Effects of Organized Convection Parameterization on the MJO and Precipitation in E3SMv1. Part I: Mesoscale Heating

C.-C. Chen1, J. H. Richter1, C. Liu1, M. W. Moncrieff1, Q. Tang1, W. Lin1, S. Xie1, and P. J. Rasch1

1National Center for Atmospheric Research, Climate and Global Dynamics Division, Boulder, CO, USA, 2National Center for Atmospheric Research, Research Applications Laboratory, Boulder, CO, USA, 3Lawrence Livermore National Laboratory, Livermore, CA, USA, 4Brookhaven National Laboratory, Upton, NY, USA, 5Pacific Northwest National Laboratory, Richland, WA, USA

Abstract Mesoscale organization of convection is typically not represented in global circulation models, and hence its influence on the global circulation is not accounted for. The heating component of a parameterization that represents the dynamical and physical effects of circulations associated with organized convection, referred to as the multiscale coherent structure parameterization (MCSP), is implemented in the Energy Exascale Earth System Model version 1 (E3SMv1). Numerical simulations are conducted to assess its impact on the simulated climate. Besides E3SMv1 simulations, we performed high-resolution (2 km) simulations using the Weather Research and Forecasting (WRF) Model to determine the temperature tendencies induced by mesoscale convective systems embedded in deep convection. We tuned the free parameters of the MCSP based on the WRF simulations. MCSP heating enhances Kelvin wave spectra in E3SMv1, improves the representation of the Madden-Julian Oscillation, and reduces precipitation biases over the tropical Pacific.

Plain Language Summary Most general circulation models do not account for important circulation changes associated with organized convection. We implement a new parameterization of heating associated with mesoscale circulations in Energy Exascale Earth System Model version 1 (E3SMv1). The parameterization improves the representation of Madden-Julian Oscillation simulated by E3SMv1 and also reduces the precipitation biases in the tropical West Pacific.

1. Introduction

The transport and mixing of heat and momentum throughout the atmosphere largely control the global circulation, and hence moisture and precipitation patterns. However, several transport processes occur on scales much smaller than a global circulation model (GCM) grid box, and therefore have to be parameterized. Improvements in the representation of subgrid heat and momentum transport can lead to significant model improvements in the representation of wind stresses, moisture and precipitation patterns, and organized modes of variability (Richter & Rasch, 2008). In particular, convection is a large source of heat and momentum transport in the atmosphere, both on scales of individual convective plumes, as well as on scales of the order of 10–1,000 km (mesoscales) (Houze, 2004; Moncrieff, 1995).

The importance of convective organization on the global circulation has been recognized for more than three decades but parameterizations of the attendant processes are missing from GCMs. Contemporary convective parameterizations commonly use a convective plume model (or a spectrum of plumes) (Arakawa & Schubert, 1974; Chen & Mapes, 2018; D’Andrea et al., 2014; Kain & Fritsch et al., 1990). This is appropriate for unorganized convection. However, the assumption of a gap between the cumulus scale and the large-scale motion that underpins contemporary convective parameterizations fails to recognize mesoscale dynamics manifested in squall lines, mesoscale convective systems (MCS), mesoscale convective complexes (MCC), and multi-scale cloud systems associated with the Madden-Julian Oscillation (MJO). Over 50% of convective precipitation in the tropics is provided by MCS defined as heavily precipitating closely coupled cumulus ensembles embedded in the more moderately precipitating stratiform regions of these systems (Nesbitt et al., 2006; Tao & Moncrieff, 2009).
Mesoscale convective organization significantly modulates the life-cycle of moist convection, the transport of heat, moisture, momentum and chemical constituents (Houze, 2004, 2014). Organized convection is abundant in environments featuring lower-tropospheric vertical wind shear (Anber et al., 2014; Robe & Emanuel, 2001) and is typically associated with counter-gradient momentum transport (Moncrieff, 1992). Moncrieff and Liu (2006) designed a hybrid “predictor-corrector” framework to parameterize mesoscale convective systems and tested this approach in the Weather Research and Forecasting (WRF) model with a 60 km grid over the United States continent in summertime meteorological conditions. Specifically, a cumulus parameterization scheme gives the convective heating profile (predictor) which is then adjusted by adding sine-like upper-tropospheric warming and low-tropospheric cooling (i.e., second baroclinic normal mode) corrector that emulates the mesoscale heating profile associated with MCSs. Moncrieff et al. (2017) (hereafter MLB17) recently implemented a similar approach in the Community Atmosphere Model (CAM) in the form of multiscale coherent structure parameterization (MCSP) where organized convection is treated as coherent structures in a turbulent environment. MCSP is approximated by a slantwise layer overturning dynamical model that exchanges tropospheric layers via convectively generated mesoscale circulations which are significantly controlled by environment vertical shear (Figure 1). Because slantwise layer overturning is not represented by existing convective parameterizations in a GCM, it is appropriate to add the “missing” mesoscale tendencies to traditional convective parameterizations. For the first time, differences between GCM simulations with and without MCSP directly measure the large-scale effects of convective organization.

The implementation of heating and momentum components of MCSP was shown in MLB17 to improve precipitation biases in CAM version 5.5 (CAM5.5), for example, in the tropical west Pacific, equatorial Africa, and the Inter Tropical Convergence Zone (ITCZ). MCSP also increased the Kelvin wave amplitude and extended the MJO signal from zonal wavenumber 1 to a more realistic wavenumber 1–5 range (MLB17).

The Energy Exascale Earth System Model, version 1 (E3SMv1) is a new Earth system model designed to meet the science needs of the nation and the mission of the Department of Energy (DOE). E3SM currently does not represent heat or momentum transport associated with mesoscale convective organization, so the MCSP introduces missing physical processes which can potentially reduce E3SM’s biases, in particular, over the ITCZ, south Pacific convergence zone, and the maritime continent. Parameterization of mesoscale transport is needed for the next generations of E3SM.

This article is structured as follows: Section 2 describes the mesoscale heating component of the MCSP, Section 3 describes the WRF simulations that are used to set tunable parameters in MCSP, Section 4 describes changes to the simulation of convection and tropical variability in E3SMv1 as a result of MCSP heating, and Section 5 includes summary and conclusions.

2. Mesoscale Heating Parameterization

The heating component of MCSP consists of a temperature tendency of multiscale convective systems which is added to the existing deep convection parameterization (Zhang & McFarlane, 1995) to account for induced impacts. Following Moncrieff and Liu (2006) and Moncrieff et al. (2017), mesoscale heating associated with unresolved atmospheric circulations in GCMs due to inadequately coarse horizontal resolution
is represented as a second baroclinic normal mode with amplitude proportional to the vertically averaged convective heating provided by the convective parameterization. While a realistic vertical distribution of mesoscale heating is considerably more complex (Moncrieff, 1992), this simplification captures its dominant physical and dynamical features. Figure 2b illustrates the vertical profile of mesoscale heating, representing heating \((H)\) in the trailing stratiform region and evaporative cooling \((C)\) beneath as shown in Figure 2a. In the Moncrieff and Liu (2006) prototype, the simplest possible vertical profile of mesoscale heating \((Q_m)\) in Figure 2, is given by:

\[
Q_m(p,t) = -\alpha Q_c(t) \sin \left[ 2\pi \frac{p - pt}{ps - pt} \right],
\]

\[
p^* = \frac{ps + pt}{2},
\]

\[
Q_c(t) = \frac{1}{ps - pt} \int_{ps}^{pt} H(p,t) dp.
\]

where \(\alpha\) is the ratio between stratiform and convective heating which is estimated based on Weather and Research Forecasting model (hereafter WRF) simulations as described in Section 3; \(ps\) and \(pt\) are the bottom and top of the convection respectively; \(p^*\) is the location where the heating profile crosses the zero line; \(H\) is the heating rate from the deep convection scheme; \(Q_c\) is the vertically averaged \(H\).

### 3. WRF Simulations

We utilize version 4.0.3 of WRF (Skamarock et al., 2008) to conduct convection resolving simulations that guide the free parameters in the mesoscale heating parameterization. WRF is a model with a terrain-following mass coordinate, suitable for use in a broad spectrum of applications across scales ranging from meters to thousands of kilometers. The computational domain is 4,600 × 1,700 km at 2 km grid spacing with 65 stretched vertical levels and model top at 25 hPa. The subgrid parameterizations employed in WRF are the single-moment 6-class microphysics (Hong & Lim, 2006), the YSU planetary boundary layer parameterization (Hong et al., 2006), the Noah-MP land surface model (Niu et al., 2011), and the Rapid Radiative Transfer Model for longwave and shortwave atmospheric radiative fluxes (Iacono et al., 2008). Spectral nudging applied to wind and water vapor fields above the planetary boundary layer preserve characteristics of the large-scale circulation, while allowing explicit sub-synoptic, mesoscale, and convective processes. Two multi-day episodes of active convection associated with an MJO event (MLB17) were selected for simulation, corresponding to a 9-days convective period in the Indian Ocean (Figure 3a) and a 14-days convective period in the western Pacific (Figure 4a), respectively. The hourly, 31-km-resolution ERA5 reanalysis data provided initial and boundary conditions. Figures 3b and 4b show that the observed convective systems in
both the Indian Ocean and the western Pacific are well captured by the WRF convection resolving simulations which provide information to guide selection of the parameter $\alpha$ in MCSP.

An MCS typically has two distinct parts, namely a convective and stratiform region. The convective region consists of intense, deep convective cells in association with strong precipitation cores and updrafts while the stratiform region, typically missing in GCM simulations, has a relatively uniform texture with lighter precipitation and lower reflectivity, mainly produced by aging convective cells (often from the convective
area) or by broader mesoscale ascent. Instantaneous condensational heating rate and reflectivity from WRF simulations are used with an algorithm of convective-stratiform separation in radar data analysis for convective-stratiform heating partitioning. As detailed in Steiner et al. (1995) and Houze (2014), the procedure is to identify convective area, and treat the remaining part as the stratiform region. The philosophy for determining the convective region is based on the intensity and peakness of the reflectivity field. Specifically, any grid point either with at least a 40 dBZ signature or exceeding the surrounding background reflectivity by a specified factor is considered to be the convective center. Additionally, grid points surrounding a convective center, as described in Steiner et al. (1995), are considered a convective area. A caveat of this approach is its inability to identify a shallow convection. In this analysis, the reflectivity is computed based on a diagnostic widely used in the WRF community, and the heating rate is explicitly computed by the WRF cloud microphysics scheme.

The heating profiles in Figure 5 are the spatial and temporal averaged heating rate over convective or stratiform grid points during each simulation period. Therefore, the derived ratio corresponds to the mean contribution of all MCSs within the entire domain. The detailed procedure for the ratio computation is as follows. The three-dimensional instantaneous condensational heating rate and reflectivity field are saved hourly during WRF simulations. Note that the reflectivity is diagnosed from the rainwater, cloud water, cloud ice, snow, and graupel hydrometeor fields via a standard WRF output variable, whereas the condensational heating is obtained directly from the cloud microphysics parameterization. The aforementioned convective-stratiform partitioning algorithm is applied to the hourly model reflectivity field to identify convective grid points and stratiform points. The profiles in Figure 5 are obtained by summing and then averaging over all convective/stratiform points for each multi-day simulation. The vertical profile of convective/stratiform heating is obtained and the vertically averaged convective heating rate $Q_c$ is computed using Equation 3.

Figure 5. Heating profiles for deep convection (red), stratiform and shallow convection (blue), and their sum (black). The left and right panels are corresponding to the convective episode (April 5–13, 2009) in the Indian Ocean and the convective episode (April 12–25, 2009) in the western Pacific, respectively.
The vertically averaged stratiform heating rate $Q_m$ between $p^*$ and $p_t$ is calculated using the profile in Figure 5. According to Equation 1, $\alpha = \frac{Q_m}{Q_c}$, and as such is estimated in terms of $Q_c$ and $Q_m$.

The ratio of stratiform to convective heating is uniform for all MCSs in order to simplify a complex process. In reality, the ratio will depend upon the MCS lifecycle, the category of MCS, and environmental conditions such as vertical wind shear. A formulation of the ratio parameter as a function of those factors is a subject for future work.

The convective-stratiform heating partitioning is thereby obtained. As illustrated in Figure 5, the stratiform heating is approximately 0.3–0.5 that of the deep convective heating in both Indian Ocean and the western Pacific. Therefore, $\alpha$ in Equation 1 is assumed to be 0.3 and 0.5 in this study.

Since a favorable condition for the development of MCSs includes substantial vertical wind shear, we add a wind shear threshold in order to determine where MCSP is triggered. To examine the sensitivity of the simulated climate on the wind shear threshold, we performed simulations with the zonal wind shear between surface and 600 hPa set to exceed 0, 3, 5, and 7.5 m/s as the trigger. This condition ensures that MCSP does not apply to conditions where the possibility of MCS formation is low. This shear-aware procedure updates the formulation of MLB17 where the wind shear trigger is set to 0 m/s, that is, MCSP is activated whenever deep convection occurs. The level of 600 hPa was chosen over the level of 700 hPa for the implementation due to the much stronger wind shear between surface and 600 hPa simulated by the baseline model of E3SMv1.

4. Climate Model Simulations

4.1. Model Description

The climate model employed in this study is the E3SMv1 (Golaz et al., 2019), which originated from the CESM1. There are significant developments in individual components of the model compared to CESM1: (a) new options for representing soil hydrology and biogeochemistry in the land model (ELM) based on the Community Land Model version 4.5 (CLM4.5); (b) new ocean and sea-ice components based on the Model for Prediction Across Scales (MPAS) framework; and (c) a new river model (i.e., Model for Scale Adaptive River Transport [MOSART]) that has not been previously used in a coupled Earth System Model.

Several updates to the atmospheric component of E3SMv1, EAMv1, follow the Community Atmosphere Model version 5.3. EAMv1 employs a unified treatment of planetary boundary layer turbulence, shallow convective, and cloud macrophysics with a third-order turbulence closure parameterization (Cloud Layers Unified by Binormals [CLUBB]) (Golaz et al., 2002; Larson, 2017; Larson & Golaz, 2005) which eliminates unrealistic separation of these physical processes. Turbulence, clouds, and convective processes are handled by the Zhang and McFarlane (ZM) deep convection scheme (Zhang & McFarlane, 1995) paired with CLUBB and an updated cloud microphysical scheme, version 2 of Morrison and Gettelman (MG2) (Gettelman & Morrison, 2015; Gettelman et al., 2015; Morrison & Gettelman, 2008). An update to the MG2 is a Classical Nucleation Theory based on ice nucleation parameterization for heterogeneous ice formation for mixed phase clouds (Wang et al., 2014). Rasch et al. (2019) provide an overview of EAMv1, while Xie et al. (2018), Qian et al. (2018), and Zhang et al. (2019) show simulated cloud and convective characteristics and the rationale for model tuning, and Tang et al. (2019) documents its regionally refined capability for developing high-resolution physics parameterizations.

The horizontal resolution of the simulations is ~100 km and there are 72 vertical levels with the model top at 60 km. The configuration of the vertical grid is shown in Figure 1 of Xie et al. (2018) with vertical spacing in the upper troposphere and lower stratosphere ~600 m.

4.2. Simulations

We conducted six simulations with the stand alone atmospheric component of E3SM (EAMv1) with prescribed observed sea-surface temperatures and sea-ice concentrations (“AMIP” simulations) and four simulations with the coupled version of E3SMv1 (“Coupled” simulations). All simulations are 30 years long.
beginning November 1980 and ending December 2009. In these simulations we vary the parameter $\alpha$ as well as the wind shear trigger (tunable parameters). The simulations are summarized in Table 1.

### 4.3. Results

We document the most significant impacts in EAMv1/E3SMv1 simulations due to MCSP heating with the focus on convection, precipitation and tropical variability.

#### 4.3.1. Deep Convection

The heating rate of MCSP and its impact on the deep convection scheme are illustrated in Figure 6. As shown in Figure 2b, when deep convection occurs, the circulation of MCSs induces lower-tropospheric cooling and upper-tropospheric warming. The amplitude of MCSP heating alone is illustrated in Figure 6a for $\alpha = 0.3$ and $\alpha = 0.5$. With $\alpha = 0.3$ ($\alpha = 0.5$) the lower tropospheric cooling peaks at 0.7 K/day (1.2 K/day), whereas the mid-tropospheric heating peaks at 0.25 K/day (0.5 K/day). MCSP produces temperature tendencies in the EAMv1 simulations that reflect the impact of MCSs. However, it is important to note that the heating tendencies illustrated in Figure 6a are averaged per occurrence of deep convection. Thus, the asymmetry in the magnitude of cooling and heating

---

**Table 1**

| Simulation type | $\alpha$ | Wind shear trigger (m/s) |
|-----------------|---------|--------------------------|
| EAMv1           | AMIP    | 0                        |
| EAMv1_a300      | AMIP    | 0.3                      |
| EAMv1_a500      | AMIP    | 0.5                      |
| EAMv1_a530      | AMIP    | 0.5                      |
| EAMv1_a550      | AMIP    | 0.5                      |
| EAMv1_a575      | AMIP    | 0.5                      |
| E3SMv1          | Coupled | 0                        |
| E3SMv1_a300     | Coupled | 0.3                      |
| E3SMv1_a500     | Coupled | 0.5                      |
| E3SMv1_a530     | Coupled | 0.5                      |

*Note. All simulations are 30 years long from 1980 to 2009.*

---

**Figure 6.** One-month (1980/11) averaged heating rate profile due to (a) deep convection scheme with Multiscale Coherent Structure Parameterization (MCSP) parameterization, and (b) MCSP parameterization by EAMv1 simulations.
tendencies in the vertical profile implies more frequent occurrence of shallower convection which results in stronger averaged cooling in the lower troposphere. Also notice that MCSP, by design, has a zero vertically integrated temperature (dry static energy) tendency as a correction term is added to the column to ensure the parameterization conserve total energy, and thus shallower albeit more intense cooling in the lower troposphere and deeper and weaker warming in the upper troposphere.

The heating rate from the ZM (Zhang & McFarlane, 1995) deep convection scheme, with and without MCSP included, is shown in Figure 6b. With the addition of MCSP heating, the ZM heating rate is weaker in the lower troposphere (below 800 hPa) because MCSP cools that region. However, the heating rate in the upper troposphere (around 600 hPa) is also weaker than the control simulation where MCSP adds heating (as will be explained below). Moreover, with a larger $\alpha$ the heating rate is weaker in the upper troposphere which is somewhat surprising since MCSP provides a heating tendency.

The circulation associated with MCSs stabilizes the troposphere due to a cooling effect in the lower troposphere and a heating effect aloft. Thus, when deep convection occurs, MCSP reduces the convective available potential energy (CAPE) by enhancing the tropospheric stability via weaker, less persistent, and less frequent ZM convection. Hence, the MCSP heating has a stabilizing effect on deep convection and leads to reduced ZM heating. With a larger $\alpha$, the weakening of the ZM scheme is more pronounced.

### 4.3.2. Tropical Variability

One well documented bias in E3SMv1 is the representation of tropical variability. In particular, the Kelvin waves are much weaker than observations (Rasch et al., 2019; Richter et al., 2019). As illustrated in Figure 7b, the baseline model of EAMv1 indeed simulates much weaker Kelvin waves than TRMM observations (Figure 7a). Thus, a focus of the convection scheme development in E3SM has been to generate more realistic Kelvin wave spectra.
Importantly, the heating component of MCSP enhances the Kelvin wave spectra in all simulations. The most pronounced increase in Kelvin wave activity occurs when there is no wind shear trigger threshold in MCSP so it gets activated whenever deep convection occurs. With a wind shear trigger, MCSP heating is activated less frequently and therefore has a weaker impact on the simulation. The results from coupled simulations are quite similar and thus not shown. In a recent study by Bengtsson et al. (2021), it is found that a parameterization of organized tropical convection also improves spectral characteristics of Kelvin waves.

4.3.3. MJO
The MJO is a dominant mode of subseasonal variability in the tropics. The key signature of MJO is eastward propagation of tropical convection originating from the Indian Ocean to the west Pacific during the boreal winter (Madden & Julian, 1971, 1972). The MJO is often not well represented in climate models (Ahn et al., 2017, 2020). In observations, the cross-lag correlation of precipitation shows pronounced eastward propagation from 60°E to 160°E and meridional propagation from the equator to 20°S and 20°N (Figure 8a).

The cross-lag correlation, of precipitation and zonal wind at 850 hPa, indicates that the baseline EAMv1 produces substantial westward propagation (top panel of Figure 8b) instead of the dominant eastward

Figure 8. Cross-lag correlation of precipitation and zonal wind at 850 hPa during the boreal winter by (a) ERA-Interim, and simulations by EAMv1: (b) baseline model, (c) Multiscale Coherent Structure Parameterization (MCSP) with α = 0.3, (d) MCSP with α = 0.5, (e) MCSP α = 0.5 and a wind shear threshold of 3 m/s, (f) MCSP α = 0.5 and a wind shear threshold of 5 m/s, (g) MCSP α = 0.5 and a wind shear threshold of 7.5 m/s. The cross-lag correlation is computed for bandpass filtered precipitation and zonal wind at 850 hPa within 10°N, 10°S, 80°E, 100°E. The latitude range for the top panels is 10°S and 10°N, and the longitude range for the bottom panels is 80°E and 100°E.
propagation (top panel of Figure 8a) in the Indian Ocean basin according to the ERA-Interim reanalysis. This indicates that there is a substantial westward propagation component of precipitation in the baseline model whereas meridionally it is mainly stationary near the equator with nearly symmetrical meridional propagation of precipitation in the 10°S/N–20°S/N region (bottom panel of Figure 8a). However, EAMv1 simulates much less meridional precipitation propagation in the Indian Ocean basin (bottom panel of Figure 8b).

With MCSP heating, EAMv1 simulates less westward propagation and more meridional precipitation propagation (Figures 8c–8g). The setup with $\alpha = 0.5$ and wind shear threshold = 3 m/s (Figure 8e) shows significant improvement over the baseline model since the simulation shows the clearest eastward precipitation propagation in the Indian Ocean basin.

When run in the fully coupled configuration (E3SMv1), the model shows significant improvement over EAMv1 in the simulation of the MJO. The fully coupled model produces clear eastward propagation of precipitation in the Indian Ocean basin (Figure 9a) and removes most of the westward propagation that was seen in EAMv1 (Figure 8b). However, meridional precipitation propagation in E3SMv1 remains weak compared to observations. With inclusion of MCSP heating, the eastward precipitation propagation is faster than in the baseline E3SMv1 (Figure 9b) when $\alpha = 0.5$ (Figures 9c and 9d). MCSP also extends meridional precipitation propagation farther from the equator.

The spectral characteristics of the MJO are considered in Figures 10 and 11. Observations (NOAA Interpolated Outgoing Longwave Radiation, Liebmann & Smith, 1996, Figure 10a) show eastward propagation in outgoing longwave radiation (OLR) in the tropics (10°S–10°N) with a period between 30 and 90 days and wave numbers between 1 and 3 during the boreal winter. The baseline EAMv1 has little power in eastward propagating modes with periods between 30 and 90 days (Figure 10b) despite substantial power for eastward propagating modes with periods greater than 90 days. In addition, there is slow (>90 days) westward propagating modes in the baseline model.

When MCSP heating is included in EAMv1, slow (>90 days) eastward propagation of OLR is further enhanced (Figures 10c–10g). More power within eastward propagating modes in the window of 30–90 days is evident in Figures 10c–10e, implying improved MJO simulation. It is worth noting that Figure 10e (model configuration $\alpha = 0.5$, and shear threshold 3 m/s) shows much stronger westward OLR propagation which indicates degradation from the baseline model simulation.
E3SMv1 indicates an improved MJO compared to EAMv1 in terms of increased power within the window of 30–90 days in eastward OLR propagation (Figures 11a), although the eastward propagation is still slow compared to observation (Figure 10a). With MCSP heating, the fully coupled model simulations (Figures 11b–11d) show further improvements, that is, increased power in the 30–90 days window in eastward OLR propagation while not enhancing westward propagation. The MJO spectra is too strong for $\alpha = 0.5$ without a wind shear threshold (Figures 11c) as compared to observations, but is reduced by a wind shear trigger threshold of 3 m/s (Figures 11d).
To quantify the impact of MCSP heating on the simulations, we use the sum of the OLR spectra within 30–90 days (eastward) and wave numbers 1–5 as the MJO spectra. For EAMv1 simulations, all configurations with MCSP increase the MJO spectra compared to the baseline model (percentage difference from observation is given in the caption of Figure 10). The MJO spectra under the configuration of $\alpha = 0.5$ and shear threshold 3 m/s differs merely 2% from observation, down from $-17.6\%$ of the baseline EAMv1. The baseline model of E3SMv1 features a much better representation of MJO since the magnitude of the percentage difference of the MJO spectra from observation is reduced to 6.7% (Figure 11) from 17.6% from baseline EAMv1 (Figure 10). The implementation of MCSP in E3SMv1 also increases the MJO spectra as found in EAMv1 and the magnitude of the percentage difference exceeds 6.7%. This could be interpreted as a degradation from the baseline E3SMv1, but after careful examination, the baseline model of both EAMv1 and E3SMv1 shows little power between 30 and 50 days whereas MCSP enhances spectra within this range, that is, indicating an improvement.

Figure 12 illustrates the life-cycle composite of MJO, which is constructed using the two leading principal components of OLR (PC1 and PC2 computed for observations and each simulation) as in Wheeler and Hendon (2004) and only days when $PC_1^2 + PC_2^2$ exceeds 1 are selected for composition. The observations indicate that convection forms in the Indian Ocean basin during the early phases of the life cycle and propagates eastward to the tropical West Pacific. The baseline model EAMv1 simulates much weaker OLR (and hence MJO) (Figure 12b). With MCSP heating, the intensity of OLR is enhanced slightly but still much weaker than observation. Additionally, EAMv1, with or without MCSP, lacks organized convection within the Indian Ocean basin during MJO phases 2 and 3, compared to observations (Figure 12a).

Figure 13a shows E3SMv1 simulates stronger OLR, but the lack of organized convection in the Indian Ocean basin during phases 2 and 3 persists. With MCSP heating (Figures 13b–13d), the intensity of OLR is stronger, mainly in the later phases (5–8) but the model also fails to produce strong organized convection in the Indian Ocean basin during phases 2 and 3.

Both EAMv1 and E3SMv1 generate little propagation in OLR during later phases of the MJO life-composite plots (phases 6–7 in Figure 12b; phases 5–6 Figure 13a) but observations show clear eastward propagation (Figure 12a). With MCSP heating, the progression of eastward propagation continues during these later phases of the composite. Thus, MCSP exhibits an important improvement over the baseline model.
As revealed by various diagnostics for MJO presented in this section, MCSP heating has a positive impact on the representation of MJO. There are two tunable parameters in MCSP, \( \alpha \) and the vertical wind shear trigger. When \( \alpha \) is larger, MCSP adds a greater temperature tendency and thus is conceivable that MCSP will have a stronger impact. The vertical wind shear trigger determines when MCSP is activated. Consequently, if the vertical wind shear trigger is larger, the criteria are met less frequently and thus less likely MCSP is activated. Therefore, MCSP heating has a smaller impact on the simulation when the vertical wind shear trigger is larger.

4.3.4. Precipitation Biases

Since MCSP is implemented within the deep convection scheme, it is anticipated to alter convective precipitation especially in the lower latitudes where active deep convection is extensive. Indeed, MCSP mainly modifies precipitation near the tropics and sub-tropics in both EAMv1 (Figure 14) and E3SMv1 (Figure 15). In both model configurations, MCSP heating robustly reduces precipitation around the tropical east Pacific, Colombia, and Ecuador (see Figures 14 and 15). The reduction of precipitation in these regions is more
pronounced when the wind shear threshold is smaller, meaning that MCSP gets triggered more frequently. Also, these regions are where the baseline EAMv1/E3SMv1 produces positive biases in precipitation so MCSP provides improvement. Nevertheless, the reduction is weaker during December-January-February (DJF), a feature also evident in CAM5.5 simulations, i.e. Figure 14b of MLB17. The annual average of the EAMv1 simulations with MCSP heating shares several other features found in CAM5.5 (e.g., Figure 14b in MLB17), such as enhancement in precipitation in the tropical west Pacific, South China Sea, and Indian Ocean basin, and reduction in tropical Africa (see Figure 14d). Such features are suppressed with a higher wind shear trigger (Figures 14b and 14c). It is also worth noting that the enhancement in precipitation in South China Sea and the Indian Ocean is mainly attributed to June-July-August (JJA, Figures 14k and 14l) since in DJF (see Figures 14g and 14h), the change in precipitation in these regions due to MCSP is negative. However, these features represent a degradation to EAMv1 because the baseline model already simulates positive biases in precipitation in these regions (Figure 14a).

Another region showing significant change in precipitation due to MCSP heating is in south and southeast Asia, and over India and the Indian Ocean basin. In DJF, a reduction in precipitation due to MCSP occurs (Figures 14f–14h and 15f–15h). EAMv1 and E3SMv1 in the baseline model configuration both produce positive biases in these regions and thus such changes in DJF bring the simulated climate closer to observation. When MCSP is incorporated in the fully coupled model, E3SMv1, the change in precipitation in the tropical west Pacific differs from those in EAMv1. For the annual average, a reduction in precipitation occurs (Figures 15b–15d), which improves the model biases. However, E3SMv1 with MCSP simulates more precipitation in these regions in JJA (Figures 15j–15l) and thus degrades the model. The reduction in annual precipitation is attributed to more pronounced lower precipitation simulated in DJF (Figures 15f–15h). In terms of annual precipitation, MCSP heating does not induce a coherent feature in the tropical west Pacific between EAMv1 and E3SMv1. However, the parameterization induces the same features in seasonal averages: MCSP heating enhances precipitation in JJA but reduces precipitation in DJF in these regions.

4.3.5. Impact on Temperature and Humidity Distribution

The impact of MCSP heating on the temperature and humidity field is illustrated in Figure 16. Since MCSP cools the lower troposphere, it is not surprising that the temperature simulated with MCSP in the lower
troposphere is reduced (Figures 16a, 16d, and 16g). While AMIP simulations with MCSP heating produce cooling only in the lower troposphere, the coupled model with MCSP generates much stronger cooling which extends throughout the troposphere. It is important to note that in both model configurations, MCSP heating produces a layer with enhanced stability between 700 and 900 hPa which can suppress convection.

The simulations with MCSP under the AMIP configuration show higher specific humidity in the lower troposphere (Figures 16b, 16e, and 16h), attributed to less persistent convection due to enhanced stability between 700 and 900 hPa. The coupled simulations with MCSP heating, however, show reduced specific humidity throughout the troposphere. The pattern of the specific humidity difference profile is quite similar between the two model configurations (AMIP and coupled).

The drastic differences in the temperature and specific humidity field produced by MCSP heating under the two model configurations is arguably due to air-sea interactions. Under the AMIP configuration, sea surface temperature (SST) is prescribed and cannot react to the impact of MCSP on the atmospheric circulation. Nonetheless, in the coupled model configuration SST can be influenced by the impact of MCSP and the resulting response is amplified. The cooling induced by MCSP heating in the lower troposphere can cause SST cooling, which lowers the latent heat fluxes and further cools the troposphere.

Although the simulated specific humidity is much reduced in the lower troposphere under the coupled model configuration with MCSP, the relative humidity profile is quite similar between the two model configurations.

Figure 16. Annual (ANN) and seasonal (DJF, JJA) averaged vertical profile of temperature (T), specific humidity (Q), and relative humidity (RH) differences from the control simulation under various configurations of Multiscale Coherent Structure Parameterization (MCSP) between 20°S/N. Solid lines are for AMIP simulations and dashed lines are for coupled simulations.
configurations (Figures 16c, 16f, and 16i). This feature implies that the temperature effect on relative humidity is stronger than that due to specific humidity. With MCSP, convection is less persistent and more low-level moisture can be preserved, especially downstream of the convective systems. This may be an important factor for the improved representation of MJO since horizontal advection of moisture is a key mechanism for its eastward propagation (Maloney et al., 2010).

5. Summary and Conclusions

We implemented the heating component of MCSP in E3SMv1 modified from its original form and investigated its impact on the simulated climate in AMIP and coupled simulations. MCSP represents important organized physical and dynamical processes induced by the MCS circulation in the form of lower-tropospheric cooling and upper-troposphere heating. The intensity of the cooling and heating is proportional to the column integral of heating by deep convection revealed by convection resolving WRF simulations. Specifically, when deep convection occurs, MCSP decreases CAPE by enhancing the tropospheric static stability, hence suppressing subsequent convection.

Recent modeling efforts have attempted to improve the simulated MJO in climate models. Implementation of the heating component of MCSP in E3SMv1 provides substantial improvement in MJO in terms of stronger eastward propagation of precipitation and OLR in the Indian Ocean basin and increased power in the eastward spectra for periods between 30 and 90 days. This suggests that over-persistent convection over this region, as simulated by the baseline model, hinders propagation. Since MCSP stabilizes the troposphere when deep convection occurs, it shortens the duration of convection within a grid-cell and assists the persistent propagation of coherent convective systems. In the simulated climate, MCSP heating reduces precipitation over the tropical western Pacific, Colombia, and Ecuador, where the baseline model simulates too much precipitation. This feature is independent of the model configurations and implies that the baseline model may have too-persistent deep convection in these regions. With MCSP heating removing some CAPE, it may serve to improve precipitation biases where deep convection is over-active in the baseline model. Under the two model configurations examined in this study, different model behavior occurs in the annual precipitation around tropical west Pacific induced by MCSP: EAMv1 simulated higher precipitation but E3SMv1 simulated lower precipitation. Our further analysis shows that MCSP enhanced precipitation in these regions in JJA and reduced precipitation in DJF. This signature was not dependent on the model configuration. It is evidently crucial to analyze the seasonal impact of MCSP instead of focusing on only the annual average since its impact can reverse sign depending upon the season.

Studies have suggested that better representation of MJO in GCM simulations often degrade the simulated mean state (see Jiang et al., 2020 for summary). However, the addition of MCSP mitigates the MJO-mean state trade-off (MLB17, Ahn et al., 2019; Moncrieff, 2019). This approach adds missing physical processes to a GCM instead of merely tuning the convection scheme. We demonstrated that an updated version of MCSP heating simultaneously improves the MJO simulation and reduces tropical precipitation biases. One key signature of the MCSP scheme is the reduction of DJF precipitation over the tropical west Pacific where the baseline model generates strong positive biases. Since total precipitation in that region is dominated by convective precipitation (not shown), which implies that convection in the baseline model is too persistent in this region. Thus, convective systems tend to stagnate when they enter this region, which may explain the lack of eastward propagation during phases 6 and 7 in the MJO life-composite plots from the baseline model (Figures 12b and 13a). The MCSP reduces persistent convection by enhancing the thermodynamic stability of the lower troposphere and thus eastward propagation in OLR may advance within this region as indicated by the MJO life-composite plots.

Herein we limited the scope of our investigation to the effects of mesoscale convective heating represented by MCSP on the simulated global climate. However, momentum transport is another important and more complex dynamical feature of organized moist convection requiring investigation. Note that, in an atmosphere-only model context, MLB17 compared the global-scale effects of MCSP convective heating and convective momentum transport and discovered structural differences. In our future MCSP work, we plan to explore and compare the effects of convective momentum transport and convective heating in atmosphere-only and coupled versions of E3SM.
Data Availability Statement

Output from the default E3SMv1 simulations can be downloaded via the following link: https://portal.nersc.gov/project/e3sm/chen24/MCSP_JAMES/

References

Ahn, M.-S., Kim, D., Kang, D., Lee, J., Sperber, K. R., Gleckler, P. J., et al. (2020). MJO propagation across the Maritime Continent: Are CMIP6 models better than CMIP5 models? Geophysical Research Letters, 47, e2020GL087250. https://doi.org/10.1029/2020GL087250

Ahn, M.-S., Kim, D., Kang, D., Y.-G. Ham, Lee, J., et al. (2019). Do we need to parameterize mesoscale convective organization to mitigate mean-state trade-off? Geophysical Research Letters, 46(4), 2293–2301. https://doi.org/10.1029/2018GL080314

Ahn, M.-S., Kim, D., Sperber, K. R., Kang, I.-S., Maloney, E., Walsper, D., et al. (2017). MJO simulation in CMIP5 climate models: MJO skill metrics and process-oriented diagnosis. Climate Dynamics, 49, 4023–4045. https://doi.org/10.1007/s00382-017-3558-4

Anber, U., Wang, S., & Sobel, A. (2014). Response of atmospheric convection to vertical wind shear: Cloud-system-resolving simulations with parameterized large-scale circulation. Part I: Specified radiative cooling. Journal of the Atmospheric Sciences, 71(8), 2976–2993. https://doi.org/10.1175/jas-d-13-0320.1

Arakawa, A., & Schubert, W. H. (1974). Interaction of a cumulus cloud ensemble with the large-scale environment. Part I. Journal of the Atmospheric Sciences, 31(3), 674–701. https://doi.org/10.1175/1520-0469-31.3.674

Bengtsson, L., Dias, J., Tulich, S., Gehne, M., & Bao, J.-W. (2021). A stochastic parameterization of organized tropical convection using cellular automata for global forecasts in NOAA’s Unified Forecast System. Journal of Advances in Modeling Earth Systems, 13, e2020MS002260. https://doi.org/10.1029/2020MS002260

Chen, B., & Mapes, B. E. (2018). Effects of a simple convective organization scheme in a two-plume GCM. Journal of Advances in Modeling Earth Systems, 10(3), 867–880. https://doi.org/10.1002/2017MS001106

D’Andrea, F., Gentine, P., Betts, A. K., & Lintner, B. R. (2014). Triggering deep convection with a probabilistic plume model. Journal of the Atmospheric Sciences, 71, 3883–3901. https://doi.org/10.1175/jas-d-13-0340.1

Gettelman, A., & Morrison, H. (2015). Advanced two-moment bulk microphysics for global models. Part I: Off-line tests and comparison with other schemes. Journal of Climate, 28(3), 1268–1287. https://doi.org/10.1175/JCLI-D-14-00102.1

Gettelman, A., Morrison, H., Santos, S., Bogenschutz, P., & Caldwell, P. (2015). Advanced two-moment bulk microphysics for global models. Part II: Global model solutions and aerosol-cloud interactions. Journal of Climate, 28(3), 1288–1307. https://doi.org/10.1175/JCLI-D-14-00103.1

Golaz, J.-C., Larson, V. E., Van Roekel, L. P., Petersen, M. R., Tang, Q., Wolfe, J. D., et al. (2019). The DOE E3SM coupled model version 1: Overview and evaluation at standard resolution. Journal of Advances in Modeling Earth Systems, 11, 2089–2129. https://doi.org/10.1029/2018MS001603

Golaz, J.-C., Larson, V. E., & Cotton, W. R. (2002). On a PDF-based model for boundary layer clouds. Part I: Method and model description. Journal of the Atmospheric Sciences, 59(24), 3540–3551. https://doi.org/10.1175/1520-0469(2002)059<3540:apbmfb>2.0.co;2

Hong, S., & Lim, J. (2006). The WRF single-moment 6-class microphysics scheme (WSM6). Journal of the Korean Meteorological Society, 42, 129–151.

Hong, S-Y., Noh, Y., & Dudhia, J. (2006). A new vertical diffusion package with an explicit treatment of entrainment processes. Monthly Weather Review, 134, 2318–2341. https://doi.org/10.1175/MWR3199.1

Houze, R. A., Jr. (2004). Mesoscale convective systems. Reviews of Geophysics, 42, RG4003. https://doi.org/10.1029/2004RG000150

Houze, R. A., Jr. (2014). Cloud dynamics. International Geophysics Series (2nd ed., Vol. 14, p. 432). Academic Press.

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, D13103. https://doi.org/10.1029/2007JD009944

Jiang, X., Adames, A. F., Kim, D., Maloney, E. D., Lin, H., Kim, H., et al. (2020). Fifty years recent progress, challenges, and of research on the Madden-Julian oscillation perspectives. Journal of Geophysical Research, 125, https://doi.org/10.1029/2019JD030911

Kain, J. S., & Fritsch, J. M. (1990). A one-dimensional entraining/detraining plume model and its application in convective parameterization. The Journal of the Atmospheric Sciences, 47, 2784–2801. https://doi.org/10.1175/1520-0469(1990)047<2784:aoedepl>2.0.co;2

Larson, V. E., & Golaz, J. C. (2017). CLUBB-SILUS: A parameterization of subgrid variability in the atmosphere. Cornell University.

Larson, V. E., & Golaz, J. C. (2005). Using probability density functions to derive consistent closure relationships among higher-order moments. Monthly Weather Review, 133(4), 1023–1042. https://doi.org/10.1175/MWR9202.1

Liebmann, B., & Smith, C. A. (1996). Description of a complete (interpolated) outgoing longwave radiation dataset. Bulletin of the American Meteorological Society, 77, 1275–1277. https://doi.org/10.1175/1520-0477-77.11.1275

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), 702–708. https://doi.org/10.1175/1520-0469(1971)028<0702:do4050>2.0.co;2

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(6), 1109–1123. https://doi.org/10.1175/1520-0469(1972)029<1109:dgsc>2.0.co;2

Maloney, E. D., Sobel, A. H., & Hannah, W. M. (2010). Intraseasonal variability in an aquaplanet general circulation model. Journal of Advances in Modeling Earth Systems, 2(2). https://doi.org/10.3894/JAMES.2010.2.5

Moncrieff, M. W. (1992). Organized convective systems: Archetypal models, mass and momentum flux theory, and parameterization. Quarterly Journal of the Royal Meteorological Society, 118, 819–850. https://doi.org/10.1002/qj.49711880703

Moncrieff, M. W. (1995). Mesoscale convective zone from a large-scale perspective. Atmospheric Research, 35, 87–112. https://doi.org/10.1016/0169-8095(94)00012-3

Moncrieff, M. W. (2019). Toward a dynamical foundation for organized convection parameterization in GCMs. Geophysical Research Letters, 46, 14103–14108. https://doi.org/10.1029/2019GL085316

Moncrieff, M. W., & Liu, C. (2006). Representing convective organization in prediction models by a hybrid strategy. Journal of the Atmospheric Sciences, 63, 3404–3420. https://doi.org/10.1175/JAS3812.1
