Hindcasting the Madden-Julian Oscillation With a New Parameterization of Surface Heat Fluxes

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Abstract The recently developed maximum entropy production (MEP) model, an alternative parameterization of surface heat fluxes, is incorporated into the Weather Research and Forecasting (WRF) model. A pair of WRF cloud-resolving experiments (5 km grids) using the bulk transfer model (WRF default) and the MEP model of surface heat fluxes are performed to hindcast the October Madden-Julian oscillation (MJO) event observed during the 2011 Dynamics of the MJO (DYNAMO) field campaign. The simulated surface latent and sensible heat fluxes in the MEP and bulk transfer model runs are in general consistent with in situ observations from two research vessels. Compared to the bulk transfer model, the convection envelope is strengthened in the MEP run and shows a more coherent propagation over the Maritime Continent. The simulated precipitable water in the MEP run is in closer agreement with the observations. Precipitation in the MEP run is enhanced during the active phase of the MJO with significantly reduced regional dry and wet biases. Large-scale ocean evaporation is stronger in the MEP run leading to stronger boundary layer moistening to the east of the convection center, which facilitates the eastward propagation of the MJO.

1. Introduction

The Madden-Julian oscillation (MJO; Madden & Julian, 1971, 1972) is a dominant large-scale phenomenon of subseasonal variability (30–90 days) that modulates deep convection and atmospheric circulation in the tropics with slow eastward propagation (~5 m s⁻¹) across the equatorial oceans. The MJO plays a crucial role in bridging weather and climate (Zhang, 2013) at diurnal to interannual scales in the atmosphere and oceans, e.g., extreme precipitation, tropical cyclone, El Niño Southern Oscillation (ENSO), and Wyrtki jets. Understanding, simulation, and prediction of MJO have attracted scientists’ interests around the world over the past several decades (Zhang, 2005).

Despite the substantial efforts devoted to investigate the MJO, we are still facing great challenges in understanding the physical mechanisms of its initiation and propagation (Zhang, 2005). The representations of MJO in global/regional weather and climate models are far from satisfactory given that they are at the heart of the subseasonal to seasonal prediction (Gottschalck et al., 2010; Hung et al., 2013; Kim et al., 2009; Lin et al., 2006; Pegion & Kirtman, 2008; Waliser et al., 2003; Zhang et al., 2013). In order to elucidate these issues, the Cooperative Indian Ocean Experiment on Intraseasonal Variability (CINDY)/Dynamics of the MJO (DYNAMO)/Atmospheric Radiation Measurement Program MJO Investigation Experiment (AMIE; hereafter DYNAMO for brevity) field campaign collected unprecedented observations of four MJO events from October 2011 to March 2012 (Gottschalck et al., 2013; Yoneyama et al., 2013; Zhang et al., 2013). Recent research suggests that the moisture dynamics plays a crucial role in the eastward propagation of the MJO (Hsu et al., 2014; Hsu & Li, 2012; Maloney, 2009; Sobel & Maloney, 2013; Yoneyama et al., 2013). The zonal asymmetry in the lower tropospheric moisture with positive anomalies appearing ahead (to the east) of the MJO convection is critical in the eastward propagation (Benedict & Randall, 2007; Hsu et al., 2014; Hsu & Li, 2012; Kemball-Cook & Wang, 2001; Myers & Waliser, 2003). However, numerous sophisticated models fail to faithfully represent these features and tend to produce weaker and faster moving MJO (Slingo et al., 1996; Sperber et al., 1997; Zhang, 2005). Unrealistic air-sea coupling in these models could be one reason for the failure (Zhang, 2005) since surface evaporation, as a source of moisture, is an important contributor to the moistening of lower troposphere (Maloney & Sobel, 2004; Powell, 2016; Sobel et al., 2014; Wang et al.,...
Sperber (2003) found that latent heat flux dominated net surface heat flux during the MJO life cycle. Sobel et al. (2008) proposed that surface heat fluxes from ocean to atmosphere play a fundamental role in driving the MJO and air-sea interaction might provide the energy source for the MJO. Other studies have shown that enhanced surface evaporation associated with positive sea surface temperature anomalies ahead of the MJO convection may improve the MJO simulation in global circulation models (Flatau et al., 1997; Kemball-Cook et al., 2002; Marshall et al., 2008; Sobel et al., 2008; Waliser et al., 1999; Zhang, 2005).

The modeling error of the bulk latent heat flux largely due to the uncertainties of surface wind speed is presumably a key factor to the biases of moist static energy and precipitation in MJO simulations (Hagos et al., 2016; Qian et al., 2016). Latent heat flux in almost all numerical models is parameterized using bulk transfer formula (e.g., Arya, 1988), i.e., the linear equations relating surface latent and sensible heat fluxes to the corresponding humidity or temperature gradient multiplied by empirical wind speed-dependent transfer coefficients. The bulk transfer models have several limitations including not conserving energy, large errors in bulk gradient variables (near-surface humidity/temperature gradient), and uncertainties in transfer coefficients in terms of surface wind speed and roughness length. A new innovative approach for modeling the surface energy budget, the maximum entropy production (MEP) model, was recently proposed to overcome some of the shortcomings of the bulk transfer models (Wang & Bras, 2009, 2010, 2011; Wang et al., 2014). The MEP model is formulated based on the contemporary nonequilibrium thermodynamics and the classical Monin-Obukhov similarity theory for boundary layer turbulence. In the MEP model, the surface radiation fluxes are partitioned into turbulent and conductive heat fluxes, automatically closing surface energy budgets at all space-time scales without using (bulk) gradient variables and wind speed and surface roughness dependent transfer coefficients. The MEP modeled surface heat fluxes have bounded and reduced modeling errors compared to the bulk fluxes. The reestimated global surface heat fluxes using MEP model agree closely with the existing estimates (Huang et al., 2016).

In this study, the MEP model, as an alternative parameterization of surface heat fluxes was incorporated into the widely used Weather Research and Forecasting (WRF) model to examine whether the new parameterization of surface heat fluxes improves the simulation of MJO. The initiation and propagation of convection during the MJO life cycle is highly related to the moistening of the boundary layer that depends strongly on surface evaporation. This study represents a first-step effort in testing the MEP model for (1) hindcasting the October MJO event observed during the DYNAMO field campaign, (2) comparing the WRF hindcast of the MJO using the bulk transfer model with that using the MEP model, and (3) validating the WRF hindcasts of MJO against available observations. Following this introduction, section 2 outlines the formulations of the MEP model. A brief description of data sets used, model configuration, and numerical experiment design is given in section 3. The results of WRF model hindcasts are discussed in section 4. Section 5 provides some concluding remarks.

2. Methods

The Earth’s surface energy balance equations depend on the transparency of the surface media to sunlight. The conservation of energy over land (Figure 1a), where surface material is nontransparent to sunlight, is expressed as

\[ R_n = R_d^S - R_d^L + R_d^L - R_u^L = H + E + Q, \]

where \( R_n \), \( R_d^S \), \( R_d^L \), \( R_u^L \) are the net radiation, incoming shortwave, reflected shortwave, downward atmospheric longwave, and surface emitted longwave radiation, respectively; \( H \), \( E \), and \( Q \) represent respectively sensible, latent, and ground heat flux. The turbulent and conductive heat fluxes away from the surface are defined as positive. Since the water, ice, and snow media are transparent to sunlight (Figure 1b), the conservation of energy is written as (e.g., Fairall et al., 1996; Saunders, 1967; Weller, 1968)

\[ R_n = R_d^S - R_d^L = R_0, \]

\[ R_n = R_d^L - R_u^L = H + E + Q, \]

where \( R_0 \) is the net shortwave radiation entering the media (i.e., water, ice, and snow layer) equal to \( R_d^S \), and \( R_n \) is the net longwave radiation. Equations (1) and (2), the surface energy balance equations over land and water-ice-snow surfaces, are identical at nighttimes.
The MEP model simultaneously solves turbulent sensible heat flux $H$, latent heat flux $E$, and conductive heat flux $Q$ over the Earth-atmosphere interface in terms of analytical functions of surface radiation fluxes, temperature and/or humidity satisfying conservation of energy at all space-time scales. The formulation of MEP model is described in Wang and Bras (2011) for the case of land surfaces, and in Wang et al. (2014) for the case of water-ice-snow surfaces. According to the MEP formalism, maximizing the entropy production function under the constraint of surface energy balance equations in equations (1) and (2), we obtain the solution of $H$, $E$, and $Q$:

$$H = R_n - R_S^d + R_L^d - R_L^u = H + E + Q$$

$$E = B(\sigma)R_n$$

$$Q = \begin{cases} R_n - E - H & \text{land} \\ R_n - E - H & \text{water, ice, and snow} \end{cases}$$

with

$$B(\sigma) = \frac{6}{\pi^2} \left( 1 + \frac{11}{36} \sigma^{-1} \right), \quad \sigma = \frac{L^2}{C_p R_n T_s^2}.$$  

Figure 1. Surface energy budget over (a) land surface and (b) ocean surface (including water, ice, and snow). $R_S^d$, $R_L^d$, $R_L^u$, and $R_U^d$ denote the incoming shortwave, reflected shortwave, downward atmospheric longwave, and surface emitted longwave radiation; $R_n$, $R_S^u$, and $R_U^d$ denote the net radiation, net shortwave radiation, and net longwave radiation; $R_0$ is the net shortwave radiation entering the media (i.e., water, ice, and snow); $H$, $E$, and $Q$ denote the sensible, latent, and ground/surface water-ice-snow heat flux.
conductivity $\lambda$ of water-ice-snow media) are $\sim 1.5 \times 10^3$, $1.9 \times 10^3$, and $0.6-1.4 \times 10^3$ J m$^{-2}$ K$^{-1}$ s$^{-1/2}$ (for snow with density varying between 100 and 500 kg m$^{-3}$), respectively. $l_0$ is the “apparent” thermal inertia of the air representing boundary layer turbulence:

$$l_0 = \rho_a C_p \sqrt{C_1 \kappa z} \left( C_2 \frac{\kappa g}{\rho_a C_p T_0} \right)^{1/2},$$

where $\rho_a$ is the air density (kg m$^{-3}$), $\kappa$ the von Kármán constant ($\sim 0.4$), $g$ the gravitational acceleration, $T_0$ a representative environment temperature ($\sim 300$ K), and $C_1$ and $C_2$ the two universal empirical constants of the surface layer characterizing the atmospheric boundary layer stability in the Monin-Obukhov similarity equations (Wang & Bras, 2009). $z$ is a local topography-dependent reference height above the surface where the Monin-Obukhov similarity equations hold. In this study, $z$ is 2.5 m for bare soil and water-ice-snow surfaces, 4.5 m for short vegetation, and 9.5 m for tall trees (Huang et al., 2016). Equation (3) has a unique solution of $H$, $E$, and $Q$ for given $R_n$, $T_s$, and $q_s$ over land surfaces and $R_n$, $R_L$, and $T_s$ over water-ice-snow surfaces. Over water, ice, and snow surfaces, the surface specific humidity $q_s$ is commonly assumed to be the saturation humidity as a function of $T_s$ according to the Clausius-Clapeyron equation.

Extensive and independent tests (e.g., Nearing et al., 2012; Shanafield et al., 2015; Yang & Wang, 2014) have indicated that the MEP model outperforms or reproduces the traditional models of surface heat fluxes, due to its unique features compared to the other flux models. Its performance roots in that this new model has firm theoretical foundation supported by the Bayesian probability theory and the physical principles. The Bayesian probability theory provides a first-principles-based inference algorithm for synthesizing the available information about the heat fluxes from surface radiation, temperature, and humidity variables. These surface variables alone would be insufficient for modeling heat fluxes using the bulk transfer method. In comparison to the traditional methods, the MEP method allows more effective use of limited surface variables most relevant to the Earth-atmospheric exchange processes for parameterizing heat fluxes. In fact, all essential flux physical processes represented in all existing flux models are parameterized into the MEP model including conservation of energy, transport of water vapor and heat through forced and free convection in the boundary layer, and phase change thermodynamics of water. The MEP model is the only physically based model of heat fluxes independent of temperature and humidity gradient. It is also the only atmospheric turbulent fluxes model without using transfer coefficients that depend on wind speed and surface roughness. Modeling of heat fluxes without using temperature/humidity gradients, wind speed, and surface roughness should not be interpreted as these variables playing no role in the turbulent transport of water vapor and heat. Using the extremum solution of the Monin-Obukhov similarity equations (MOSE; Wang & Bras, 2010), temperature/humidity gradient and wind speed are expressed as certain analytical functions of sensible heat (and momentum) flux, hence implicitly represented by the thermal inertia parameters (e.g., $l_0$ in equation (3)) in the MEP formulation. Excluding surface roughness parameter in the MEP model is also made possible by the extremum solution of the MOSE where the original differential form instead of the integrated form of the MOSE is needed. The MEP fluxes constrained by radiation fluxes have bounded modeling errors and uncertainties not exceeding those of net radiation flux. It is evident from equation (3) that the sensitivity of the MEP fluxes to other physical parameters of the model such as the thermal inertia of surface media is much reduced compared to the traditional bulk flux models (Huang et al., 2016). In summary, the major advantages of the MPE model include closure of the surface energy balance at all space-time scales, independence of temperature and moisture gradient, and parsimony of model parameters (i.e., not using wind speed and surface roughness, and free of site-specific empirical tuning coefficients).

### 3. Model Hindcasts and Data

The Advanced Research version of Weather Research and Forecasting (WRF) model (Skamarock et al., 2008), version 3.7.1 is used in this study. WRF is a nonhydrostatic and fully compressible numerical model, with an Arakawa C-grid staggering spatial discretization for variables and a terrain-following vertical coordinate. The simulation domain covers the tropical Indian Ocean and western Maritime Continent, from 20°S to 20°N and 48°E to 120°E (Figure 2), similar to that in Wang et al. (2015a). All simulations are run at 5 km horizontal grids without cumulus parameterization as cumulus parameterization could add more uncertainties to MJO...
simulations (Hagos & Leung, 2011; Pilon et al., 2016; Zhang, 2005). The simulation has 42 vertical levels with 12 levels in the lowest 1 km and a nominal top at 20 hPa. Initial, lateral, and surface boundary conditions including sea surface temperatures (SSTs) are derived from the European Centre for Medium-range Weather Forecasts (ECMWF) Re-Analysis Interim (ERA-Interim; Dee et al., 2011) with 0.75° horizontal resolution. The lateral boundary condition is updated every 6 h while SSTs are updated daily. The hindcast start from 3 October 2011 to 15 November 2011, during which one MJO episode was observed in the DYNAMO. We focus on the free run period from 6 October to the end of the hindcast with outputs every 3 h.

Physics packages used include the WRF Single-Moment microphysics 6 class scheme (WSM6; Song-You & Jeong-Ock, 2006), the global circulation model version of the Rapid Radiative Transfer Model (RRTMG) long-wave and shortwave radiation schemes (Iacono et al., 2008), the Yonsei University (YSU) planetary boundary layer (PBL) scheme, the unified Noah land surface physics scheme (Tewari et al., 2004) among others. The two experiments analyzed in the subsequent sections share the same model setup listed above. In the control experiment (hereafter CTL), the MM5 similarity scheme (Beljaars, 1995; Dyer & Hicks, 1970; Paulson, 1970; Webb, 1970; Zhang & Anthes, 1982) bulk flux formula based on the Monin-Obukhov similarity theory with Carslon-Boland viscous sublayer were used to calculate surface sensible and latent heat fluxes for both lands and oceans. Specifically, the Coupled Ocean–Atmosphere Response Experiment (COARE) version 3.0 formula (Fairall et al., 2003) were used to calculate heat fluxes over water surfaces. In the MEP experiment (hereafter MEP), the MEP model (described in section 2) is incorporated into the WRF for calculating the surface sensible and latent heat fluxes for both land and water surfaces.

The reanalysis and observational data sets used for evaluating the WRF hindcast include total precipitable water from the ERA-Interim, surface precipitation from the 3-hourly 0.25° Tropical Rainfall Measurement Mission (TRMM) 3B42 product version 7 (Huffman et al., 2007), and surface heat fluxes from the DYNAMO.
research vessels Mirai and Roger Revelle (Yoneyama et al., 2013). The surface heat fluxes observations from the Mirai were derived using the eddy covariance method, while those from the Revelle were obtained using three algorithms, i.e., the eddy covariance method, inertial dissipation method (Fairall & Larsen, 1986), and COARE version 3.5 bulk formula (Edson et al., 2013).

4. Results

The WRF simulated surface heat fluxes were compared with ship observations from the research vessels Mirai and Roger Revelle (Yoneyama et al., 2013). Figure 3 shows the daily mean surface latent and sensible heat fluxes from the ship observations and two model experiments. The magnitude of latent heat flux in the CTL run tends to be greater than that in the ship observations, while latent heat fluxes of the MEP run in general agrees more closely with the observations (Figures 3a and 3c). Although the MEP sensible heat fluxes are systematically higher than those in the CTL run and Mirai observation (Figure 3b), the magnitude of the MEP sensible heat flux is consistent with that estimated using other methods (e.g., eddy covariance, inertial dissipation, and Coupled Ocean-Atmosphere Response Experiment [COARE] bulk formula) with the same input data (Figure 3d). The variations of surface heat fluxes in the CTL run are more consistent with the ship observations than those in the MEP run. This is partly a result of the explicit and strong dependence of the bulk fluxes on surface wind speed, and the implicit use of wind speed in the MEP formulation has reduced the sensitivity of MEP heat fluxes to surface wind speed. In other words, the variations of daily bulk fluxes are highly correlated with surface wind speed, while the MEP fluxes are constrained by surface (net) radiation with weaker day-to-day fluctuations (figure not shown). Note that the instruments onboard of the research vessels for measuring turbulent heat fluxes were installed on the top of the ship foremast approximately 20 m above the average ocean surface, the observations actually represent heat fluxes at that height instead of the actual sea surface. Given these measurement uncertainties and the significant spread among fluxes estimated using different algorithms (Figure 3d), the MEP heat fluxes are considered here to be consistent with both ship observations and those from the CTL run.

Figure 3. Time series of daily mean (left) latent heat flux and (right) sensible heat flux based on the observations from the research vessels (top) Mirai and (bottom) Revelle derived from eddy covariance (EC, solid black line), inertial dissipation (ID, long dash black line), and COARE bulk formula (BK, short dash black line). WRF simulated heat fluxes from the CTL run (blue line) and MEP run (red line) are obtained from the model grid points closest to the locations of the vessels.
Figure 4 shows Hovmöller diagrams of 3-hourly precipitation averaged between 7.5°S and 7.5°N latitude from (a) TRMM observation, (b) CTL run, and (c) MEP run. The MJO event, starting from ~60°E and propagating eastward, is reasonably captured in both the CTL and MEP runs (Figures 4b and 4c). The initiation of MJO around 16 October is more evident in the CTL run than in the MEP run where scattered convections are found over the western Indian Ocean instead. The eastward propagation of convection is less coherent in the CTL run than that in the MEP run and in the observations indicated by an interruption of convection during the active phase around the Maritime Continent (~110°E) in the CTL run. The eastward propagation of the convection is faster in the MEP run than that in the observation. The MEP run also produces more convections during the entire period than the observations, especially in the suppressed phase after 7 November. The weak eastward propagating convection occurring around 10 November is only captured in the MEP run, not in the CTL run. The CTL run generally underestimates precipitation, particularly during the active phase of the MJO across the Indian Ocean, while the MEP run produces much stronger precipitation with its intensity in closer agreement with the observations.

Since the intensity of precipitation is directly related to precipitable water in the atmosphere, we examine the domain averaged daily precipitable water (Figure 5a). Water vapor in the atmosphere steadily
regions (greater than 50 mm d\(^{-1}\)) during the entire hindcast period. TRMM precipitation spreads widely over most of the Indian Ocean with strong precipitation to the south of the equator, over the eastern Indian Ocean, and over the western Maritime Continent (Figure 6a). There are significant dry biases in the strong precipitation regions (greater than \(\sim 8\) mm d\(^{-1}\)) in the CTL run, while wet biases are seen in the weak precipitation regions (less than \(\sim 8\) mm d\(^{-1}\); Figure 6c). The MEP run tends to reduce the dry and wet biases over most regions except for the significant dry biases occurring around 5\(^{\circ}\)S, 60\(^{\circ}\)E (Figure 6e). The spatial distributions of heavy precipitation frequency are similar to those of the time-mean precipitation (Figures 6b, 6d, and 6f versus Figures 6a, 6c, and 6e), suggesting that the biases in the time-mean precipitation could largely come from biases in simulating heavy precipitation. The heavy precipitation frequency in the CTL run is at least 50% lower compared to that in the TRMM data (Figure 6d versus Figure 6b). These results demonstrate that the new MEP-based parameterizations of surface latent and sensible heat fluxes improve the simulation of local precipitation through producing more heavy rain in the WRF model.

Figure 7 compares the simulated daily mean surface latent, sensible, and total heat fluxes (sum of the latent and sensible heat fluxes) between the MEP run and the CTL run. The magnitudes of latent heat fluxes in the CTL and MEP runs are comparable, while the MEP sensible heat fluxes are weaker than the bulk sensible heat flux (Figures 7a and 7b). The MEP latent heat flux is stronger (weaker) than the bulk latent heat flux when the latter is below (above) \(\sim 100\) W m\(^{-2}\) (Figures 7a and 7d). The MEP sensible heat flux is stronger (weaker) than the bulk sensible heat flux over ocean (land) surfaces (Figures 7e and 7h). The MEP latent and total heat fluxes are in general consistent with the bulk latent and total heat fluxes over land surfaces (Figures 7g and 7i) as the heat fluxes parameterized in the land surface scheme of the CTL run are constrained by the surface energy balance. The MEP total heat fluxes are in general higher than the bulk total heat fluxes, particularly over the oceans (Figures 7c and 7f), leading to larger moist static energy in the atmosphere in the MEP run. Through extra moistening and heating of the lower atmosphere, the increased moist static energy tends to destabilize the atmosphere and precondition the convections in the MEP run, resulting in stronger precipitation compared to that in the CTL run, as discussed previously.

The PBL moistening (below 700 hPa), ahead of the convective center of MJO, favors the eastward propagation of MJO (Hsu & Li, 2012). This feature was observed during the DYNAMO (Szeoeke et al., 2015) and in the Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC) radio occultation measurements (Zeng et al., 2012). The eastward propagation appears to be less coherent in the CTL run than that in the MEP run as previously discussed. Two snapshots of the convection envelope, along with the 850 hPa relative humidity and surface latent heat flux, during the eastward propagation of MJO are shown in Figure 8: 0600 UTC, 26 October and 0300 UTC, 3 November. The convection is stronger in the MEP run than in the CTL run, with a greater convection area and less scattered convection (Figures 8b and 8d versus...
Figures 8a and 8c). The moisture increase in the PBL ahead (to the east) of the convection is much more widespread in the MEP run than in the CTL run, accompanied with greater surface latent heat flux. The surface latent heat flux, as a local direct moisture source, contributes positively to the moistening of the PBL. Stronger MEP latent heat fluxes thus appear to favor the eastward propagation of the MJO by adding more water vapor to the PBL ahead of the convection.

5. Findings and Discussions

In this study, we made an initial effort to incorporate the MEP model, as an alternative parameterization of surface heat fluxes, into the WRF model and examined the capability of the modified WRF for hindcasting a MJO event observed during the DYNAMO field campaign. A pair of cloud-resolving numerical experiments were conducted with the bulk transfer model (default in the WRF) and the MEP model of surface heat fluxes. The MEP latent heat flux in general agrees closely with the ship observation, while latent heat flux is overestimated in the CTL run compared to the observation. The MEP sensible heat flux tends to be stronger than that in the CTL run but the difference in magnitude appears to be within the uncertainty bounds given by the observation.
significant spreads among flux estimates with three different algorithms. Both the CTL and MEP simulations capture the October MJO event, while the convection envelope is stronger and propagates eastward more coherently in the MEP run than that in the CTL run. Despite these, it is important to note that the MEP model still fails to capture some important characteristics of the MJO convection. For example, the eastward propagation of the convection envelope is faster in the MEP run compared to that in the observation. The enhanced total heat fluxes (sum of the latent and sensible heat fluxes) in the MEP run result in a closer

Figure 7. The scatterplots of daily mean (left) latent heat flux (W m$^{-2}$), (middle) sensible heat flux (W m$^{-2}$), and (right) total heat fluxes (sum of latent and sensible heat fluxes, W m$^{-2}$) over (top) the WRF domain, (middle) ocean surface, and (bottom) land surface, showing the comparison of the heat fluxes calculated in MEP run versus in CTL run.
agreement between the simulated and observed precipitable water, and improves the simulation of heavy precipitation during the active phase of the MJO event. The regional dry and wet biases of precipitation in the CTL run are substantially reduced as a result of enhanced (reduced) heavy (light) rain in the MEP run. Enhanced large-scale PBL moistening ahead (to the east) of the convection center in the MEP run favors the eastward propagation of the MJO. Despite smaller scale convections in the MEP run tends to slightly blur the edge between the convective and nonconvective sections of the MJO envelop, the WRF-MEP model improves several aspects of simulation including reduction in precipitation bias and more coherent propagation over the Maritime Continent. However, additional MJO ensemble hindcast simulations under various scenarios (e.g., strong and weak MJO, and propagating and nonpropagating MJO over Maritime Continent) are necessary to further verify the findings from this analysis and provide a thorough assessment of this promising new surface flux scheme.

The MEP modeled surface heat fluxes not only automatically close the surface energy budget at all space-time scales but also have reduced uncertainties without explicitly using near-surface temperature/humidity gradients, wind speed, and surface roughness as model inputs, all of which tend to have substantial estimating uncertainties in weather and climate models. The preliminary yet encouraging results of the WRF-MEP model presented here justify further tests using regional and global coupled models for the study of tropical climate. We emphasize that the MEP model as it is now is used as an alternative algorithm of surface heat fluxes instead of a replacement of existing algorithms in the WRF. The performance of the WRF-

Figure 8. Latent heat flux (shaded), OLR (blue contour, 240 W m$^{-2}$), and 850 hPa relative humidity (dots, exceeded 85%) from (left) CTL run and (right) MEP run. Two time snapshots are shown: (top) 0600 UTC, 26 October and (bottom) 0300 UTC, 3 November.
MEP model may be further improved through, for example, improving the representation of water vapor transport and diffusion processes in the lower troposphere to potentially obtain more accurate simulations of the propagation speed and convection frequency of the MJO. Another ongoing effort is to employ the MEP model in large-eddy simulations to examine its skill in representing PBL structures over oceans under various atmospheric stability conditions.

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References

Arya, S. (1988). Introduction to micrometeorology. New York, NY: Academic Press.

Beljaars, A. C. M. (1995). The parametrization of surface fluxes in large-scale models under free convection. Quarterly Journal of the Royal Meteorological Society, 121(522), 255–270. https://doi.org/10.1002/qj.49712152203

Benedict, J. J., & Randall, D. A. (2007). Observed characteristics of the MJO relative to maximum rainfall. Journal of the Atmospheric Sciences, 64(7), 2332–2354. https://doi.org/10.1175/JAS3968.1

Doe, D. P., Uppala, S. M., Simmons, A. J., Berrisford, P., Poli, P., Kobayashi, S., ... Vitart, F. (2011). The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. Quarterly Journal of the Royal Meteorological Society, 137(656), 553–597. https://doi.org/10.1002/qj.828

Dyer, A. J., & Hicks, B. B. (1970). Flux-gradient relationships in the constant flux layer. Quarterly Journal of the Royal Meteorological Society, 96(410), 715–721. https://doi.org/10.1002/qj.49709641012

Edson, J. B., Jamparna, V., Weller, R. A., Bigmore, S. P., Plueddemann, A. J., Fairall, C. W., ... Hersbach, H. (2013). On the exchange of momentum over the open ocean. Journal of Physical Oceanography, 43(8), 1589–1610. https://doi.org/10.1175/JPO-D-12-0173.1

Fairall, C. W., Bradley, E. F., Godfrey, J. S., Wick, G. A., Edson, J. B., & Young, G. S. (1996). Cool-skin and warm-layer effects on sea surface temperature. Journal of Geophysical Research, 101(C15), 1295–1308. https://doi.org/10.1029/95JC03190

Fairall, C. W., Bradley, E. F., Hare, J. E., Grachev, A. A., & Edson, J. B. (2003). Bulk parameterization of air–sea fluxes: Updates and verification for the COARE algorithm. Journal of Climate, 16(4), 571–591. https://doi.org/10.1175/1520-0442(2003)016<0571:BPOASF>2.0.CO;2

Fairall, C. W., & Larsen, S. E. (1986). Inertial-dissipation methods and turbulent fluxes at the air-sea interface. Boundary Layer Meteorology, 34(3), 287–301. https://doi.org/10.1007/BF00212383

Flatau, M., Flatau, P. J., Phoebus, P., & Niler, P. P. (1997). The feedback between equatorial convection and local radiative and evaporative processes: The implications for intraseasonal oscillations. The Atmospheric Sciences, 54(19), 2373–2386. https://doi.org/10.1175/1520-0469(1997)054<2373:FTBEFP>2.0.CO;2

Gottschalck, J., Roundy, P. E., Schrecker, C. J. I. I., Vintzileos, A., & Zhang, C. (2013). Large scale atmospheric and oceanic conditions during the 2011–12 DYNAMO field campaign. Monthly Weather Review, 141(1), 4173–4196. https://doi.org/10.1175/MWR-D-13-00021.1

Hagos, S., Kinnison, D. E., Solomon, S., Mauason, G., & Anderson, J. G. (2001). Potential for significant interannual variability in tropical convection. Journal of the Atmospheric Sciences, 58(13), 2373–2391. https://doi.org/10.1175/1520-0469(2001)058<2373:PSFIIV>2.0.CO;2

Hagos, S. M., Feng, Z., Burleyson, C. D., Zhao, C., Martini, M. N., & Berg, L. K. (2016). Moist process biases in simulations of the Madden–Julian oscillation. Scientific Data, 3, 160031. https://doi.org/10.1038/sdata.2016.31

Hagos, S. M., & Leung, L. R. (2011). Moist Thermodynamics of the Madden–Julian oscillation in a cloud-resolving simulation. Journal of the Atmospheric Sciences, 68(12), 3551–3563. https://doi.org/10.1175/JAS-D-11-0142.1

Hung, M. P., Lin, J. J., Wang, W., Kim, D., Shinoda, T., & Weaver, S. J. (2013). MJO and convectively coupled equatorial waves simulated by CMIP5 climate models. Journal of Climate, 26(7), 6185–6214. https://doi.org/10.1175/JCLI-D-12-00541.1

Iacono, M. J., Delamere, J. S., Mlawer, E. J., Shephard, M. W., Clough, S. A., & Collins, W. D. (2008). Radiative forcing by long-lived greenhouse gases: Calculations with the AER radiative transfer models. Journal of Geophysical Research, 113, D11310. https://doi.org/10.1029/2007JD009544

Kemble-Cook, S., & Wang, B. (2001). Equatorial waves and air–sea interaction in the boreal summer intraseasonal oscillation. Journal of Climate, 14(13), 2923–2942. https://doi.org/10.1175/1520-0442(2001)014<2923:EWASAI>2.0.CO;2

Kemble-Cook, S., Wang, B., & Fu, X. (2002). Simulation of the intraseasonal oscillation in the ECHAM-4 model: The impact of coupling with an ocean model. Journal of the Atmospheric Sciences, 59(9), 1433–1453. https://doi.org/10.1175/1520-0469(2002)059<1433:TOASAI>2.0.CO;2

Kim, D., Spengler, K., Stern, W., Waliser, D., Kang, I., Maloney, E. D., ... Zhang, G. (2009). Application of MJO simulation diagnostics to climate models. Journal of Climate, 22(23), 6413–6436. https://doi.org/10.1175/2009JCLI3063.1

Kim, D. J., Kiladis, G. N., Mapes, B. E., Weckmann, K. M., Spengler, K. R., Lin, W., ... Scinocca, J. F. (2006). Tropical intraseasonal variability in 14 IPCC AR4 climate models. Part I: Convective signals. Journal of Climate, 19(12), 2665–2690. https://doi.org/10.1175/JCLI3735.1

Madden, R. A., & Julian, P. R. (1972). Description of global-scale circulation cells in the tropics with a 40–50 day period. Journal of the Atmospheric Sciences, 29(5), 702–708. https://doi.org/10.1175/1520-0469(1972)029<0702:DGDSCI>2.0.CO;2

Madden, R. A., & Julian, P. R. (1971). Detection of a 40–50 day oscillation in the zonal wind in the tropical Pacific. Journal of the Atmospheric Sciences, 28(5), 902–908. https://doi.org/10.1175/1520-0469(1971)028<0902:DADOI>2.0.CO;2

Madden, R. A., & Julian, P. R. (1971). Description of global-scale circulation cells in the tropics with a 40–50 day period. Journal of the Atmospheric Sciences, 28(6), 1109–1110. https://doi.org/10.1175/1520-0469(1972)029<1109:DOGSC>2.0.CO;2

Maloney, E. D. (2009). The moist static energy budget of a composite tropical intraseasonal oscillation in a climate model. Journal of Climate, 22(3), 711–729. https://doi.org/10.1175/2008JCLI2542.1

Maloney, E. D., & Sobel, A. H. (2004). Surface fluxes and ocean coupling in the tropical intraseasonal oscillation. Journal of Climate, 17(22), 4368–4386. https://doi.org/10.1175/JCLI-3212.1
Zeng, Z., Ho, S. P., Sokolovskiy, S., & Kuo, Y. H. (2012). Structural evolution of the Madden-Julian oscillation from COSMIC radio occultation data. *Journal of Geophysical Research*, 117, D22108. https://doi.org/10.1029/2012JD017685

Zhang, C. (2005). Madden-Julian oscillation. *Reviews of Geophysics*, 43, RG2003. https://doi.org/10.1029/2004RG000158

Zhang, C. (2013). Madden–Julian oscillation: Bridging weather and climate. *Bulletin of the American Meteorological Society*, 94(12), 1849–1870. https://doi.org/10.1175/BAMS-D-12-00026.1

Zhang, C., Gottschalck, J., Maloney, E. D., Moncrieff, M. W., Vitart, F., Waliser, D. E., . . . Wheeler, M. C. (2013). Cracking the MJO nut. *Geophysical Research Letters*, 40, 1223–1230. https://doi.org/10.1002/grl.50244

Zhang, D., & Anthes, R. A. (1982). A high-resolution model of the planetary boundary layer—Sensitivity tests and comparisons with SESAME-79 Data. *Journal of Applied Meteorology and Climatology*, 21(11), 1594–1609. https://doi.org/10.1175/1520-0450(1982)021<1594:AHRMOT>2.0.CO;2