with equivalent depth of 12, 25, and 50 m.
Figure 7. Latitude-time diagram of relative vorticity at 850 hPa (shading, $10^{-5}$ s$^{-1}$) and daily surface rainfall (gray contour: 12 mm/day), both averaged in the longitude bands 55-90° E.
Figure 8. Left panels: total precipitable water from TIM/SSMI (shaded), TRMM daily precipitation (white, 10, 30 and 50 mm/d), and ERA-I horizontal winds at 850 hPa (vectors). Right panels: the same as the left, but for the WRF simulation. Three time snapshots are shown: 00UTC Oct 27 (top), 00UTC Nov 3 (middle), and 00UTC Nov 27 (bottom).
Figure 9. Time series of (a) precipitation (mm/day) from WRF, TRMM, and budget derived rainfall, (b) OLR (W/m²) from WRF and CERES, (c) Reflective short wave, and (d) net Top-of-Atmosphere radiation averaged over the NSA region. Quantities from the WRF model is averaged over the region 73°E-80°E, Equator-5°N.
Figure 10. (a) Vertical motion (cm/s) from WRF averaged over the region 73°E - 80°E, Equator – 5°N. and (b) the DYNAMO northern sounding array. Time mean and standard deviation of W are shown in (c) and (d), respectively.
Figure 11. (a) Zonal winds (U, m/s) from WRF averaged over the region 73 °E - 80 °E, Equator-5 °N, and (b) the DYNAMO northern sounding array. Time mean and standard deviation of U are shown in (c) and (d), respectively.
Figure 12. (a) Temperature anomaly (K) from (a) WRF averaged over the region 73°E-80°E, Equator – 5°N, and (b) the DYNAMO northern sounding array. A 3-day moving average is applied. (c) Time mean temperature profile. (d) Standard deviation of temperature anomaly.
Figure 13. (a) Relative humidity from (a) WRF averaged over the region 73°E – 80°E, Equator–5°N, and (b) the DYNAMO northern sounding array. Time mean and standard deviation of relative humidity are shown in (c) and (d), respectively.
**Figure 14.** Precipitation (15 mm/d) and 850 zonal wind averaged between the equation and 5°N from (a) daily SST (identical to Fig. 3a); (b) time independent SST averaged from October to December; (c) as (a) but initialized at Nov 13; (d) as (c) but with time independent SST averaged from Nov 13 to Dec 13.
Figure 15. Column integrated water vapor budget over the NSA regions (0 – 5°N, 73°E-80°E) (a) Vertical advection, horizontal advection, rain, residual, and the tendency term, calculated at $L = 9$ km (the grid spacing). (b) as (a), but $L = 63$ km. (c) Column integral of vertical advection (mm/day) on different horizontal scales: 9, 18, 63, 135, 225 km. (d) As (c), but for horizontal advection (mm/day). (e) as (c), but for the sum of horizontal and vertical advection on different scales. These daily time series are also smoothed by 3-day moving average.
Figure 16. Budget of moist static energy over the NSA region. (a) Radiation and surface fluxes; (b) Horizontal and vertical advection; (c) Sum of these terms (black solid) and the directly estimated tendency d<MSE>/dt (dashed) at L = 63 km. Simulated rain (gray) is also shown to indicate MJO events.
Figure 17. Column-integrated nonlinear advection terms of the MSE budget on different scales with $L = 9, 18, 63, 135,$ and 225 km: (a) vertical advection, (b) horizontal advection, and (c) the sum of the two terms.
Figure 18. Time mean of the nonlinear advection terms averaged over NSA versus $L$ varying from 9 to 225 km for vertical advection term (red), horizontal advection (blue), and the sum of the two terms (black). Gray line indicates the sum of the rest terms in the MSE budget, but excluding diffusion.
Figure 19. (a) Time series of normalized gross moist stability; (b) time-longitude diagram of directly estimated NGMS; (c) as (b), but for indirectly estimated NGMS computed using the MSE budget. NGMS is not shown when the denominator is less than 1/5 of its time mean value.