Moist Thermodynamics of the Madden–Julian Oscillation in a Cloud-Resolving Simulation

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ABSTRACT

The moist thermodynamic processes that determine the time scale and energy of the Madden–Julian oscillation (MJO) are investigated using moisture and eddy available potential energy budget analyses on a cloud-resolving simulation. Two MJO episodes observed during the winter of 2007/08 are realistically simulated. During the inactive phase, moisture supplied by meridional moisture convergence and boundary layer diffusion generates shallow and congestus clouds that moisten the lower troposphere while horizontal mixing tends to dry it. As the lower troposphere is moistened, it becomes a source of moisture for the subsequent deep convection during the MJO active phase. As the active phase ends, the lower troposphere dries out primarily by condensation and horizontal divergence that dominates over the moisture supply by vertical transport. In the simulation, the characteristic time scales of convective vertical transport, mixing, and condensation of moisture in the midtroposphere are estimated to be about 2 days, 4 days, and 20 h respectively. The small differences among these time scales result in an effective time scale of MJO moistening of about 25 days, half the period of the simulated MJO. Furthermore, various cloud types have a destabilizing or damping effect on the amplitude of MJO temperature signals, depending on their characteristic latent heating profile and its temporal covariance with the temperature. The results are used to identify possible sources of the difficulties in simulating MJO in low-resolution models that rely on cumulus parameterizations.

1. Introduction

Discovered by Madden and Julian (1971), the Madden–Julian oscillation (MJO) is an important component of tropical intraseasonal variability. Originating over the Indian Ocean, MJO is an equatorial planetary-scale envelope of complex multiscale cloud systems that propagates eastward at a speed of about 5 m s⁻¹ across the Maritime Continent and western Pacific. It gradually degenerates into a fast-moving dry Kelvin wave as it crosses the date line. The MJO has far reaching impact on weather systems within and outside the tropics including the onset and break of the Asian and Australian summer monsoons (e.g., Yasunari 1979; Hendon and Liebmann 1990; Wheeler and McBride 2005; Goswami 2005) and the formation of tropical cyclones (e.g., Liebmann et al. 1994; Bessafi and Wheeler 2006). It is also believed to influence the onset and demise of some El Niño events (e.g., Kessler et al. 1995; McPhaden 2008).

Decades of observational studies have documented the evolution of various fields associated with MJO propagation and several theories have been proposed to explain its characteristics. However, a comprehensive theory that explains all the important observed characteristics, especially its sources of energy and its peculiar propagation speed is lacking. Hence, proper parameterization of subgrid-scale dynamic and moist thermodynamic processes that enables a satisfactory simulation of MJO remains a challenge for global climate models. For extensive review of MJO research, the reader is referred to Zhang (2005), Lau and Waliser (2005), and the references therein, here only the most relevant recent studies on MJO moist thermodynamic processes and parameterization are discussed.

Several studies have shown that the characteristics of a simulated MJO in global models are very sensitive to cumulus parameterizations (Zhu et al. 2009; Wang and Schlesinger 1999; Maloney and Hartmann 2001). Most often, the limitations of climate models in simulating the MJO appear in the form of too fast propagation speed (Slingo et al. 1996) and/or relatively weak and weakly organized convective activity (Lin et al. 2006). Some
studies suggest the latter problem is related to the threshold conditions for the onset of convection. For example a higher threshold in entrainment rate results in a more organized and stronger MJO (Tokioka et al. 1988), and similarly Wang and Schlesinger (1999) show that inclusion of a relative humidity threshold for triggering deep convection improves the simulation. Yet a study by Fu and Wang (2009) shows that increasing stratiform heating by changing the entrainment rate has a similar effect as well.

As important as they are, limitations in convective parameterizations have to be considered as symptoms of our lack of understanding of the processes that provide the MJO with its unique propagation characteristics, such as its time and space scales as well as its amplitude (energy). Thus, development and modification of parameterizations for a realistic simulation of the MJO have to be founded on solid theoretical and observational grounds for them to be general enough to consistently address the above issues. To this end, a thorough, systematic and multiscale analysis of observed MJO moist thermodynamics is necessary. In the absence of detailed observations, analysis of these processes in high-resolution or cloud-resolving simulations that capture the important features of MJO without relying on cumulus parameterizations is a good starting point. The purpose of this study is to identify the roles of various processes associated with heat and moisture transport and microphysics of various types of clouds in determining the time scale and the energetics of MJO in a high-resolution regional simulation that explicitly resolves cloud-scale properties.

2. Model, experiment design, and evaluation

a. Model description and experiment setup

The model used in this study is the Advanced Research Weather Research and Forecasting version 3.1 (ARW-WRF3.1) (Skamarock et al. 2008). In the high-resolution WRF simulation (HIRES) the grid spacing is set at 4 km and the model is run without any cumulus parameterization. The Rapid Radiative Transfer Model (RRTM) (Mlawer et al. 1997) for longwave radiation, the Dudhia (1989) scheme for shortwave radiation, and the Yonsei University scheme (Hong et al. 2006) for PBL processes are used. The Noah land surface model (Mitchell et al. 2000) is used to represent land surface processes. The microphysics scheme used in these experiments is the WRF Single Moment Microphysics 3 class scheme (WSM3) (Hong et al. 2004).

Initial, lateral, and surface boundary conditions are derived from the National Centers for Environmental Prediction (NCEP) Global Forecast System Final Analysis.
the Maritime Continent (120°E) and the relatively strong signal to the immediate east in the middle of November.

Figure 3 shows the comparison of the model precipitation with the TRMM and GPCP products. The model generally has stronger and more widespread precipitation than both TRMM and GPCP. However, the locations and times of peak precipitation (e.g., near 100° and 150°E) are well captured. The fact that the OLR signal is not as strong as the satellite observations while the precipitation signal is stronger suggests that the model is likely overestimating shallow convection at the expense of deep convection. The MJO signal is also clearly observed as an eastward-propagating westerly low-level wind signal in an otherwise easterly trade wind environment. Figure 4 shows the zonal wind at 850 hPa from the model and two global reanalyses. Although the model captures...
the propagation of the westerly winds well, the signals are slightly stronger than both reanalyses. The differences are most apparent near mid-November (near 140°E) and the end of December (between 100° and 130°E) when the model westerly wind speed is almost double its reanalysis counterparts. Figure 5 shows the comparison of zonal winds at 200 hPa. The upper-tropospheric easterly is well captured in the model albeit slightly stronger, especially near 120°E both during the middle of November and the middle of January.

The fact that the high-resolution simulation captures all of the major features of the two MJO episodes without any cumulus parameterization suggests that most cloud processes involved in the MJO are simulated and warrants the analysis of these small-scale processes to gain insight into the dynamics and thermodynamics of MJO.

c. Composite analysis

The composite analysis performed on all subsequent analysis is similar to that described in Hagos et al. (2011).
As discussed in that study in more detail, the analysis transforms the two-dimensional time–longitude data into one-dimensional data at a hypothetical point that moves with the MJO, in this case, at 5 m s\(^{-1}\). In other words, the composite is created by averaging along the 5 m s\(^{-1}\) lines in Fig. 2. 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 a first-order understanding of its moist thermodynamics. Thus, the composite incorporates all longitudinal grid points in the domain into a single point that represents the longitudinal averages the MJO characteristics. A 10-day running mean is then applied to filter out high frequency variability. The composite is performed over the longitudinal range 57\(^\circ\)–130\(^\circ\)E. Figure 6 shows the composite evolution of precipitation, perturbations of potential temperature, moisture mixing ratio, and total diabatic heating. The precipitation maxima at day 30 and 71 are marked by the dashed lines. The precipitation peaks are accompanied by anomalous warming and moistening of the middle troposphere (Figs. 6b and 6c). Note that the potential temperature perturbations are baroclinic with midtropospheric anomalies out of phase with those of the lower troposphere (below 600 hPa). This peculiar tilt in the potential temperature anomalies has important implications for MJO thermodynamics that will be discussed in section 4. In the model, there is still a significant amount of precipitation during the inactive phase of the MJO. This precipitation is related to shallow convection, which is manifested in the form of a secondary diabatic heating peak near 700 hPa in addition to the upper-tropospheric heating associated with the active phase of the MJO (Fig. 6d).

3. Moisture budget analysis

a. Moistening and drying during the MJO life cycle

In this section, the moistening processes in the simulated MJO episodes are analyzed in the composite framework discussed above. The moisture budget equation in the model can be written as

\[
\frac{\partial q}{\partial t} = -\frac{\partial (qv_h)}{\partial p} - \nabla \cdot (qv_h) + Q_{2(\text{microphysics})} + Q_{2(\text{bl})},
\]

(1)

where \(v_h\) is the horizontal wind vector and \(\omega\) is the vertical pressure velocity. The terms on the rhs of (1) represent the rate of change of moisture mixing ratio \(q\) due to vertical and horizontal moisture flux gradients,
microphysics (condensation/evaporation $Q^2_{\text{microphysics}}$) and moisture fluxes by diffusion in the PBL ($Q^2_{\text{PBL}}$). These components of the moisture budget equation are shown in Fig. 7. During the inactive phase of MJO, shallow convection (explicitly simulated as vertical transport) moistens the lower troposphere (Fig. 7) while horizontal divergence of moisture tends to dry it. This is most apparent at 800 hPa about 20 days before the time of peak precipitation. The moisture sources for the upward moisture flux are both boundary layer (below 900 hPa) moisture convergence and surface fluxes, which have comparable contributions. As time progresses, the lower troposphere is moist enough that it becomes the moisture source as is shown by the overall deepening of the moisture source layer (Fig. 7a). About 10 days ahead of the MJO, the lower troposphere (near 800 hPa) is a source of the vertical moisture transport and the sink is in the midtroposphere (near 600 hPa) which subsequently rises even higher, marking the triggering of the MJO active phase. At about 10 days after the MJO peak precipitation, horizontal transport and condensation dry the lower troposphere and convection becomes shallow once again. There is little variation in the boundary layer moisture convergence and diffusion.

Another interesting aspect of moisture convergence in the model is the difference between the zonal and meridional components (Fig. 8). The moisture for shallow convection that is prevalent during both active and inactive phases of MJO is primarily supplied by the
meridional convergence. Thus, the apparent model bias in overestimating shallow convection is likely related to the meridional convergence. Much of the horizontal moisture convergence is contributed by the meridional component while the zonal component has a drying effect in the boundary layer, but it moistens the lower troposphere only during the active phase of MJO. This zonal moisture convergence is related to the convergence of westerly trade winds from the east and the MJO westerly winds to the immediate west of the convection (Fig. 4).

b. Time scales of moist processes and the MJO

To understand the roots of low frequency variability of the MJO, time scales of moistening and drying are considered in this section. In the previous section, we have seen that midtropospheric moisture content is controlled primarily by three processes: vertical moisture flux divergence (convective transport), horizontal moisture flux divergence (mixing), and condensational drying. When the four month mean is removed, Eq. (1) for the middle troposphere (600–300 hPa) can be approximated by

\[ \frac{\partial q'}{\partial t} \approx Q_{2(\text{vertical})}' + Q_{2(\text{horizontal})}' + Q_{2(\text{microphysics})}', \]  

(2)

where \( q' \) is the perturbation water vapor mixing ratio and \( Q_{2(\text{vertical})}' \), \( Q_{2(\text{horizontal})}' \) and \( Q_{2(\text{microphysics})}' \) are the perturbations of the first three rhs terms in (1). Correlation between each of the three rhs terms in (2) with \( q' \) is calculated for the duration of the life cycle of the two MJO episodes after the composite analysis. If there is strong correlation between each of the terms on the rhs in (2) and moisture mixing ratio perturbation, one can approximate these terms as linear functions of \( q' \) each with a characteristic time scale such that

\[ \frac{\partial q'}{\partial t} \approx \frac{q'}{\tau_{\text{vertical}}} + \frac{q'}{\tau_{\text{horizontal}}} + \frac{q'}{\tau_{\text{microphysics}}}, \]  

(3)

where \( \tau_{\text{vertical}} \), \( \tau_{\text{horizontal}} \) and \( \tau_{\text{microphysics}} \) are the characteristic time scales of each process. According to (3) the effective time scale for moistening is then

\[ \tau_{\text{effective}} = \frac{\tau_{\text{vertical}}\tau_{\text{horizontal}}\tau_{\text{microphysics}}}{\tau_{\text{vertical}}\tau_{\text{horizontal}}\tau_{\text{microphysics}}}. \]  

(4)

such that

\[ \frac{\partial q'}{\partial t} \approx \frac{q'}{\tau_{\text{effective}}}. \]  

(5)

Note that the moisture perturbation given by (4) is not oscillatory, but it provides an approximation for half of the period of oscillation (the recharge or discharge phases of MJO in this case) at levels where the moistening (drying) processes are strongly correlated to the moisture perturbations. Once the correlations between \( q' \) and each of the rhs terms in Eq. (2) are calculated, the slope of the best-fit line represents the time scale associated with each process.

Figure 9 shows the correlation coefficients and the associated time scales for each process and the effective time scale. Figure 9a shows that above 500 hPa there is a strong correlation between vertical transport of moisture (convection) and the moisture content. The associated time scale maximizes (about 2 days) at about 500 hPa and rapidly decreases with height. This is expected because latent heat release provides kinetic energy for the vertical transport and the heating depends on the environmental saturation mixing ratio, which decreases rapidly with temperature. The horizontal transport, which represents the exchange of moisture between the convection core and the environment through entrainment and detrainment, is also strongly correlated with the perturbation mixing ratio (Fig. 9b). Below about 500 hPa, moisture convergence and mixing ratio are positively correlated but above 500 hPa, the environment is drier than the convective core so moisture is detrained and dry air entrained and the correlation is negative. The associated time scale is about 4 days. Figure 9c shows that condensation \( Q_{2(\text{microphysics})}' \) is strongly correlated with moisture perturbation. The correlation is particularly strong above the 500-hPa level where the air is saturated or close to saturation, so changes in mixing ratio result in condensation/evaporation. The maximum associated time scale is about 20 h. Finally, the effective time scale is calculated using \( Q_{2(\text{effective})}' \), which is the sum of the 3 terms in the rhs of (1). The correlation is very weak because, in adding up the moistening/drying processes, the robust correlation signals for the large part cancel, while the uncorrelated components are being added up. In other words, the low frequency variability is relatively weak in comparison to the high frequency signals. Above 500 hPa, the effective time scale is between 15 and 25 days, which corresponds to an approximate period of 30–50 days. Note that the time scales involved in convective updraft, condensation, and mixing are not necessarily related to large-scale processes of that particular
characteristic time scales; instead, they should be interpreted as results of aggregation of small-scale processes. The above analysis shows that the low-frequency variability of the MJO is a result of small differences in the time scales of convective vertical moisture transport, mixing, and condensation. This might be an important factor for the sensitivity of MJO time scale (and phase speed) for different cumulus parameterizations.

4. Eddy available potential energy of MJO

a. The composition of latent heating in the model MJO

Simulation of the MJO episodes with a high-resolution regional model without cumulus parameterization enables the analysis of the roles of the explicitly resolved cloud types in the thermodynamics of the MJO. Before proceeding to the eddy available potential energy budget analysis similar to Hagos et al. (2011), the types of clouds in the model MJO are identified. In this study the cloud types in the model are identified by the profile of their associated latent heating. A bivariate PDF of maximum latent heating with respect to the level of maximum and minimum heating is performed. In other words, \( Q_{\text{ln}}[\text{max(lev), min(lev)}] \) is defined such that it is the sum of all the maximum latent heating from clouds with a maximum at \( \text{max(lev)} \) and minimum at \( \text{min(lev)} \). Figure 10 shows the PDF of \( Q_{\text{ln}}[\text{max(lev), min(lev)}] \). According to Fig. 10a the simulated clouds can be approximately categorized into four subsets (defined by the boxes). For example, subset 2 is dominated by cloud types

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**FIG. 9.** Correlations of the components of moistening/drying [rhs of (1)] with moisture perturbations and the associated time scales (days) calculated using linear best fit.
with peak heating at about 400 hPa and a minimum near about 600 hPa. The mean profile of each subset is shown in Fig. 10b.

To help interpret these profiles and define the cloud types, normalized, idealized heating profiles derived by Schumacher et al. (2007) using Doppler radar observation from TRMM field campaigns are included (Fig. 10c) for comparison. Despite some differences, such as the fact that cloud type 1 of the model has some cooling near the surface and the midtropospheric cooling associated with cloud type 4 is relatively weak compared to its heating, the heating profiles of the four cloud types obtained from the high-resolution simulation roughly match those derived from the TRMM field campaign. Thus, the four cloud types obtained from the high-resolution simulation are interpreted as deep convective, stratiform, shallow convective, and congestus clouds in accordance with their corresponding radar estimated profiles.

b. Eddy available potential energy budget analysis

To understand the roles of the various cloud processes in the MJO thermodynamics, eddy available potential energy budget analysis is performed. As discussed in more detail in Hagos et al. (2011), the rate of change of the amplitude of oscillation of potential temperature can be written as

\[
\frac{\partial \theta^*}{\partial t} = -\theta^* (\nabla \cdot (\mathbf{v} \theta^*)) + \frac{\partial}{\partial t} (\nabla \cdot \mathbf{Q})_{\text{latent}} + \frac{\partial}{\partial t} (\nabla \cdot \mathbf{Q})_{\text{radiation}} + \frac{\partial}{\partial t} (\nabla \cdot \mathbf{Q})_{\text{pbl}}.
\]  

The primes indicate the deviation of the terms from their time mean (in this case four months) value for every point in the domain. The term on the left-hand side represents the rate of change of the amplitude of potential temperature, while the first term on the rhs represents processes such as the buildup of potential energy because of the introduction of potential temperature difference between a parcel and its environment by transport of warm or cold air. The next three terms on the rhs represent instability (damping) due to selective heating of air parcel when it is warm (cold) and cooling it when it is cool (warm) by diabatic processes. Note that, in a homogenous environment where there is no systematic difference between time mean potential temperature and the environmental (space mean) potential temperature, the column-integrated form of (6) governs the generation and destruction of convective available potential energy (CAPE).

Figure 11 shows the contributions of the conversion to EKE, latent heating, and radiation to the eddy available potential energy (EAPE) budget after the composite analysis. The main balance is between conversion to/from EKE and latent heating. Above about 600 hPa, latent heating is a source of EAPE, while conversion to EKE is a sink. This is because, above the level of free convection, latent heating leads to warming and upward movement of the air parcel (conversion to EKE), which
leads to further condensation and warming. Thus, heating perturbations are in phase with temperature perturbations. At the lower troposphere, on the other hand, cold air continuously sinks to the lower troposphere and mechanically forces the moist boundary layer air upward, resulting in shallow convection and heating. In other words, shallow convective heating responds to (and damps) the instability created by mechanical forcing imposed on the moist boundary layer by the subsidence of cold air. Radiation perturbations are out of phase with temperature perturbation as one would expect from the fact that radiative cooling increases with temperature. Therefore, radiation also plays a small role in damping the MJO temperature fluctuations in the lower troposphere. The sign of the effect of radiation in intraseasonal variability is known to depend on the type and depth of the clouds involved. For example, anvil clouds are associated with positive forcing and shallow and midlevel clouds have negative forcing (Hartmann et al. 2001; Cess 1976). Given the apparent overestimation of shallow convection, its damping role is likely overestimated as well. The contribution of radiation forcing to condensational heating, according to Lin and Mapes (2004), is about 10%–15% and in our analysis its contribution to the EAP budget is about 10% positive (negative) in the upper (lower) troposphere. The contribution of sensible heat fluxes to EAP budget is negligible.

To identify the contributions of each cloud type to the EAP budget, the latent heating component [the second term on rhs of (5)] is separated into contributions from the four cloud types identified by the bivariate PDF. Figure 12 shows these contributions. Congestus clouds are sinks of EAP all the time. During the inactive phases of the MJO, their damping effect gradually deepens and reaches a peak, then the effect declines reaching near minimum when precipitation maximizes, after which it gradually deepens once again. The contours in Figs. 12a and 12b represent the level at which the contributions of the congestus and shallow clouds to EAP tendency are zero, respectively. Shallow convection (Fig. 12b) is also primarily a sink of EAP; its damping effect deepens and reaches a maximum value, afterward it generates EAP-like deep convection (Fig. 12c). Deep convection is a source of EAP for it leads to heating and further vertical moisture flux and condensation (Fig. 12c).

The stratiform heating perturbations are in phase with the potential temperature anomalies both at the lower and upper troposphere. This is related to the baroclinic properties of the potential temperature perturbations shown in Fig. 6b. Thus, near 600 hPa, where the generation of EAP associated with deep convection in the inactive period of convection is near zero (where the role of convection changes from damping to generating EAP), there is an extra source of EAP associated with stratiform variability that provides the EKE for further convection. In other words, as the gradual deepening of shallow convection reaches the level at which there is an extra source of EAP from stratiform heating variability, deep convection starts to appear. This is demonstrated by the change in sign of the EAP contribution by shallow...
convection, as it reaches its maximum depth about 15–20 days ahead of the peak precipitation.

5. Summary and discussion

In this study a high-resolution regional model simulation without a cumulus parameterization is used to simulate the two MJO episodes observed during boreal winter 2007. Moisture and eddy available potential energy budget analyses are then used to identify the processes contributing to the life cycle of the MJO. During the inactive phase of MJO, shallow and congestus clouds moisten the lower troposphere while horizontal mixing tends to dry it (Fig. 7a). The main sources of moisture at the boundary layer are meridional moisture convergence and PBL diffusion (evaporation, Figs. 7c and 8b). Zonal moisture convergence is slightly elevated, weak and positive only during the active phase of the MJO (Fig. 8a). As the lower troposphere is moistened, it becomes a source of moisture for the subsequent deep convection (Figs. 7a,b). During this period, the horizontal convergence also deepens. As the MJO active phase ends, the lower troposphere dries out primarily by condensation and horizontal divergence, which dominates over the moisture supply by vertical transport.

The longer time scale involved in the moistening of the midtroposphere during recharging (discharging) of the MJO in the model is related to the small differences in the time scales of the vertical transport of moisture, horizontal mixing, and condensational drying. At the midtroposphere, for example, condensational drying has a time scale of about 20 h, while horizontal mixing has time scale of ~4 days and vertical transport has time scale of ~2 days (Fig. 9). As a result, the effective time scale of moistening is ~25 days according to (4), which approximates half of the period of the simulated MJO.

In the high-resolution model simulation four distinct types of clouds are identified using a bivariate PDF of their latent heating profile (Fig. 10). These are congestus, shallow convective, deep convective, and stratiform clouds. They play differing roles in the thermodynamics of the MJO. Eddy available potential energy budget analysis shows that the heating from shallow and congestus clouds is negatively correlated with the potential temperature perturbations, which suggests that it damps the EAPE generated by convergence of eddy available potential energy (Figs. 12a,b). The deep convection, on the other hand, is marked by a positive correlation between latent heating perturbations and temperature perturbations, and hence is a source of eddy available potential energy.
energy. In other words, while shallow convective heating acts to stabilize the lower troposphere, deep convection destabilizes the midtroposphere by converting condensational heating to eddy kinetic energy (Fig. 12c).

Both in the lower and upper troposphere, stratiform heating is an important source of EAPE and provides additional EAPE for the transition into deep convection once the shallow convection is deep enough that the level of shallow damping reaches the level of stratiform instability near 600 hPa (Figs. 12b and 12d) about 15–20 days before the date of peak precipitation (Figs. 12b,d). Radiation has a damping effect on the midtroposphere EAPE budget throughout the MJO life cycle.

The above summarized results can also be evaluated in the context of previous observational and modeling studies. The moistening by shallow convection during inactive phases of the MJO and the immediate drying following the active phase are observed in various studies (Benedict and Randall 2007). The important role of the tropospheric moisture content has recently been recognized in observations (Bretherton et al. 2004) and is supported by the relative successes of convective parameterizations that relate tropospheric relative humidity with precipitation (Raymond 2001; Wang and Schlesinger 1999). In this study the relationship between precipitation (condensational drying) and tropospheric moisture content is quantified and found to be very strong (Fig. 9c) and, along with the relationship of vertical transport of moisture and horizontal mixing with the environment, it sets the time scale of the low frequency variability.

The results of the EAPE budget analysis are fairly consistent with the results from our recent study (Hagos et al. 2011) using a low-resolution simulation (36 km) with constrained moistening to produce a realistic MJO. The roles of shallow, deep, and stratiform heating in the MJO thermodynamics obtained from the previous study using a low-resolution constrained model simulation are fairly well reproduced in this high-resolution simulation without cumulus parameterization. The simulated characteristics of the covariance between heating and temperature during the MJO life cycle are also found in reanalysis products (Yanai et al. 2000; Benedict and Randall 2007). Furthermore, the results of our study support the moisture–convection feedback hypothesis, forwarded by Grabowski and Moncrieff (2004), in that downdrafts of cold air associated with deep convection deepen the moist layer, leading to further deep convection as demonstrated by the conversion of EAPE to EKE in the middle and upper troposphere and the conversion of EKE to EAPE at the lower troposphere and the associated forcing of shallow convection followed by deep convection (Fig. 12).

The results of this study suggest that efforts at developing convective parameterizations to accurately simulate the time scale and amplitude of the MJO will likely benefit from improved representations of (i) the relationships between midtropospheric moisture content, convective transport and mixing, as well as their associated time scales; (ii) shallow convection and its relationship to convective downdrafts; and (iii) stratiform heating and its contribution to wave energy.

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