Cloud-permitting simulations of the October and November MJO events during the DYNAMO field campaign. Part 1. Comparison with observations.

Shuguang Wang, Adam H. Sobel, Fuqing Zhang, Yongqiang Sun, Lei Zhou
Abstract
1. Introduction

The Madden Julian Oscillation is a dominant intra-seasonal signal over the entire tropical regions. Because of its influence on global weather and climate, understanding, simulation and prediction of MJOs has tremendous scientific and societal impact. Nevertheless, our capability to predict the MJO initialization in the Indian Ocean remains a long standing issue despite decades of research efforts. The field campaign DYNAMO (Dynamics of MJO) were designed specifically to address this issue. The DYNAMO observational network captured several MJO events from October 2011 to March 2012 (Zhang et al. 2013, Goncharck et al. 2013). Unprecedented amount of data were collected over multiple observational platform in the region of the central Indian Ocean. In particular, the northern sounding network captured time evolution of two MJO events during the first two months.

This study investigates the two MJO events using a cloud-permitting regional modeling system. Recent development of the MJO theories, despite of their detailed treatment of dynamics, have converged to the point that they all agree upon on the prominent roles of free-tropospheric moisture, including its spatial and temporal variations over the MJO life cycle, horizontal/vertical transport by shallow/deep convection, or large scale flow, and phase change and associated diabatic heating. The MJO skeleton model () Sobel and Maloney 2013

Because of limited observations are available for moisture distribution, most of these efforts are based on reanalysis dataset, which is known to have not so small residuals with the moisture budget (e.g., Mapes et al. 2012, Kim et al. 2013). Due to its prominent roles in the moist static energy, analysis of moist static energy often suffers similar deficiency. The moisture budget derived from sounding network also have its own
Large scale upper tropospheric diabatic heating is a critical component of the MJO. However, precipitation associated with this MJO diabatic heating and their hydrological cycle has been known to be very difficult for numerical weather prediction models and climate models. Even the reanalysis dataset has deficiencies in reproducing rainfall pattern during various MJO phases (e.g., Stephens et al. 2010, Kim et al. 2013, Sobel et al.) – a common problem in various reanalysis dataset is that too much rainfall is produced during the suppressed phase and rain intensity is underestimated during the active phase.

Continuously development of numerical models has led to progressive improvement of the MJO simulations in climate models (e.g., Hung et al., 2013) and global weather prediction models, On the other hand, regional models that explicitly simulated convective elements have not been explored. MJO simulations and prediction from regional models are still mixed.

Several prior studies (Holloway et al. 2013, Khouider and Han, 2013, Hogos et al. 2012) have reported that high resolution models may provide reasonably well simulations of MJO signals at lead time of one to two weeks.

The rationale of regional simulations the MJO initializations is that such modeling system is capable to (1) use fine resolution to better resolve high frequency mesoscale viabilities and their upscale impact; (2) minimize error outside the region of interests, thus identify the key processes within the regions of interests. The latter actually permits Ray et al. (2010) to simulate MJO wind signals with a lead time more than one month, far beyond the prediction limit (~10-15 days) from the GCMs (Waliser et al., 2003).
Ray and Zhang (2010) found that MM5 model simulations of two MJO events produced realistic MJO wind fields but too little surface rainfall (personal communication). Ray et al. (2010) suggested that the mean state biases in the tropical channel WRF model may fail simulations of the MJO events. Case studies by Hagos and Leung (2011) and Hagos et al. (2011) showed that convection-permitting resolution (4 km) was able to capture realistically two MJO events, whereas free-running simulations at coarser resolution (36 km) failed. These mixed results seem to suggest that while high-resolution regional models are promising for tropical weather forecast beyond 2 or 3 weeks, their current skills are not fully realized.

While metrics for the MJO at global scales have been quite well established using the RMM index (Mathew and Henden 2004), measure of the MJO at regional scales particular at IO remain to be controversial issue. Straub discussed that global RMM index may not always be an accurate index for the MJO initialization event at IO. While surface rainfall remains to be Ling et al. discussed another measure of the MJO initialization using the surface rainfall. Precipitation is the most important quantity for MJOs. Both large scale models (e.g., Kim et al. 2013) and high resolution regional models (e.g., Zhang, personal communication, who was able to simulate wind signals but not precipitation) both has considerable difficulty in precipitation.

This purpose of the present study is to use a regional modeling system to (1) simulate the October and November DYNAMO MJO events, (2) validate the simulation of the mean state and intraseasonal variability using DYNAMO observation dataset, (3) demonstrate the potential of using a regional system in the intraseasonal prediction of weather and climate for MJO and tropical cyclones.

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 ECMWF-Interim reanalysis (Dee et al. 2011) is adopted for produce the initial, bottom and lateral boundary conditions for the regional modeling. The lateral boundary consists of a narrow 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.

The numerical domain include lateral boundary from 20 S to 20 N, 48 E to 120 E, as shown in Figure 1. A total of 45 vertical levels are used, with 9 levels in the 1 km and a nominal top at 26 km. Horizontal grid spacing is 9 km. This is not adequate for individual convective cells, but can partly resolve organized cloud systems and many other mesoscale circulations, as well as their coupling with large scale dynamics, hence their upscale impact.

At the lower boundary, SST is updated every 6 hours for lower boundary conditions over sea. Surface temperature over land is parameterized through the Unified Noah land-surface physics scheme. This creates diurnally varying surface temperature over Maritime continent, and important for simulated synoptic and mesoscale viabilities over the Indian Ocean. The CAM radiation scheme (Collins et al. 2004) is used to parameterize both long wave and short wave radiative heating.

The following physical parameterization schemes are used: vertical turbulent eddy mixing in the boundary and the free troposphere is parameterized using the YSU PBL schemes (Hong et al., 2006), the Monin-Obukov scheme for surface fluxes, the surface skin temperature using (Zeng et al. 2005), the WDM double momentum microphysics from WRF 3.5.1. 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 2\textsuperscript{nd} order horizontal turbulent eddy mixing is parameterized using the Smagorinsky first order closure, and evaluated in the physical space. The combination of these physical schemes has been tested in idealized double-periodic radiative-convective simulations to ensure that both column-integrated budget of moisture and moist static energy are well closed. The numerical check also ensures that drift of model simulated climatology is minimal and does not degrade simulated temporal variability.

The model simulations start from 1 Oct, 2011. In the first 3 days, horizontal winds are relaxed to ERA-Interim using the spectral nudging technique with a zonal wavenumber 4 (\sim 2000 km) and a meridional wavenumber 2 (\sim 2000 km). No nudging is used for other variables. This nudging does not introduce additional sources/sinks for moisture, but tightly constrain mean flow and horizontal convergence over large scale. This 3-day spectral nudging also allows the mesoscale to saturate on the spectral space (e.g. Zhang et al. 2007).

Starting from Oct 4, model is integrated without any nudging, while the tendencies of all the variables within the five points near the lateral boundaries are gradually replaced by the ERA-I tendencies, and SST is updated at the ocean surface.

\textbf{2.2 Observational dataset}

Several DYNAMO dataset is used to validate our regional simulation of the MJO events. We use DYNAMO northern sounding array (Johnson and Ciesielski, 2013, Ciesielski et al., 2013). ECMWF-Interim dataset (Dee et al. 2011), CERES radiative fluxes, …JTWC hurricane best rack dataset.
3. Results

3.1 Rain and Radiative fluxes

The mean zonal wind in the equatorial northern Indian Ocean maintains westerly at the lower troposphere, and strong easterly at the upper troposphere over the entire period, as shown in Figure 1. This mean westerly provides a favorable condition for MJO events. Several GCM modeling studies suggested that the lower level westerly is very important for the MJO (e.g., ). Indeed, the coherent MJO structure observed during the DYNAMO were strongest over the NSA, but less so in the SSA.

Figure 2 a and b shows the Hovmoller diagram of daily rainfall averaged between 5S and 5N from TRMM (3B42V7) and the WRF simulation. Both quantities shows clearly two MJO events, starting from ~ 60 E, propagating slowly eastward in the Indian Ocean, with ~ 5m/s phase speed for the October MJO event, and slightly faster speed for the November event. The eastward propagation of rainfall is greatly disrupted near the maritime continent (~ 100 E) in both the model and observation: the October event essentially shows no propagation in rainfall in the WRF model, while the November event is more coherent in both model and Obs after its passage of the Sumatra island. The MJO propagation over the maritime continent is generally very poorly simulated in most climate models (e.g., Huang et al. 2013).

Despite its long period (10 days apart) from the model initialization date, the simulation correctly captured the initialization of the first events ~ Oct 16 at 60 E, and ~3-5 days earlier of the November MJO event. The model also simulated correctly the relative phase of both events - similar to observation, the model simulates relatively slower eastward propagation of the October event and its slower propagation to the east.
The slow eastward progression of the MJO is also shown in the low level zonal winds. Figure 3 shows that Figure 3 shows the Hovmoller diagram of the large scale pattern of 850 and 200 hPa zonal winds, and 500 hPa pressure velocity.

**Figure 4 and 5** show the spatial structure of 3-day averaged large scale omega averaged between the equator and 5°N for the two MJO events, along with precipitation (gray curves). During Oct 12-14, precipitation maximizes near 66°E, and a relatively weak but broad lower tropospheric ascent is located between 65 to 85°E. During Oct 15-17 when the MJO event is already initialized, the large scale structure features westward tilt of the ascent region and a top-heavy omega with distinct wavenumber 2 structure (i.e., descent in lower troposphere and ascent in the upper troposphere) between 60°E to 65°E, while the leading edge of ascent (70°E - 77°E) is dominantly bottom heavy. During 21 – 23 Oct, a top-heavy omega with wavenumber 1 structure (ascent over the whole troposphere with a strong peak in the upper troposphere) collocates with the precipitation maximum between 67°E and 73°E; eastward of this region is mostly bottom-heavy ascent, while the westward of this region is wavenumber 2 top-heavy omega. Overall, the bottom heavy ascent eastward of the precipitation center, deep wavenumber-1top-heavy ascent in the region of maximum precipitation, and wavenumber-2 top-heavy ascent trailing the maximum precipitation is maintained throughout the eastward progression of the October MJO event. The region of wavenumber-2 top-heavy ascent is particularly broad, spanning a longitude of 10 – 15 degrees.

This structure of vertical motion is a reminiscent of the cross-scale self-similarity in the cloud fields - as summarized in Mapes et al. (2006), mesoscale convective systems, 2-day waves, synoptic convectively-coupled Kelvin waves, and MJOs, all have shallow structure at the leading edge, followed by deep clouds, and trailing stratiform processes. Here, we show a smilnar
structure in omega. Dynamical implication of this structure will be further explored in the later section on the analysis of gross moist stability.

The November MJO event shows similar longitude structure of vertical motion, but at a slightly faster pace. The WRF model simulation also shows that precipitation and omega associated with MJO do not propagate continuously, but often assumes a stepwise eastward progression, e.g., from 21-23 Nov to 27-29 Nov. This is probably related to synoptic variability (e.g., Rossby waves).

High frequency convective events are revealed using 3 hourly surface precipitation data. Figure 2 c and d show clearly westward fast propagating signals in Obs and WRF, but missing in the daily quantities. Some of these westward waves are prominent during the active phase of the two MJO events in the open Indian Ocean; within the MJO envelope, while others are geographically linked to large islands over the maritime continent, e.g., the Samatra Island. These fast signals are in phase with diurnal variations, and closely related to the so-called two-day waves, which has been observed by the precipitation Radar deployed at the Gan radar supersite (e.g., Zuluga and Houze 2013).

The high frequency variability in the time-longitude diagram of 3-hourly precipitation also bears out in the spectral space. Figure 6 shows the wavenumber-frequency diagram, a regional equivalent of the global Wheeler and Kiladas diagram, for the 3-hourly surface rainfall from TRMM and WRF. A 30-day time segment with a 10 day overlap is used for Fourier transformation at each latitude spanning the longitude range 50E to 95E, and the average over the latitude bands from the equator to 5 N is further applied. The background diagram is obtained by applying multiple pass of a 1-2-1 filter on the wavenumber space – 40 passes in the present study. Because of the relative short longitude and temporal window, low frequency and low
wavenumber component are not resolved. Despite this caveat, primary features standing in contrast to the background red spectrum are the n=1 WIGs (westward inertia-gravity waves), bounded by dispersion curves for the theoretical inertial-gravity waves with equivalent depths from 12, 25 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. The WRF simulation also has substantial power at sub-diurnal time scales below 1000 km.

3.2 Comparison with northern sounding array

The northern sounding network deployed in the central Indian Ocean was designed to observed time evolution of the MJO. Johnson and Ciesielski (2013) shows that the two MJO events are better captured in the northern sounding array. In the following, we evaluate the model simulation using the sounding array averaged derived from sounding observations in the northern sounding array network (NSA) by Johnson and Ciesielski (2013).

Time series of OLR averaged over the region is shown in Figure 7. CERES 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 ~150 W/m2 during late November. Reflect TOA short wave shows a significant increase from 50 W/m2 at the suppressed phase to more than 200 W/m2 at its active phase. Such dramatic change in OLR is a marked feature of the MJO events in the Indian Ocean, significantly exceeding OLR anomalies observed in the western Pacific during the TOGA-COARE (Lin and Mapes 2004). The radiation anomalies may lead to radiative-convective instability, as several authors have postulated the MJO initialization in the Indian Ocean (Lee et al. 2001, Raymond 2001, Sobel et al. 2008). The WRF simulated OLR anomalies during late October and November is about 10 W/m2, and the minimum values are 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 3c) is 10 W/m² less than CERES. At TOA the long wave and short wave compensate each other to some degree, but remain negative of the atmosphere-ocean column in both CERES and the WRF simulation.

Vertical motion is directly related to deep convection and surface precipitation. Figure 8a shows the WRF simulated large scale W (cm/s) averaged over NSA. Large scale omega derived from the NSA (Figure 8b) shows signatures of multiple precipitation episodes in the October MJO events, and two strong ascents in the November events. The WRF model shows similar characteristics of large-scale ascent for both events. Both the model and NSA sounding data shows that, vertical motion has a first baroclinic mode structure during the MJO active phase, and develop a second baroclinic mode structure at late stage of the 2 MJO events, although this second baroclinic mode structure is less clear in Obs. This time evolution resembles the time-longitude diagram in Figure 3 and 4.

Model bias is evident in several aspects: as precipitation, WRF simulated W is larger for the October event, but weaker for the November event. Time-averaged vertical profile of W from NSA is top-heavy, similar to W from the TOGA-COARE in the western Pacific, with a peak value of ~1.2 cm/s. WRF simulated mean values of W agrees well with the observation, (Figure 8c), but its upper-tropospheric peak is located slighted lower altitude (~400 hPa), and its amplitude is slightly weaker (~1.3 cm/s). The standard deviation (Figure 8d) also shows similar vertical structure with a local maximum in the upper troposphere.

Figure 9 compares time series of model simulated the zonal wind (Fig. 9a) with the NSA observation (Fig. 9b). In both NSA and model, during the first half of October, lower troposphere over the NSA region is dominantly westerly, become easterly prior to the October
MJO event, and turns into westerly during the MJO active phase (later October and early November). The simulated gradual transition from easterly to westerly may be regarded as a bottom-top process, starting from near surface ~ Oct 20\textsuperscript{th} and reaching to 400 hPa in the first week of November, while NSA shows a similar but less distinct gradual deepening of westerly during over a ~5-day period near Nov 1\textsuperscript{st}. After the passage of the October MJO event, easterly prevails in both model and Obs until ~ Nov 20\textsuperscript{th}, when the November MJO event arrives at the NSA. The low-tropospheric westerly wind burst observed during the Toga-Coare MJO events (Lin and Johnson 1996) is very strong during the last 5 days of November, nearly 10 m/s in Obs, 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, and a weak westerly in the lower troposphere, as seen in both model and obs (Fig. 9c). The standard variation (Fig. 9d) has a peak in ~ 150 hPa in both model and obs, but the model is stronger from 900 hPa to 400 hPa.

Figure 10 shows relative humidity with respect to ice, computed from the mean profiles of temperature and water vapor over the NSA region. The prominent dry condition over the troposphere during suppressed MJO phase is evident in both WRF simulation and NSA observations. Gradual lower tropospheric moistening prior to the MJO events (Oct 10\textsuperscript{th} – 20\textsuperscript{th}, Nov 15\textsuperscript{th} – 20\textsuperscript{th}) is similar in both. Drying post MJO active phases is less dramatic in the WRF simulation. Time mean relative humidity also shows a local maximum near the tropopause, between 100 and 200 hPa (Fig. 10 c), ~70% in NSA and 80% in WRF. This local peak also persists even during persisted during the suppressed MJO phase. T may be related to the anomalous cirrus observed from the satellite (e.g., Vitat et al. 2010, Del Genio et al. 2012).
3.3 Column integrated budget of water and moist static energy

3.3.1 Budget of water vapor

Consider the budget of column integrated water vapor:

\[
< \frac{\partial q_v}{\partial t} > + HADV + VADV = E - P
\]

\[
VADV = < \vec{v}_h \nabla q_v >, \text{ and } VADV = < \omega \frac{\partial q_v}{\partial p} >,
\]

where <> denotes mass weighted vertical integral, HADV and VADV are horizontal and vertical advection at resolved scales by the model. \( E \) is surface evaporation by the subgrid scale eddies; \( P \), precipitation. The nonlinear advection terms are evaluated using the 9 km model data. Figure 1 shows water in three area:

Compared to the NSA area, a distinct large term is

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

\[
HADV = < \vec{v}_h \nabla \vec{h} > + < \vec{v}_h \nabla \vec{h} ' >, \text{ and } VADV = < \omega \frac{\partial \vec{h}}{\partial p} > + < \omega \frac{\partial \vec{h} '}{\partial p} >.
\]

For a quantity \( A \) written as \( A = \vec{A} + A' \), overbars denote spatial and/or temporal average, and primes’ denote subgrid eddies. Using the reanalysis data or observational sounding network, the budget is typically performed over a finite area covering a few hundreds of kilometers, similar to or smaller than the DYNAMO sounding network. These two terms are commonly evaluated without contribution from subgrid eddies. Because of inherent ambiguity of “average”, we use a range of horizontal scale for averaging: 9 km (the model horizontal grid spacing), 45 km, 90 km, 225 km, and 450 km (slightly smaller than the NSA).
The total of advection, HADV + VADV, is scale independent, meaning that they are

Note that horizontal average has been dropped out in these expressions, but one needs to keep in mind that nonlinear terms are evaluated as \( \overline{V_h \nabla h} \) and \( \omega \frac{\partial h}{\partial p} \), including contribution from both mean and eddies.

### 3.3.2 Budget of moist static energy

The seminal work of Yanai et al. 1973 formulated a framework to evaluate contribution of convective ensemble to large scales from an observational sounding network. The MSE Budget of MJOs has been explored by several authors in different context (Maloney 2009; Kiranmayi and Maloney, 2011; Wu and Deng 2013, Anderson and Kuang 2012). Sobel et al. 2014 recently studied the MSE budget of the DYNAMO MJO events using both observation and the ERA-I reanalysis data, and showed that both the horizontal and vertical advection term had a negative contribution to the MSE budget during the active phase of the DYNAMO events. They also estimated a very small contribution of the energy conversion from kinetic energy to static energy - a term that was not considered in Yanai et al 1973, yet its magnitude has been properly evaluated (Neelin 2007).

The following section will show the MSE budget over the same area of NSA, but for the WRF simulation, which also allows us to explicitly compute horizontal advection by subgrid eddies.
Column-integrated budget for moisture, dry static energy, and moist energy, may be written as:

\[
< \frac{\partial q_v}{\partial t} > + HADV_q + VADV_q = E - P
\]

\[
VADV = < \mathbf{v}_h \nabla q_v >, \text{ and } VADV = < \omega \frac{\partial q_v}{\partial p} >
\]

\[
< \frac{\partial q_v}{\partial t} > + < \mathbf{v}_h \nabla q_v > + < \omega \frac{\partial q_v}{\partial p} > = E - P
\]

\[
< \frac{\partial s}{\partial t} > + HADV_s + VADV_h = R + SH + LH + D_h
\]

\[
< \frac{\partial s}{\partial t} > + HADV + < \omega \frac{\partial h}{\partial p} > = R + SH + LH + D_h
\]

where \( q_v \), \( s \), and \( h \) represents water vapor, dry enthalpy \( (s=c_pT) \), and frozen moisture enthalpy \( (h = c_pT+L_vq_v-L_iq_i) \), respectively. \( \omega \) is pressure velocity; \( E \), surface evaporation; \( P \), precipitation; \( R \) column net radiation; \( SH \), surface sensible flux; \( LH \), surface latent flux, \( D_h \) subgrid diffusion.

The effective resolution of the WRF model is about 5-7 times of grid spacing.

Equation (1) restates the Weak Temperature Gradient approximation by neglecting the local time tendency and horizontal advection of temperature. In the moist budget equation (2), the terms from the right to the left represent the local tendency, horizontal advection, vertical advection, surface evaporation, and precipitation, respectively. This equation bypasses detailed microphysical phase transformations, but provides a rigorous check on the budget of moisture. Equation (3) is the sum of equation (1) and (2); it plays a key role in several theories of MJOs.

Figure 5 shows the various budget terms of column-integrated moisture budget, including vertical transport, horizontal transport, surface evaporation and precipitation, and the storage term, all being averaged over the region of the NSA.
The dominant balance of column water budget during the active MJO phase is between vertical transport by both convective and stratiform processes and removal of water vapor by the precipitation. The residual between the two is also not small, often on the order of a few mm/day (e.g., during Oct 20-28, Nov 20-25), and mostly balanced by horizontal dry air advection. On the other hand, surface evaporation also plays active roles, in particular as the dominant water source during the suppress phase from Nov 2 - Nov 17, when model still simulates rainfall of 3-4 mm/day over the region, while TRMM or budget derived rainfall only shows rainfall of 1-2 mm/day or even less.

Surface evaporation and horizontal advection also plays a prominent role regulating large scale processes. This is evident in the MSE equation. The former, often termed wind-induced-surface-heat-instability, were proposed as the leading mechanism for MJO and large scale waves in its linear form.

Time series of moist static energy over NSA is shown in Figure X. As the observation (Sobel et al. 2014), it is small and even negative before the peak of precipitation at the large scales, and increases steadily to 0.8 at later stage. Such structure reflect a transition of the large scale vertical motion from shallow, bottom heavy, to deep convection regimes, and finally to the stratiform cloud with a peak.

3.5 Tropical cyclones

The two MJO events have a meridional propagating component. Figure 3 displays the latitude-time diagram of rainfall from the model and TRMM, both averaged between 55E-90E.
Due to the slowly northward migrating Rossby waves, northward propagation of the October MJO event appears to initiate ~ Oct 15 southern of the equation (5S – equator), and large scale precipitation migrate northernward at a speed of ~ 1-2 m/s. The November event These equatorial features (e.g., Rossby waves) transform into TC types as the system pass 5 N, the critical latitude for TC genesis.

Note that persistent rainfall south of 4S in TRMM may be due to the ITCZ structure or the effect of the ocean thermalcline ridge, where a shallow ocean mixed layer play a significant role in exciting rain. WRF produces little rain south of this 4S, partly due to the lack of the ocean coupling, which is crucial for the ocean thermalcline ridge.

4. Discussion and conclusions

The field campaign Dynamo (Dynamics of MJO) observed several MJO events during its entire period.
Acknowledgement

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http://data.eol.ucar.edu/

REFERENCE


Figure 1. The WRF domain indicated by block box in the top panel. Time mean zonal winds at October and November.
Figure 2. 3-hourly and daily surface precipitation (mm/day) from WRF and TRMM averaged over the latitudes 5°S - 5°N.
Figure 3. Time-longitude plot of zonal wind averaged between 0-5 N at 850 hPa from ERA-Interim (left panel) and WRF (right).
Figure 4 Longitude-pressure diagram of 3-day mean pressure velocity (hPa/hour, red shading: descent; blue shading: ascent) and surface rainfall (mm/day, gray curve with the vertical axis on the right) for the October MJO event. Omega and rain is first averaged over the latitudes 0 – 5N; and a 250 km running average along longitude is further applied.
Figure 5  As in Figure 4 but for the November MJO event.
Figure 6. Spectrum of rain in the Indian Ocean (50 E to 90 E) averaged between 0 to 5N for (a) CMORPH (b) WRF. The temporal window is 32 days, spatial resolution is 8 km for CMORPH, 9 km for WRF, and the temporal resolution is every 3 hours. The solid curve corresponds to $n=0$ inertial gravity waves and Kelvin waves with equivalent depth of 12, 25, and 50 m.
Figure 7. Time series of (a) precipitation (mm/day) from WRF, TRMM, and budget derived rainfall, (b) OLR (W/m$^2$) 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 8. (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 9. (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 10. (a) Temperature anomaly (K) from WRF averaged over the region 73°E-80°E, Equator-5°N. (b) the DYNAMO northern sounding array. Time mean and standard deviation of U are shown in (c) and (d), respectively.
Figure 11. As figure 8, but for relative humidity.
Figure 12. Column integrated water vapor budget over three regions: (a) the equatorial Indian Ocean 10°S-10°N, (b) NSA, 0 – 5°N, 73°E-80°E; and (c) 0 – 5°N, 60°E-70°E. Vertical advection, horizontal advection, rain, residual, and the tendency term, are shown.

Figure 13. (a) Time series of normalized gross moist stability; (b) time-longitude diagram of directly estimated NGMS; (c) indirectly estimated NGMS.

in the Top panel: diabatic sources: radiation and surface fluxes; mid panel: horizontal and vertical advection; bottom: gross moist stability.
Figure 13. Latitude-time diagram of relative vorticity at 850 hPa (shading) and daily surface rainfall (gray contour: 12 mm/day), both averaged in the longitude bands 55-90 °E.
Figure 14. Left panels: total precipitable water from SSMI (shaded), ERA-I daily precipitation (white, 10, 30 and 50 mm/d), and ERA-I horizontal winds (vectors). Right panels: the same as the left but for the WRF model. Three time snapshots are shown: 00UTC Oct 27 (top), 00UTC Nov 3 (middle), and 00UTC Nov 27 (bottom).