Thermodynamics of the Madden–Julian Oscillation in a Regional Model with Constrained Moisture

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ABSTRACT

To identify the main thermodynamic processes that sustain the Madden–Julian oscillation (MJO), an eddy available potential energy budget analysis is performed on a regional model simulation with moisture constrained by observations. The model realistically simulates the two MJO episodes observed during the winter of 2007/08. Analysis of these two cases shows that instabilities and damping associated with variations in diabatic heating and energy transport work in concert to provide the MJO with its observed characteristics. The results are used to construct a simplified paradigm of MJO thermodynamics.

Furthermore, the effect of moisture nudging on the simulation is analyzed to identify the limitations of the model cumulus parameterization. Without moisture nudging, the parameterization fails to provide adequate low-level (upper level) moistening during the early (late) stage of the MJO active phase. The moistening plays a critical role in providing stratiform heating variability that is an important source of eddy available potential energy for the model MJO.

1. Introduction

The Madden–Julian oscillation (MJO) is one of the most important components of tropical intraseasonal variability. Discovered by Madden and Julian (1971), this oscillation is an equatorial planetary-scale envelope of several complex multiscale processes. It originates over the Indian Ocean and propagates eastward across the Maritime Continent and western Pacific at a speed of about 5 m s\(^{-1}\). A number of observational studies have provided important insights into the evolution of various fields associated with MJO propagation and several theories have been proposed to explain its various characteristics. However, a comprehensive theory that explains its important observed characteristics and climate models that satisfactorily simulate them remains elusive.

In this section only a brief summary of the relevant results from observational, theoretical, and modeling studies is presented. For extensive reviews of decades of MJO research, the reader is referred to Zhang (2005), Lau and Waliser (2005), and the references therein. One of the earliest theories proposed to explain large-scale intraseasonal variability is a mechanism often referred to as wave conditional instability of the second kind (wave-CISK; Lindzen 1974; Lau and Peng 1987). In this theory, convective heating and moisture convergence interact to create unstable modes. This theory has been refined to account for inconsistencies with observations such as the fact that the smallest wavelength waves are most unstable and the propagation speed of the waves involved in this mechanism is 3–4 times that of MJO. The refinements include the effect of friction on moisture convergence and appear to alleviate the above problems, resulting in slowly propagating eastward planetary-scale oscillations comparable to observations (Wang 1988; Salby et al. 1994). Furthermore, the inclusion of Rossby waves in the dynamics results in wave packets that propagate at slower speeds than convectively coupled Kelvin waves. Despite its success in producing MJO-like oscillations in some analytical models, the theory has also been criticized on observational and theoretical grounds (Raymond 1994). Wu (2003) proposed a shallow CISK
mechanism through which the CISK operates via shallow heating to provide moisture for an eventual triggering of deep convection that uses up the moisture without a positive feedback onto the low-level moisture convergence. This paradigm assumes that deep convection is triggered after the low-level moistening somehow eroded the stable cap above the shallow convection. An important feature of wave-CISK-like mechanisms is the covariance of heating (vertical motion) and temperature, through which eddy available potential energy (EAPE) is generated to feed the wave growth.

On the flip side of shallow-CISK processes is what is referred to as stratiform instability. In this mechanism, the baroclinic mode of temperature perturbation interacts with stratiform heating, resulting in instability (Mapes 2000; Kuang 2008). This process has been proposed to prolong the lifetime of convective activity and seems to exist during the early and late stages of MJO active phase according to a composite MJO study by Benedict and Randall (2007).

Perhaps the most widely accepted theory of MJO dynamics is what is often referred to as a “discharge-recharge” process. According to this theory, shallow cumuli gradually moisten lower troposphere and that leads to development of deep heating and drying, after which radiative cooling and low-level frictional convergence once again and moisten (recharge) the lower troposphere (Bladé and Hartmann 1993; Hu and Randall 1994). Yet, other theories point to radiation feedback (Raymond 2001; Bony and Emanuel 2005), and wind-induced surface heat exchange (WISHE; Emanuel 1987) to explain some characteristics of MJO. The processes through which MJO gets its observed scale, propagation speed, and energy sources for its convective activity are far from understood. As suggested by Benedict and Randall (2007), it is quite likely that the mechanisms discussed above are in action at various stages of the MJO life cycle, and that the MJO process is just too complex for one simple linear or quasi-linear theory to explain.

A practical manifestation of our lack of understanding of MJO dynamics is the difficulty faced by current climate models in simulating it correctly (Waliser et al. 2003; Lin et al. 2006, Kim et al. 2008). Results of several decades of effort on the simulation of MJO portray a lack of fundamental understanding of the interactions of moist convection with the environment and their proper representation in the models (Tokioka et al. 1988; Wang and Schlesinger 1999; Maloney and Hartmann 2001). Several cases have been reported where a certain closure would improve the simulation of MJO in a certain model and would have an opposite effect in another. For example, Wang and Schlesinger (1999) showed that a large relative humidity threshold in convective parameterizations is critical for the amplification of MJO by causing a time lag between condensational heating and large-scale convergence, while Community Climate Model (CCM3.6) experiments by Maloney and Hartmann (2001) indicated that the model’s MJO signal is not sensitive to the relative humidity threshold. Recently Fu and Wang (2009) showed that, by providing additional eddy available potential energy, increased stratiform heating improved the simulation of the MJO in the European Centre for Medium-Range Weather Forecasts (ECMWF) Hamburg Atmospheric Model (ECHAM), while similar increase in stratiform heating in a Chinese version of the Flexible Global Ocean Atmosphere Land System (FGOALS) model (Li et al. 2009) resulted in an MJO with an unrealistically fast phase speed. The MJO has also been studied using regional models forced by observed lateral and surface conditions to diagnose the processes responsible for its initiation (Ray et al. 2009; Gustafson and Weare 2004). These studies suggest that stochastic high-frequency forcing from the subtropics may play important roles during the early stages of MJO development.

This study aims to contribute to bridging the gap among theory, observations, and modeling of the MJO. The strategy involves observationally constraining the moistening process in a model such that other relevant aspects of MJO are correctly simulated. The results of the simulation are then utilized toward our objective in two ways:

1) Thermodynamic budget analysis is used to identify the main sources and sinks of energy associated with MJO convective activity and the roles of the various types of instabilities in the life cycle of the MJO.

2) By systematically evaluating the effect of the moistening constraint on the model simulation, an insight is gained into the physical processes that are critical for a robust, realistic MJO in models.

The advantage of performing thermodynamic budget analysis with an observationally constrained model output, as opposed to pure observations or pure model output, is that on the one hand the variables necessary for a closed energy budget analysis are often not available in the observations, and on the other hand models without observational constraints rarely produce a convincing MJO. The technique used in this study aims to alleviate both these problems. In this study we focus on an already initiated and propagating MJO signal.

2. Model, experiment design, and evaluation

a. Model description and setup

The model used in this study is the Advanced Research Weather Research and Forecasting (WRF) model version...
3.1 (ARWRF3.1; Skamarock et al. 2008). It is run with the Kain–Fritsch cumulus scheme (Kain and Fritsch 1990), the Rapid Radiative Transfer Model (RRTM; Mlawer et al. 1997) for longwave radiation, the Dudhia (1989) scheme for shortwave radiation, and the Yonsei University PBL scheme (Hong et al. 2006). The Noah land surface model (LSM; Mitchell et al. 2000) is used to represent land surface processes. The microphysics scheme used in these experiments is the WRF single-moment microphysics three-class scheme (WSM3; Hong et al. 2004).

Initial, lateral, and surface boundary conditions including SSTs are derived from the National Centers for Environmental Prediction’s (NCEP’s) Global Forecast System (GFS) final analysis and no filtering is applied on them. Furthermore, in the case where the moisture is constrained, the moisture tendency in the model is nudged toward this global analysis. Figure 1 shows the model domain. It includes the Indian Ocean, the Maritime Continent, and the western Pacific. Meridionally, the domain extends between 22.5°S and 22.5°N and includes the South Asian and Australian monsoon regions. Longitudinally it ranges from 45° to 175°E. The dashed lines mark the 7.5°S and 7.5°N latitudes between which meridional averages of all the fields in the subsequent analyses are performed. The model grid spacing is set at 36 km. The simulations run between 1 October 2007 and 31 January 2008, during which two consecutive episodes of MJO were observed.

b. Experiments

The two experiments analyzed in the subsequent sections share all of the above experimental setup. In the control experiment (hereafter NONUDGE) the only observational constraints are the initial, lateral, and surface boundary conditions. In the constrained case, the setting is identical except that moisture at every grid point and at all levels above the PBL is nudged toward the GFS analysis every 6 h with a nudging time scale of 15 min using the WRF’s grid four-dimensional data assimilation utility. In a test experiment, nudging the PBL moisture was found to have little effect and was thus unnecessary for the purpose of the study. This experiment will be referred to as the GFDDA case. The rationale for this experimental design is that while the moisture budget is not closed because of the nudging term, the variations in diabatic heating and temperature (which are the focus of this study) remain consistent. In other words, the thermodynamic equation remains balanced for the given diabatic heating, while the overall model dynamics and thermodynamics approximate the observations, as will be shown in the subsequent subsection. Furthermore, as noted in the introduction, this enables us to identify the limitations of moistening process in the model by directly analyzing what processes the nudging in the GFDDA experiment compensates for.

c. Evaluation

Evaluation of the model performance in simulating the MJO is performed by (i) direct comparison of Hovmöller diagrams of outgoing longwave radiation (OLR) and zonal wind with those from observations and (ii) a univariate empirical orthogonal function (EOF) analysis of the OLR signal based on the recommendations of Climate Predictability and Variability (CLIVAR) MJO working group for the first-level diagnostics of the MJO in models (Waliser et al. 2009). The datasets used for evaluating the model are the National Oceanic and Atmospheric Administration (NOAA)–Climate Prediction Center (CPC) interpolated OLR (Liebmann and Smith 1996), the NCEP–Department of Energy (DOE) reanalysis (Kanamitsu et al. 2002), and the interim ECMWF Re-Analysis (ERA-Interim; Simmons et al. 2007).

Figure 2 shows the comparison of OLR from the two experiments with those from NCEP–DOE reanalysis and NOAA–CPC data. Note that the NCEP–DOE reanalysis OLR is model produced while that from NOAA–CPC is from satellite measurements. In the NONUDGE case, OLR is rarely below 200 W m⁻² whereas in the nudging case it gets lower than 100 W m⁻², in much better agreement with the NOAA–CPC observation, albeit slightly lower. The OLR signal in the NONUDGE case is weakly organized in comparison to those of the observations or the GFDDA case.

The MJO is also manifested as an eastward propagating westerly low-level wind signal in an otherwise
westerly trade wind environment. Figure 3 shows the zonal wind at 850 hPa from the two model experiments and the two global reanalyses. The eastward propagation of the westerly winds exists in both model experiments as well as the reanalyses. But the smaller-scale features in the GFDDA case show better agreement with reanalyses than the NONUDGE case. In particular, the westerly winds from the NONUDGE case at 120°E during late October and at 60°E during late November are stronger, once again making the standing signal more robust than those of the GFDDA case and the reanalyses. The return flow associated with this eastward propagating westerly wind signal are upper tropospheric easterlies that propagate eastward. The easterlies at 200 hPa are depicted in Fig. 4. In general the upper-level winds from both model experiments are in good agreement with the reanalyses. In the NONUDGE case, however, the easterlies are once again dominated by a more or less standing signal. The weak easterly signal near late November that separates the two MJO episodes is more visible in the GFDDA case and the reanalyses. The opposite signs of the eastward propagating zonal wind at 850 and 200 hPa reveal the baroclinic nature of MJO signals.

Univariate EOF analysis is performed on the OLR data from the two model simulations and the NOAA–CPC interpolated data after they are averaged between 7.5°S and 7.5°N and bandpass filtered between 20- and 90-day
periods. Figures 5a and 5b show the first and second EOFs. The variance explained by the two modes is 34% and 29% for NONUDGE, 49% and 25% for GFDDA, and 43% and 35% for NOAA–CPC OLR. Once again the zonal structures of the EOFs of the GFDDA experiment are in better agreement with the observation. Figure 5c shows the MJO index as defined by the sum of the squares of the first two principal components ($PC_1^2 + PC_2^2$). The index is greater than 1 for most of the simulation period, which represents a particularly robust MJO signal in both model simulations as well as the observation. This is of course expected for the simulation domain, and simulation times are selected with that purpose in mind. Finally, Fig. 5d shows the lag-correlation between the first two principal components (PCs). The correlation pattern is another indication of the coherent propagation of the signals. Comparison of Fig. 5 with Fig. 10 of Waliser et al. (2009) provides further evidence that the episodes presented here bear all the significant characteristics of a typical robust MJO signal.
d. Composite analysis

In the previous subsection it was shown that the GFDDA experiment reproduces several characteristics of the two observed MJO episodes to justify the assumption that the thermodynamic processes in the model sufficiently represent those of the observed MJO episodes. In this subsection the method of compositing used in this study is briefly described. As shown by the solid lines in Fig. 2, the observed pair of MJO episodes propagates at approximately 4 m s\(^{-1}\). To obtain a robust signal of the evolution of a variable during the MJO life cycle, one has to produce a composite. Even though it is well documented that the MJO structure shows variations as it propagates eastward across the Maritime Continent (Kiladis et al. 2005), we neglect these variations in favor of constructing a robust composite MJO using the entire longitudinal model domain for the purpose of obtaining first-order understanding of its thermodynamics. Thus, the underlying assumptions in the composite analysis are that the MJO propagates at a constant speed over the longitudinal domain of the model and there are no
inherent differences among the longitudinal and temporal points other than the differences in the phases of the MJO. In other words, suppose a particular phase of the MJO, say peak precipitation (pcp), is observed at points \((t_1, x_1)\) and \((t_2, x_2)\). Then the MJO is propagating at speed \(c = (x_2 - x_1)/(t_2 - t_1)\). Therefore, from a reference frame of a hypothetical point \(x_c\) that moves at the phase speed of the MJO, the variable (precipitation in this case) is \(pcp = pcp[t, x_1 + c(t - t_1)]\). This transformation essentially compresses the hundreds of snapshots of the MJO from the entire longitudinal domain into a 4-month-long time series from a single (hypothetical) point. The reference point \((t_1, x_1)\) is taken to be 14 October and 63°E. Figure 6 shows the evolution of precipitation, convective latent heating, and perturbation potential temperature from the reference frame of a point moving with the MJO at \(c = 4\) m s\(^{-1}\). The two MJO peak precipitation times are on days 34 and 81 and are marked by the dashed vertical lines. These lines also more or less coincide with the peak heating, which is also in phase with the positive perturbation potential temperature (potential temperature minus its 4-month mean). The subsequent analysis is presented in this framework.

e. Eddy available energy budget analysis

As a brief introduction to the concept of eddy available potential energy, consider a simple system where periodic heating \(Q(t)\) is applied to a body of heat capacity of \(C\). The temperature of the body would vary as

\[
\frac{dT}{dt} = \frac{Q(t)}{C}.
\]

(1)

The equation for the amplitude of a specific mode of temperature oscillation is then given by

\[
\frac{dT' T''}{dt} = \frac{2T' Q'}{C},
\]

(2)

where the primes indicate the fluctuations from the mean and the averaging (denoted by the overbar) is performed.

FIG. 5. (a) First and (b) second EOFs of OLR from the two simulations and NOAA–CPC interpolated data. (c) MJO index and (d) cross correlation of the associated principal components.
over the period of the mode. Now consider a situation where the perturbation heating and the temperature anomaly are positively correlated. In other words, the external heating selectively heats up the body when it is hot and cools it down when it is cold. In this case the amplitude of the temperature of the body would grow; on the other hand, if the external heating selectively cools it down when it is warm and warms it up when it is cold, the amplitude would decay. Now assume the body is a parcel of air in the atmosphere and this heating with positive covariability with the temperature is being applied only to the parcel but not to the environment. In this case we say eddy available potential energy is being supplied to the system. Since the parcel will be periodically cooler (denser) and warmer (lighter) than its environment, this results in rearrangement of mass.

**FIG. 6.** Composites of (a) precipitation (mm day$^{-1}$), (b) convective latent heating (K day$^{-1}$), and (c) potential temperature perturbations (K).
The thermodynamic budget equation can be written as

$$\frac{\partial \theta}{\partial t} = -\nabla \cdot (\mathbf{v} \theta) + Q_{1(\text{convective})}$$

$$+ Q_{1(\text{stratiform})} + Q_{1(\text{radiation})} + Q_{1(pbl)}$$

(3)

where $\theta$ is potential temperature and $\mathbf{v} = (u, v, \omega)$ is the three-dimensional wind on pressure surfaces. The four terms on the right-hand side correspond to potential temperature tendencies due to convective latent heating, stratiform (resolved) latent heating, radiation, and turbulent heat fluxes. Note that the potential temperature tendencies are the actual heating multiplied by $\theta/c_p T$, where $c_p$ is the heat capacity of air at constant pressure and $T$ is the temperature. Unlike the moisture budget, which is not closed (does not conserve moisture) because of nudging, the thermodynamic budget is closed for the given diabatic heating, which is indirectly constrained by the moistening, while there is no nudging term in the thermodynamic equation. The energy budget equation is directly written out from the model in closed form; therefore, there is also no residual term associated with errors in the calculation.

One of the advantages of using potential temperature over temperature is the simplicity of the advection term on pressure surfaces. The equation that governs the evolution of amplitude of potential temperature fluctuations about the mean (in this case the 4-month mean) is given by

$$\frac{\partial \theta^* \theta^*}{\partial t^*} = -\bar{\theta} [\nabla \cdot (\bar{\mathbf{v}} \bar{\theta})] + \bar{\theta} \theta^*Q_{1(\text{convective})}$$

$$+ \bar{\theta} \theta^*Q_{1(\text{stratiform})} + \bar{\theta} \theta^*Q_{1(\text{radiation})} + \bar{\theta} \theta^*Q_{1(pbl)}$$

(4)

The primes indicate the deviation of the terms from their time mean value for every point in the domain and the overbars represent latitudinal averaging, which is taken as a proxy for the scale of the MJO. The relationship between this equation and the eddy available potential energy equation often used in global energy budget analyses deserves some discussion. The equation is often integrated in space, but in this case, we are interested in locations as well as time of the generation and destruction of EAPE, so integration in space is not useful. So, while we will continue to refer to this equation as the EAPE tendency, one has to keep in mind that this is a “local” form useful for wave energy analysis as opposed to the integrated form often used in global energy budget analyses. The implicit assumption is that the time mean state of the parcel under consideration is representative of the state of its environment at any time (see Hsu et al. 2009; Maloney and Esbensen 2003; Yanai et al. 2000).

A brief discussion of the physical meaning of Eq. (4) in the general context of wave theory is provided here. The first term on the right is EAPE generation by the threedimensional transport of heat; if a wave propagates in such a way that it transports heat from a cold region to a warm region (warm region to cold region) then it is generating (destroying) EAPE. Thus, it represents conversion from eddy kinetic energy (EKE) to eddy available potential energy (and vice versa). This term incorporates both transport of EAPE as well as generation of EAPE by the interaction of transported heat with local temperature fluctuations. This term incorporates several important processes that will be discussed in some detail in a section 3b.

The last four terms on the right-hand side of Eq. (4) represent diabatic heating processes that could generate (or damp) EAPE depending on their covariance with potential temperature perturbations. The first one represents a wave-CISK-like mechanism where temperature (and wind convergence) waves that are in phase with the heating grow. When the term is negative it represents convective damping (Wang 2006; Yanai et al. 2000) where the EAPE created by some other processes is compensated by convective heating over the relatively colder regions, so the contrast in perturbation potential temperature is reduced. The next term on the right-hand side of Eq. (4) represents EAPE generation by “stratiform instability.” For example, when a stratiform heating perturbation with heating in the upper troposphere and cooling at lower troposphere has a positive covariance with the baroclinic mode of potential temperature variability, the baroclinic mode grows. Similarly when covariance of radiative processes with potential temperature (e.g., through radiation moisture feedback; Bony and Emanuel 2005) exists, EAPE can be generated or destroyed depending on the nature of the covariance. Finally, waves that modulate the surface heat fluxes across the planetary boundary layer (e.g., through WISHE) in such a way that the potential temperature fluctuations are in phase with the surface heat fluxes would lead to EAPE generation according to the last term in Eq. (4).
3. Thermodynamics of the MJO

a. Eddy available potential energy budget of the MJO in the model

EAPE tendency calculations are performed using 6-hourly model outputs of the potential temperature and the thermodynamic budget equation term from the model. The fluctuations are deviations from the 4-month (the length of the simulation) average. Once the EAPE budget equation terms are calculated, they are latitudinally averaged and composited as discussed in section 2d. Figure 7 shows the four dominant terms in the EAPE budget equation. According to Fig. 7a, convective heating has two distinct roles in the EAPE budget. In the upper troposphere, where EAPE generation is positive, it generates EAPE and wave-CISK-like processes dominate. But in the lower troposphere shallow heating is a sink of EAPE as convective damping prevails. Generation of EAPE by convective heating is strong around the times of the precipitation peak (the dashed lines) as well as when there is less precipitation (heating) and the troposphere is particularly cool (the first 5 days). A consequence of the EAPE damping by shallow heating and EAPE generation by deep heating is that there is a level (marked by solid dark curve) where convective heating has no effect on EAPE generation. EAPE generation by stratiform heating is positive throughout the troposphere (Fig. 7b), suggesting that there is covariance of stratiform heating with the baroclinic mode of potential temperature fluctuations. However, the EAPE generation is quite weak in the lower troposphere but strong between 500 and 300 hPa. The fact that it shows relatively less temporal variability suggests that its variability more or less matches the potential temperature fluctuations associated with the MJO (Fig. 6c).

Radiative heating plays a minor role in the EAPE budget, in general, because it is negatively correlated
with temperature, as one would expect (Fig. 7c), but the negative correlation is weakened during precipitation peaks, likely because of the radiative effect of moisture as suggested by Bony and Emanuel (2005). Most of the EAPE generated by convective and stratiform heating in the upper troposphere is converted to eddy kinetic energy (Fig. 7d). At the lower troposphere there is conversion from eddy kinetic energy to eddy potential energy instead. In other words, the lower troposphere is destabilized mechanically by the transport of heat.

The nature of the balances in the EAPE budget is shown in Fig. 8. In the lower troposphere (Fig. 8a), heat transport and stratiform instability generate EAPE, which is damped by shallow heating. Note that heat transport provides at least half of the initial EAPE; however, it decays quickly and stratiform instability sustains the shallow heating for up to 10 days beyond the day of precipitation maximum. In the upper troposphere, stratiform instability and wave-CISK-like processes generate EAPE, which is converted to kinetic energy (Fig. 8b).

The fact that shallow convection damps EAPE and deep convection generates it has an important implication. At the level of transition, the kinetic energy necessary for the transport of heat (and moisture) across the transition level (where EAPE generation by convective heating is zero) has to be supplied by some other mechanism than convective latent heating. Thus, analysis of the EAPE budget at this transition level would provide us with insight into what is triggering the transition from shallow to deep convection. Figure 8c shows the EAPE budget terms calculated at this level of transition by interpolating between the levels above and below the transition level. The conversion to kinetic energy follows the EAPE generation by stratiform variability. The increase in stratiform heating is related to the fact that the transition level is progressing upward to the levels where stratiform heating dominates. Once deep convective heating is triggered, the transition level decreases its elevation (Fig. 7a) and stratiform heating decreases as well (Fig. 7b). In simple terms the transition level is gradually climbing upward until it reaches the level at which there is a large EAPE generation available for conversion to KE to trigger deep convection.

b. EAPE transport in the MJO

While stratiform instability plays an important role in the transition from shallow to deep convection and it sustains the shallow heating well afterward, the EAPE that forces the shallow heating at the early stages of the MJO is supplied by transport (conversion from KE; Figs. 7d and 8a). Therefore, understanding of the processes that generate EAPE through transport of heat between the lower and upper troposphere is critical to the closure of the MJO EAPE cycle.

Fig. 8. Sources and sinks of EAPE (K£2 km£2 day£1) averaged (a) between the surface and the transition level (solid curve in Fig. 5a), (b) between 200 hPa and the transition level, and (c) at the transition level.
The first term on the right-hand side of Eq. (4) can be expanded into its zonal, meridional, and vertical components as follows:

\[
-\theta' [V \cdot (\mathbf{v} \theta)] = -\theta' \frac{\partial (u \theta)}{\partial x} - \theta' \frac{\partial (v \theta)}{\partial y} - \theta' \frac{\partial (w \theta)}{\partial p}.
\]

Figure 9 shows the contributions of the three terms to the total EAPE generation by heat transport into the lower troposphere of the MJO region. Vertical transport is the dominant term while both zonal and meridional components transport EAPE away from the region. Similarly, at the upper troposphere most of the EAPE generated by latent heating is consumed by the vertical component of EAPE generation by transport.

Figure 10 shows the contributions of the upward and downward parts of this vertical transport. The downward component oscillates up and down with the MJO cycle as the transition line does to a lesser extent. The oscillation in the upward component is less apparent. The triggering of deep heating is followed by subsidence that coincides with a temperature minimum. The subsidence of cold air, presumably associated with mesoscale downdrafts, at the low levels gradually retreats upward until the convective damping layer is deep enough for the next MJO deep heating to be triggered.

In this section the EAPE budget analysis was performed on a composite MJO. According to the results of
the analysis, shallow heating destroys EAPE generated by transport and stratiform instability, while deep convection contributes to the generation of EAPE during the active phase of MJO. Stratiform instability plays an important role in the transition from shallow to deep convection as well as in sustaining the shallow convective heating well beyond the day of precipitation maximum. Analysis of the transport processes showed that subsidence of cold air is the main component of the EAPE generation by heat transport in the lower troposphere. This subsidence forces shallow convection and heating, which in turn deepens the layer of convective damping toward the layer of stratiform instability upon which deep convection is triggered.

4. The role of the nudged moistening

As discussed in the introduction, the model’s improved simulation of the two MJO episodes is due to the fact that moisture is frequently nudged toward the GFS reanalysis. A perfect model would provide the correct moistening profile and variability on its own. Nevertheless, looking at how the moisture nudging process influences the model is a good starting point for understanding the moistening process, helping guide model improvements by pointing out its limitations.

Figure 11 shows the role of the perturbation moisture nudging term during the life cycle of the two MJO episodes. The moistening from the nudging takes place about 15 days before the MJO peak precipitation and it moistens the upper troposphere for up to 15 days after the precipitation maxima. There is a low-level drying immediately following the precipitation peak. A similar drying is also noted in analysis of observations (Benedict and Randall 2007; Houze et al. 2000). The pairs of early low-level moistening and upper-level moistening during and after peak precipitation results in the peculiar tilt in the moistening profile.

The effect of moisture nudging on the evolution of diabatic heating is most significant on the stratiform heating and is displayed in Fig. 12, which shows the stratiform variability in the NONUDGE and GFDDA experiments. The absence of stratiform heating before the precipitation peak (“cooling” relative to the environment) and its presence immediately following the precipitation peak in the GFDDA case is consistent with the well-known tilt in stratiform heating variability (Lin et al. 2004). As discussed in the last section, this stratiform variability interacts with the potential temperature perturbations (Fig. 6c), resulting in a more or less constant stratiform instability and EAPE generation between 500 and 300 hPa (Fig. 7b). This tilt is essentially absent in the NONUDGE case. An important point to take from the analysis of the nudged moistening and the associated stratiform variability is the fact that the model lacks proper shallow as well as deep moistening mechanism. This ultimately points to the convective parameterization’s
inability to correctly represent the full life cycle of convective processes that start out as shallow convection and may eventually evolve to deep convection and stratiform cloud with the correct time scales.

5. Conclusions

One of the central problems in the study of MJO is the nature of the instabilities involved in providing the oscillation in its energy and its time and space scales. This study aims to contribute to the efforts of addressing these issues using a regional model whose moisture is constrained by observations. This approach circumvents the main challenges in performing a complete MJO thermodynamic budget analysis from observations because not all the necessary variables are readily available, and from models, which do not simulate the MJO realistically. Indeed, several sensitivity experiments performed using WRF with different convective and cloud microphysical parameterizations produced results similar to the NONUDGE case, although the OLR values are quite sensitive to the cloud microphysical parameterizations.

In this study two consecutive MJO episodes observed during the winter of 2007/08 are considered. To identify the roles of various instabilities during the life cycles of the two MJO episodes, an eddy available potential energy budget analysis is performed. The roles of the various mechanisms are analyzed and a simplified paradigm of MJO thermodynamics is constructed. This paradigm is summarized in Fig. 13. According to the model results, there are two distinct tropospheric layers with different thermodynamic roles. The upper one is the “stratiform instability layer” between 500 and 300 hPa (Fig. 7b, represented by the shading in Fig. 13) where EAPE is constantly generated by the covariance between stratiform heating and potential temperature. The second is the “convective damping layer.” In this layer, below the solid curve in Fig. 7a (also represented in Fig. 13), the EAPE generated by conversion from EKE (Fig. 7d) and stratiform instability is damped by shallow convective heating, which is also moistening the layer (Fig. 11). The curve in Fig. 13 marks the level at which EAPE generation by convective latent heating is zero. If air parcels are to rise above that layer, some other processes have to provide the EKE. The layer of convective damping is gradually deepening, primarily because of EAPE generation by subsidence (Fig. 10a). As it rises and reaches the stratiform instability layer, this mechanism of EAPE generation becomes available (Fig. 8c) and allows the air parcels and moisture to rise deep into the upper troposphere, triggering deep convection. Once deep convection is triggered, the wave-CISK-like mechanisms of generating EAPE take over and the damping layer rapidly shrinks back to the lower troposphere where the subsidence once again starts to deepen it for the next MJO cycle. This deepening of the convective damping layer is accompanied by low-level moistening [i.e., buildup of moist static energy (recharge)], which is discharged in the form of conversion from EAPE to EKE through convective latent heating, as suggested by the recharge–discharge hypothesis of Bladé and Hartmann (1993) and Hendon and Liebmann (1990). The upward transport of moisture is mimicked by the moisture nudging term (Fig. 11); it is also manifested in the gradual deepening of the layer of damping by shallow convective heating (Fig. 7a).

It is important to note that these processes take place in an environment where there is generation of EAPE
by upward fluxes of EAPE (Fig. 10b) primarily by low-level wind convergence. Additionally, however, stratiform instability also contributes to the generation of low-level EAPE that sustains the shallow heating well after the EAPE generation by transport is exhausted (Fig. 8a). The interactions between potential temperature perturbations and radiative cooling as well as PBL fluxes are found to be not as significant. Therefore, the proposed hypotheses involving radiation feedback mechanism and WISHE do not appear to play significant roles in the MJO EAPE budget. According to this specific modeling study, MJO convection is forced from the top by subsidence rather than from the surface, given the sufficiently moist boundary layer.

Analysis of the effect of the nudged moistening is important not only to understand the MJO moistening process but also to diagnose the roots of the shortcomings of the model. Lower-level (upper-level) moistening at the early (late) stages of the MJO active phase (Fig. 11) is found to be crucial for the life cycle of the MJO. This introduces the tilt in stratiform heating variability that interacts with potential temperature perturbations (Figs. 6 and 12) that is an important source of EAPE (Fig. 7).

Most of the processes identified in this study have been shown in various observational and modeling analyses. The tilt in diabatic heating associated with the fact that stratiform heating dominates in the later period of MJO cycle (Lin et al. 2004); the low-level heating and moistening associated with shallow convection and the subsequent low-level drying as well as the covariance between temperature and stratiform heating at later stages of the MJO cycle (Benedict and Randall 2007), EAPE generation by covariance between MJO deep heating and temperature in the upper troposphere (Maloney and Esbensen 2003), and convective damping in the lower troposphere (e.g., Yanai et al. 2000; Wang 2006; etc.) have been observed. The results of this study depict a coherent synthesis of the MJO EAPE cycle consistent with observations and contribute to reconciling the theories forwarded to explain the various observed aspects of the thermodynamics of MJO. This study has also provided guidance on the requirements for both shallow and deep convective parameterizations working in tandem for improved simulations of the MJO.

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REFERENCES


Wang, W., and M. E. Schlesinger, 1999: The dependence on convective parameterization of the tropical intraseasonal oscillation simulated by the UIUC 11-layer atmospheric GCM. J. Climate, 12, 1423–1457.

