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The convection connection: How ocean feedbacks affect tropical mean moisture and MJO propagation

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11	Key Points:
12	• Coupled and uncoupled simulations with identical SST climatology yield non-identical
13	MJOs in four GCMs
14	• For all models, coupling enhances mean meridional moisture gradients and improves
15	MJO propagation via meridional moisture advection
16	• Coupling enhances convective moistening at high rain rates and sharpens the mois-
17	ture gradients

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18 Abstract

The response of the Madden-Julian oscillation (MJO) to ocean feedbacks is studied with 19 coupled and uncoupled simulations of four general circulation models (GCMs). Monthly 20 mean SST from each coupled model is prescribed to its respective uncoupled simulation, 21 to ensure identical SST mean state and low-frequency variability between simulation pairs. 22 Consistent with previous studies, coupling improves each model's ability to propagate 23 MJO convection beyond the Maritime Continent. Analysis of the MJO moist static en-24 ergy budget reveals that improved MJO eastward propagation in all four coupled mod-25 els arises from enhanced meridional advection of column water vapor (CWV). Despite 26 the identical mean state SST in each coupled and uncoupled simulation pair, coupling 27 increases mean-state CWV near the Equator, sharpening equatorward moisture gradi-28 ents and enhancing meridional moisture advection and MJO propagation. CWV com-29 posites during MJO and non-MJO periods demonstrate that the MJO itself does not cause 30 enhanced moisture gradients. Instead, analysis of low-level subgrid-scale moistening con-31 ditioned by rainfall rate (R) and SST anomaly reveals that coupling enhances low-level 32 convective moistening for R > 5 mm day⁻¹; this enhancement is most prominent near 33 the Equator. The low-level moistening process varies among the four models, which we 34 interpret in terms of their ocean model configurations, cumulus parameterizations, and 35 sensitivities of convection to column relative humidity. 36

37 1 Introduction

The importance of ocean feedbacks to the Madden-Julian oscillation (MJO; Mad-38 den & Julian, 1972) has been a focus of inquiry for decades (DeMott et al., 2015). While 39 consensus thinking holds that the MJO is primarily an atmospheric phenomenon, its sen-40 sitivity to SST-driven surface flux feedbacks is supported by observational (Riley Del-41 laripa & Maloney, 2015; DeMott et al., 2016), theoretical (B. Wang & Xie, 1998), and 42 modeling studies (e.g., Zhang & McPhaden, 2000; Seo et al., 2007; Klingaman & Wool-43 nough, 2014, and others). Understanding how, and the degree to which, these feedbacks 44 influence MJO intensity or propagation is fraught with challenges. For example, the ob-45 served MJO always develops in a coupled environment, yet the nature of ocean feedbacks 46 to the MJO vary from one event to the next (Gottschalck et al., 2013; Fu et al., 2015; 47 Moum et al., 2016). In models, coupling changes MJO surface fluxes, which initiates changes 48 to the entire MJO, affecting the balance of atmospheric processes that dominate MJO 49

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maintenance and propagation (e.g., DeMott et al., 2014; Klingaman & Woolnough, 2014). 50 SST-driven surface fluxes may affect the MJO through any of a variety of processes that 51 influence both the background state and the intraseasonal convection throughout the MJO 52 lifecycle. Modeling studies focused on understanding the role of coupling within the MJO 53 are hampered by different mean-state biases in coupled and uncoupled simulations that 54 themselves affect the MJO (Slingo & co authors, 1996; Sperber et al., 2005; Zhang et al., 55 2006; Klingaman & Woolnough, 2014) and complicate the interpretation of ocean feed-56 backs. 57

One way to minimize complications from mean-state differences between coupled 58 and atmosphere-only general circulation models (CGCMs and AGCMs, respectively) is 59 to prescribe temporally smoothed SSTs from the CGCM to the AGCM. While this ap-60 proach allows a cleaner comparison of intraseasonal variability in CGCM and AGCM 61 simulations, CGCM mean-state SST biases present in both the coupled and uncoupled 62 simulations may affect MJO propagation in a manner different than observed. CGCM 63 mean-state biases can be mitigated by replacing the three-dimensional (3D) dynamical 64 ocean in the CGCM with either a single-layer "slab" ocean (Waliser et al., 1999; Wat-65 terson, 2002; Maloney & Sobel, 2004; Marshall et al., 2008; Benedict & Randall, 2011) 66 or a multi-layer, one-dimensional (1D) ocean mixed-layer (Klingaman & Woolnough, 2014; 67 Tseng et al., 2015). With these approaches, SST can respond to atmospheric forcing, but 68 mean SST is constrained to the observed climatology via surface-flux adjustments (for 69 the slab ocean) or seasonally varying salt and heat advection (for the mixed layer). How-70 ever, slab- and mixed layer-coupled GCMs exclude ocean dynamic feedbacks that may 71 influence intraseasonal SST perturbations (Harrison & Vecchi, 2001; Saji et al., 2006; McPhaden 72 & Foltz, 2013; Seiki et al., 2013; Moum et al., 2013; Halkides et al., 2015). 73

Many experiments with CGCMs and AGCMs with the same SST climatology have 74 demonstrated that coupling improves several aspects of the simulated MJO, including 75 its amplitude, periodicity, propagation, and prediction (see DeMott et al. (2015), Sec-76 tion 5 for a full review). Improved phasing of SST, surface fluxes, and MJO convection 77 are often cited as reasons for these improvements (e.g., Marshall et al., 2008; Pegion & 78 Kirtman, 2008), but changes to MJO circulation (Zhang et al., 2006), including stronger 79 free tropospheric wind anomalies (Watterson, 2002; DeMott et al., 2014), and enhanced 80 boundary-layer frictional convergence ahead of MJO convection (Kemball-Cook et al., 81 2002; Benedict & Randall, 2011; Fu et al., 2015) have also been noted. 82

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83	Many of these CGCM-AGCM studies predate important advances in understand-
84	ing and diagnosing the processes that regulate MJO activity. For example, the moist static
85	energy (MSE) budget (Maloney, 2009; Andersen & Kuang, 2012) provides a framework
86	to assess processes that contribute to MJO maintenance and propagation, including the
87	relative roles of intraseasonal SST variations on MJO surface fluxes (Maloney & Sobel,
88	2004; DeMott et al., 2016); the effects of suppressed-phase conditions on subsequent MJO $$
89	propagation (Kim et al., 2014); the importance of cloud–radiative feedbacks for desta-
90	bilizing and maintaining MJO convection (Andersen & Kuang, 2012; Arnold & Randall,
91	2015); the role of background moisture gradients on MJO moistening tendencies (Kim
92	et al., 2017; Gonzalez & Jiang, 2017; DeMott et al., 2018); and SST-driven amplifica-
93	tion of boundary-layer frictional convergence east of MJO convection (W. Wang & Seo,
94	2009; Benedict & Randall, 2011; H su & Li, 2012; L. Wang et al., 2017).
95	We revisit the role of ocean coupling for the MJO using CGCM and AGCM sim-
96	ulations of four GCMs. Mean state differences between each pair of simulations are re-
97	duced by prescribing monthly mean or 31-day running mean SST from the CGCM (rather
98	than observed SSTs) to the AGCM We seek to determine if the models share a fun-
99	damental ocean feedback that improves MJO simulation, or if each model relies upon
100	its own unique coupled feedback mechanism to improve its MJO simulation. Our paper
101	is organized as follows: Models and reanalysis products are described in Section 2. MJO
102	skill metrics and MSE budgets analyses are summarized in Section 3. In Section 4, we
103	compare mean CWV during MJO and non-MJO periods to determine the effect of the
104	MJO on the meridional CWV distribution in CGCMs. In Section 5, we propose that the
105	CWV changes with coupling can be understood by considering convective moistening
106	conditioned by rainfall rate and SSTA. In Section 6, we discuss how each model's par-
107	ticular combination of cumulus parameterization and ocean vertical resolution may lead
108	to moistening characteristics that favor MJO propagation. A summary and suggestions

¹⁰⁹ for future research are provided in Section 7.

110 2 Methods

Here, we describe the models used in this study, and introduce the moist static energy (MSE) and moisture budgets. Metrics to quanitfy MJO simulation fidelity (i.e., its accuracy or realism), structure, and circulations are described in Section 3.2.

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Figure 1. November–April mean state SST (a–d) and CWV (e–h) biases, and coupled– uncoupled mean state CWV differences (i–l) for the four models analyzed. Overlaid boxes are averaging regions used to compute i) zonal, j) MJO "detour" region and k) meridional CWV differences (Section 3.2)

114 2.1 Models and Data

The four models used in this study are: the Super-Parameterized Community At-115 mospheric Model (version 3) (SPCAM3; Khairoutdinov et al., 2005), the Global Ocean 116 Mixed Layer configuration of Met Office Unified Model (MetUM; Walters et al., 2011; 117 Hirons et al., 2015)), the Max Plank Institute's European Centre-Hamburg (ECHAM, 118 v6; Stevens et al., 2013), and the Centre National de Recherches Météorologiques (CNRM; 119 Voldoire et al., 2012). SPCAM3 and CNRM are coupled to a 3D ocean model while Me-120 tUM and ECHAM are coupled to many columns of a 1D ocean mixed-layer model. For 121 the latter, oceanic salt and heat tendency profiles are prescribed to minimize mean-state 122 SST biases (see Hirons et al. (2015) for details for MetUM). The one-column ocean model 123 in ECHAM5-CPL describes changes in temperature, momentum, salinity, and turbulent 124 kinetic energy driven by vertical fluxes parameterized using the classical K approach [Tseng 125 et. al., 2015]. No such corrections are applied to coupled simulations with SPCAM3 and 126 CNRM. November–April mean SST biases for each model are shown in the left column 127 of Figure 1. In an area-averaged sense, magnitudes of the CWV biases generally reflect 128 the magnitudes of the SST biases (middle column of Figure 1). The right column of Fig-129 ure 1 is discussed in Section 3.2. 130

Coupled, uncou-	Atmosphere	Ocean (for coupled)	
pled simulation			
SPCAM3-CPL	Super-Parameterized Commu-	3D Parallel Ocean Program ver-	
SPCAM3-ATM	nity Atmospheric Model (v3);	sion 1.4.3; $\approx 3^{\circ}$ resolution near	
	$\approx~2.5^\circ\times1.8^\circ$ (Khairout dinov et	Equator (POP; Smith & Gent,	
	al., 2005; Stan et al., 2010)	2002)	
MetUM-CPL	UK Met Office Unified Model;	1D Global K Profile Parameter-	
MetUM-ATM	${\approx}1.9^{\circ}{\times}1.25^{\circ}$ (GA3; Walters et	ization ocean mixed layer; same	
	al., 2011)	resolution as atmosphere (Large	
		et al., 1994; Hirons et al., 2015)	
ECHAM-CPL	ECHAM (v5); $\approx 1.8^{\circ} \times 1.8^{\circ}$	1D Snow-Ice-Thermocline (SIT);	
ECHAM-ATM	(Stevens et al., 2013)	same resolution as atmosphere	
		(Tseng et al., 2015)	
CNRM-CPL	CNRM (v5.2); \approx 1.4° × 1.4°	3D NEMO; ${\approx}1^{\circ}$ resolution near	
CNRM-ATM	(Voldoire et al., 2012)	Equator (Madec, 2008)	

 Table 1. Atmospheric and oceanic model descriptions and resolution.

Each CGCM is integrated for 20–25 years. Monthly mean (SPCAM3 and CNRM) 135 or 31-day running mean (MetUM and ECHAM) SST time series from each CGCM are 136 prescribed to its respective AGCM. This ensures that the CGCM and AGCM have iden-137 tical SST mean state and low-frequency variability; only SST variability on frequencies 138 higher than 31 days is absent in the AGCMs. Removal of high-frequency SST variabil-139 ity from the AGCM may affect our results, but including this variability by prescribing 140 daily mean SST to the AGCM, for example, is known to unrealistically alter the phas-141 ing of rainfall and SST (e.g., Pegion & Kirtman, 2008; DeMott et al., 2015). The four 142 CGCMs are denoted as SPCAM3-CPL, MetUM-CPL, ECHAM-CPL, and CNRM-CPL, 143 while their AGCM counterparts are SPCAM3-ATM, MetUM-ATM, ECHAM-ATM, and 144 CNRM-ATM, respectively. Model descriptions and references are summarized in Table 1. 145

- Results from all simulations are compared to daily mean data from the European
 Centre for Medium Range Forecasts (ECWMF) Interim Reanalysis (ERAI; Dee & co authors, 2011) on a 2.5°×2.5° grid for 1986–2013 (Dee & co authors, 2011).
- 150 2.2 Analysis Methods

Processes responsible for the maintenance and propagation of MJO convection are assessed with the aid of the moist static energy (MSE) budget:

$$\frac{\partial \langle m \rangle}{\partial t} = -\langle V \cdot \nabla m \rangle - \langle \omega \partial m / \partial p \rangle + \langle LW \rangle + \langle SW \rangle + LH + SH \tag{1}$$

where m is the MSE, defined as $m = C_p T + L_v q + gz$ [J kg⁻¹], T is temperature [K], q 153 is specific humidity [kg kg⁻¹], z is height [m], V is the horizontal wind [m s⁻¹], ω is the 154 vertical pressure velocity [Pa s^{-1}], LH and SH are surface latent and sensible heat fluxes 155 $[W m^{-2}]$, respectively, and LW and SW are longwave and shortwave radiative heating 156 $[K s^{-1}]$. Angled brackets denote integration from the surface to 100 mb, yielding units 157 of $[W m^{-2}]$ for all terms. From left to right, terms on the right hand side of Equation 1 158 are MSE horizontal and vertical advection, column-integrated longwave and shortwave 159 radiative heating, and surface latent and sensible heat fluxes. 160

In the Tropics, MSE tendencies are largely driven by moisture tendencies, as the 161 weak Coriolis force enables rapid dissipation of temperature and density anomalies via 162 gravity waves. Compared to moisture budgets, MSE budgets are useful for understand-163 ing tropical convective variability because rainfall and MSE are highly correlated, MSE 164 is conserved during diabatic phase changes (eliminating the need to accurately observe 165 rainfall), and the column MSE budget includes radiative and surface sensible heat fluxes. 166 MSE budget analysis is a standard tool to assess processes responsible for MJO main-167 tenance and propagation (e.g., Maloney, 2009; Andersen & Kuang, 2012). 168

Each term in Equation 1 is computed using daily mean input variables; its anomaly 169 is computed as the difference between the daily mean value and the slowly varying "back-170 ground" state, which is obtained by applying a low-pass filter with a 100 day cutoff to 171 the daily mean time series. This partitions subseasonal and higher-frequency variabil-172 ity into the anomaly time series, and seasonal and lower-frequency variability into the 173 background-state time series. SST and other state variables are partitioned the same way. 174 Intraseasonal rainfall anomalies are obtained with a 20-100 day 201 point Lanczos fil-175 ter. 176

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We also computed daily mean, vertically resolved moisture budget terms and the moisture budget residual, which is the apparent moisture sink $(Q_2; \text{Yanai et al., 1973})$:

$$Q_2/L_v = -\left[\frac{\partial q}{\partial t} + \nabla \cdot (qV) + \frac{\partial (q\omega)}{\partial p}\right] = (c-e) + \frac{\partial (q'w')}{\partial p}$$
(2)

where q is the specific humidity, c and e are the condensation and evaporation of water vapor, respectively, and $\partial(q'w')/\partial p$ is the unresolved vertical eddy flux of water vapor. When written as $-Q_2/L_v$ [kg kg⁻¹ s⁻¹], positive values of the budget residual represent unresolved moistening by convection (Section 5). For ERAI, $-Q_2/L_v$ also includes moistening from data assimilation increments.

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3 MJO simulation assessment

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3.1 MJO propagation and MSE budget overview

MJO propagation in ERAI and the eight simulations is illustrated with rainfall lagged-186 regression plots (Figure 2). ERAI and the coupled simulations (left column) show co-187 herent propagation across the Indian Ocean and Maritime Continent. Robust propaga-188 tion continues to about 165°E in ERAI, SPCAM3-CPL, and CNRM-CPL, while weaker 189 propagation is observed in MetUM-CPL and ECHAM-CPL. The 20–100 day rainfall stan-190 dard deviation (R) for each CGCM is larger than that in ERAI. Some of this difference 191 may be a reflection of suspected "missing" rainfall processes in ERAI, as indicated by 192 less frequent heavy rainfall rates compared to satellite-derived rainfall products (e.g., Adames 193 et al., 2017). R in AGCMs (right column) is reduced in all models but ECHAM-ATM, 194 an indication that improved MJO fidelity with coupling is not uniformly linked to en-195 hanced intraseasonal heating variability. AGCMs produce westward-propagating (SPCAM3-196 ATM), stationary (MetUM-ATM), or weakly eastward-propagating (ECHAM-ATM, CNRM-197 ATM) disturbances. MJO propagation fidelity is assessed using a method adapted from 198 Jiang et al. (2015): the MJO "pattern correlation" metric, r, is the correlation between 199 the rainfall lagged-regression diagram for each model (Figures 2b-i) with the ERAI lagged-200 regression diagram (Figure 2a). Correlations are computed for regressions against 90°E 201 and 150°E basepoints (not shown) and averaged. Rainfall within $\pm 15^{\circ}$ longitude of each 202 basepoint is omitted since it unfavorably weights the result by MJO periodicity, rather 203 than by MJO propagation (c.f., L. Wang et al., 2017). 204

Figure 3 presents CGCM MSE component contributions (Equation 1) to MSE maintenance (Figure 3a) and tendency (Figure 3b) integrated over the MJO lifecycle and the

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Figure 2. Lag-regression of November–April 15° S- 15° N averaged rainfall anomalies onto 206 20–100 day filtered rainfall at a 90°E basepoint (80° E– 100° E and 15° S– 15° N area average). Cou-207 pled (uncoupled) systems are plotted in the left (right) column. 20–100 day basepoint rainfall 208 standard deviation (R), and the average pattern correlation for 90°E and 150°E basepoints (r; 209 see text for details) are listed on each panel. Units for R and contoured values are mm day⁻¹. 210 Stippling denotes significance at the 95% conficence interval.

Warm Pool. Following DeMott et al. (2016), these contributions were computed by re-213 gressing individual MSE component anomalies onto November–April 20–100 day filtered 214 MSE and its tendency, respectively, at each grid point. Regression coefficients are av-215 eraged from 15° S to 15° N and 30° E to 240° E to include the Warm Pool and regions far-216 ther east affected by MJO circulation anomalies (results are similar for a 180°E eastern 217 boundary). Figure 3 combines $\langle LW \rangle$ and $\langle SW \rangle$ and LH and SH; the second term in 218 each pair is an order of magnitude smaller than the first. Consistent with many previ-219 ous studies (Andersen & Kuang, 2012; Kiranmayi & Maloney, 2011; DeMott et al., 2016), 220 MSE anomalies are principally maintained by $\langle LW \rangle$. MSE vertical advection ($\langle -\omega \frac{\partial m}{\partial p} \rangle$; 221 VADV) and surface fluxes are both MSE sinks (SPCAM3-CPL is an exception, where 222 vertical advection is small source term). The MSE tendency, which maximizes to the east 223 of MJO convection, and contributes to its eastward propagation, is dominated by hor-224 izontal MSE advection ($\langle -V \cdot \nabla m \rangle$; HADV), with minor contributions from $\langle -\omega \frac{\partial m}{\partial p} \rangle$. 225 From this area- and lifecycle-averaged perspective, surface fluxes do not significantly con-226 tribute to MJO maintenance or propagation. 227

This raises an important question: How can ocean feedbacks—which are commu-228 nicated to the atmosphere through SST-modulated surface fluxes—so distinctly improve 229 the fidelity of the simulated MJO? DeMott et al. (2016) analyzed the spatial structure 230 of SST-modulated surface flux contributions to the MJO with ERAI data and found that 231 direct ocean feedbacks account for up to 1-2% of MSE maintenance near the Equator 232 and $\approx 10\%$ of MSE tendency across the Warm Pool. We obtained similar results for the 233 CGCMs (not shown). Based on these findings, it seems unlikely that direct ocean feed-234 backs to MJO convection or its tendency can explain the improved MJO propagation 235 in the coupled simulations. Instead, coupled surface fluxes most likely affect MJO prop-236 agation through more indirect feedbacks, such as altering the atmospheric stability pro-237 file to strengthen the circulation response to MJO heating, or by changing mean state 238 conditions. 239

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3.2 MJO and process metrics

To better understand how coupling affects the MJO, we computed a variety of metrics of MJO propagation, period, circulation structure, and mean state, and compared how these metrics change with coupling. Here, we define each metric and its shorthand name, which is later referenced in Figure 5. MJO propagation is assessed with the afore-

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Figure 3. November–April area-averaged (15°S-15°N; 30°E-240°E) covariability of MSE budget terms (Equation 1) with 20–100 day filtered a) $\langle m \rangle$ and b) $\langle \frac{\partial m}{\partial t} \rangle$.

- mentioned pattern correlation ("Pattern Corr."). Another measure of MJO propagation
 is the east-west power ratio which is the ratio of rainfall variance for eastward- and westwardpropagating zonal wave numbers 1–3 and periods 30–60 days ("EW ratio;" Jiang et al.,
 2015). MJO period is defined as twice the number of days between the maximum and
 minimum lagged auto-correlation of 20-100 day filtered 90°E rainfall ("Period").
- MJO circulation metrics are computed with the aid of maps of anomalous 850 hPa 252 zonal winds (Figure 4a–e), $\langle LW \rangle$ (Figure 4f–j), and 850 hPa vertical moisture advection 253 (Figure 4k-o) regressed onto 20-100 day band pass filtered 90°E rainfall. MJO-associated 254 zonal wind anomalies show an elongated region of low-level easterlies to the east, and 255 a more truncated region of low-level westerlies to the west. The low-level easterlies arise 256 from the Kelvin-wave response to Indian Ocean convective heating (Gill, 1980) and the 257 equatorial Rossby-wave response to Maritime Continent and West Pacific longwave cool-258 ing (Kim et al., 2012). The low-level westerlies are part of the equatorially symmetric 259 Rossby wave response to positive Indian Ocean heating. The magnitudes of the Kelvin-260 wave easterlies ("KW wind") and Rossby-wave westerlies ("ER wind") are the maximum 261

 5° N to 5° S averaged easterly and westerly wind anomalies. Their ratio ("KW/ER ra-262 tio") is positively correlated with MJO propagation fidelity (B. Wang et al., 2018). A 263 "dry phase intensity" index, defined as anomalous longwave cooling $< -1 \, [W m^{-2}]/[mm day^{-1}]$ 264 integrated from 20°S–20°N and 100°E–210°E (the "Dry Phase" metric), measures the 265 strength of longwave cooling that initiates poleward flow east of MJO convection as part 266 of the anticyclonic Rossby gyre response to the negative heating anomaly (Kim et al., 267 2012). 850 hPa vertical moisture advection, $-\omega \frac{\partial q}{\partial p}|_{850}$, averaged over 5°S–5°N and 110°E– 268 180°E (the "BL export" metric) measures low-level convergence-driven export of mois-269 ture from the boundary layer to the free troposphere, which may be critical to MJO prop-270 agation (e.g., Hsu & Li, 2012; B. Wang et al., 2018). 271

MJO propagation is critically sensitive to mean-state moisture gradients, which reg-272 ulate horizontal moisture advection by MJO wind anomalies (Kim et al., 2012, 2017; Gon-273 zalez & Jiang, 2017; Lim et al., 2018; DeMott et al., 2018). November-April mean CWV 274 for ERAI and CGCMs and the averaging regions used to compute moisture gradient met-275 rics are shown in Figure 1i–k. The Warm Pool zonal moisture gradient ("dCWV/dx (WP)") 276 is the difference between CWV averaged over the Maritime Continent ($10^{\circ}S-10^{\circ}N$; $110^{\circ}E-$ 277 $130^{\circ}E$) and the central Indian Ocean ($10^{\circ}S-10^{\circ}N$; $60^{\circ}E-80^{\circ}E$). We also compute the zonal 278 moisture gradient in the MJO "detour" region (Kim et al., 2017) as the difference be-279 tween area-averaged CWV north of Australia (20°S–10°S; 120°E–135°E) and south of 280 Sumatra (20°S–10°S; 105°E–120°E (the "dCWV/dx (Aus)" metric). Northern ("dCWV/dy 281 (nIO)") and southern ("dCWV/dy (sIO)") meridional moisture gradients are computed 282 as the area-averaged CWV difference between the November–April "moisture Equator" 283 $(10^{\circ}\text{S}-0^{\circ}\text{N}; 40^{\circ}\text{E}-120^{\circ}\text{E})$ and $(0^{\circ}\text{N}-10^{\circ}\text{N}; 40^{\circ}\text{E}-120^{\circ}\text{E})$ and $(20^{\circ}\text{S}-10^{\circ}\text{S}; 40^{\circ}\text{E}-120^{\circ}\text{E})$, 284 respectively. Northern and southern gradients are averaged to yield the mean equator-285 ward moisture gradient ("dCWV/dy (IO)"). 286

295

3.3 Changes to MJO metrics with coupling

The effects of coupled feedbacks on the above MJO metrics and MSE budget term contributions to MJO maintenance and propagation are summarized in Figure 5. Changes to metrics that characterize MJO propagation, circulation, intensity, and period are plotted in Figure 5a as the coupled minus uncoupled percentage difference (i.e., $\Delta M = 100$ · $(M_{CPL}-M_{ATM})/M_{ATM}$ where subscripts "CPL" and "ATM" refer to CGCM and AGCM metrics, respectively). As previously noted, coupling does not uniformly increase intrasea-



Figure 4. Anomalous 850 hPa zonal wind (a-e), vertically integrated longwave heating (f-j), 287 and 850 hPa vertical moisture advection (k-o) regressed onto 20–100 day filtered rainfall for a 288 $90^{\circ}E$ basepoint (i.e., the $80^{\circ}E-100^{\circ}E$; $15^{\circ}-15^{\circ}E$ area average). Units listed in each panel are per 289 1 mm day^{-1} of basepoint rainfall. Results for coupled (uncoupled) systems are shaded (con-290 to ured; interval as in shaded field). Shaded regions correspond to significance at the 95% confi-291 dence interval over most of the domain. Red (blue) contours in a-j are positive (negative) SST 292 contours (units [0.1 K]/[mm day⁻¹]). Magenta boxes denote averaging regions used to compute 293 MJO metrics (see text and Figure 5). 294

sonal rainfall variability. Similarly, coupling increases MJO period in some CGCMs and 302 reduces it in others. The MJO pattern correlation, east-west power ratio, KW/ER asym-303 metry, and dry phase intensity metrics all increase with coupling, and are significantly 304 correlated with one another, as shown in Table 2. This suggests that these indices should 305 be thought of as MJO metrics, which characterize MJO fidelity, rather than as process 306 metrics, which measure the magnitude of processes that maintain or propagate MJO con-307 vection. For example, consider the case of widespread subsidence east of MJO convec-308 tion. Chen and Wang (2018) report that the western part of the subsiding region arises 309 from compensating subsidence driven by upper-level divergence associated with Indian 310 Ocean MJO convection, while the eastern part of the subsiding region arises from wake 311 subsidence from the previous MJO event. The dry phase intensity metric, then, is an MJO 312 metric because it combines measures of MJO propagation and upper-level circulation. 313 The low-level poleward flow associated with the dry phase advects mean-state moisture 314 poleward, which is characterized using the meridional advection process metric. 315

Zonal mean state moisture gradients that promote MJO propagation do not uni-319 formly increase with coupling (dCWV/dx (WP) and dCWV/dx Aus; Fig. 5b). Merid-320 ional mean-state moisture gradients, however, uniformly increase with coupling and many 321 of these increases are statistically significant. This unexpected result warrants further 322 consideration. Conventional wisdom holds that constraining CGCM and AGCM sim-323 ulations of the same model to the same mean-state SST effectively eliminates mean-state 324 CWV differences that affect MJO propagation (e.g., Klingaman & Woolnough, 2014). 325 Instead, our results demonstrate that, despite no change in SST climatology and low-326 frequency variability, coupled feedbacks consistently yield mean-state CWV patterns that 327 favor MJO propagation (Figure 1i–l). Higher-frequency ocean-atmosphere interactions 328 (i.e., 30-day or shorter timescales) adjust, or rectify onto, the mean-state moisture dis-329 tribution to favor MJO propagation. This is the result of an asymmetry or non-linearity 330 of moistening processes for positive and negative SST anomalies in coupled models that 331 yield a "rectified" non-zero net moistening. For the atmosphere, the "rectifier effect" can 332 be seen as the net effect of a short or small scale process on a longer or larger scale pro-333 cess (e.g., Denning et al., 1999; Kessler & Kleeman, 2000; Shinoda & Hendon, 2002). 334

How do coupled feedbacks affect maintenance or propagation of intraseasonal MSE anomalies? Coupled minus uncoupled projections of MSE budget terms (Eq. 1) onto 20– 100 day filtered $\langle m \rangle$ and $\frac{\partial \langle m \rangle}{\partial t}$ are shown in Figures 5c and 5d, respectively. Note that,

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Table 2. Correlations between MJO skill and descriptor metric across ERAI and the eight

simulations. Significance at the 95% (90%) confidence interval is shown in bold (italicized bold)

text. Metrics are defined in Sections 3.1 and 3.2.

Metric	Pattern	EW	Period	Rain	KW/ER	KW _{max}	ER_{max}	Dry	BL exp.
	Corr.	ratio		Amp.	ratio			Phase	
Pattern		0.66	0.35	-0.30	0.92	0.74	-0.11	0.79	0.25
Corr.									
EW	0.66		0.62	0.21	0.75	0.79	0.13	0.33	0.23
ratio									
Period	0.35	0.62		0.07	0.53	0.69	0.57	-0.09	-0.09
Rain	-0.30	0.21	0.07		-0.21	0.28	-0.39	-0.45	-0.37
Amp.									
KW/ER	0.92	0.75	0.53	-0.21		0.84	0.02	0.63	0.26
ratio									
KW_{max}	0.74	0.79	0.69	-0.28	0.84		0.49	0.57	0.51
ER_{max}	-0.12	0.13	0.57	-0.39	0.02	-0.49		-0.13	0.30
Dry	0.79	0.33	-0.09	-0.45	63	0.57	-0.13		0.65
Phase									
BL exp.	0.25	0.23	-0.09	-0.37	0.26	-0.51	0.30	0.65	

in contrast to the relative differences plotted in Figure 5a, coupled minus uncoupled MSE 338 projections are plotted as absolute differences. We find that coupling has no consistent 339 effect on $\langle m \rangle$ maintenance, and that the effect of coupling on $\langle m \rangle$ maintenance by ver-340 tical MSE advection ($\langle -\omega \frac{\partial m}{\partial p} \rangle$; VADV) and surface fluxes may be sensitive to whether 341 the AGCM is coupled to a 3D or 1D ocean (i.e., yellow bars in Figure 5c). We elabo-342 rate on these differences in Section 6. Changes in processes that support $\frac{\partial \langle m \rangle}{\partial t}$ lend fur-343 ther weight to the argument that mean-state moisture gradient changes improve MJO 344 propagation in the CGCMs. Horizontal moisture advection $(\langle -V \cdot \nabla m \rangle; \text{HADV})$ sig-345 nificantly enhances $\frac{\partial \langle m \rangle}{\partial t}$ and MJO propagation in all coupled simulations. By decom-346 posing the advection term into its zonal $(\langle -u\frac{\partial m}{\partial x}\rangle; u\text{-HADV})$ and meridional $(\langle -v\frac{\partial m}{\partial y}\rangle;$ 347 v-HADV) components, it is clear that enhanced meridional MSE advection is the com-348 mon process responsible for enhanced MJO propagation in the CGCMs. Furthermore, 349 close inspection of Figure 5d reveals that meridional MSE advection increases are larger 350 than increases in nearly all other terms. The sole exception is vertical MSE advection 351 in SPCAM3-CPL. We consider this difference further in Section 6. Coupling uniformly 352 reduces surface flux contributions to MSE tendency, thereby inhibiting, rather than en-353 couraging, MJO eastward propagation. The combination of mean-state low-level west-354 erlies and stronger MJO-associated low-level easterlies in CGCMs reduces the total wind 355 speed and surface fluxes east of MJO convection in CGCMs. 356

³⁶⁶ 4 Cause or effect: MJO and mean-state moisture

The above analysis demonstrates the connection between improved MJO propagation and sharpened meridional moisture gradients in CGCMs, but does not provide any insight into causality. It is possible that coupled feedbacks alter MJO convection and circulations to enhance the equatorward CWV gradients, so that sharpened meridional moisture gradients are a consequence, rather than a cause, of improved MJO propagation.

To test this hypothesis, we used a version of the Real-time Multivariate MJO indices (RMM; Wheeler & Hendon, 2004) adapted for climate-length simulations (Madden-Julian Oscillation Working Group, 2009) to identify MJO and non-MJO periods. The RMM indices are obtained through the multivariate empirical orthogonal function (EOF) decomposition of 20–100 day filtered outgoing longwave radiation (OLR) and zonal wind at 200 and 850 hPa. The two leading EOFs capture the eastward propagating MJO; their

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Figure 5. CGCM minus AGCM differences for a) MJO descriptive metrics, b) mean state 357 CWV gradients, c) MSE maintenance and d) MSE tendency by MSE budget terms (Eq. 1). 358 Same-signed changes across all four models are denoted with gray background stripes. Changes 359 that differ according to ocean model (i.e., 3D or 1D) are denoted with yellow background stripes. 360 Significance at the 90% and 95% confidence levels are denoted with small and large open circles, 361 respectively, while \times indicates differences that are not significant. For a) and b), significance was 362 assessed with a student's t-test based on multiple three- and one-year data subsets, respectively. 363 In c) and d), significance was tested with a student's t-test based on the area-averaged regression 364 365 coefficients for the domain plotted in Figure 4.

principal component (PC) time series can quantify MJO phase and amplitude. We computed the RMM EOFs from NOAA daily mean OLR (Liebmann, 1996) and ERAI winds and projected these onto model output, to obtain RMM indices for each CGCM. The PC timeseries are normalized, and MJO and non-MJO periods are identified as periods when the RMM amplitude ($A = \sqrt{PC_1^2 + PC_2^2}$) is greater than or less than one, respectively.

To determine if the improved MJO in the CGCM is responsible for the sharpened 385 meridional moisture gradient, we computed the average CWV difference for MJO and 386 non-MJO periods in the CGCM and compared it to the average CWV difference between 387 the CGCM and AGCM. MJO and non-MJO CWV differences were computed for each 388 month (November-April) and then averaged across the entire season to account for sea-389 sonal shifts in MJO amplitude that arose in some CGCMs. The results are shown in Fig-390 ure 6a-e. Shading (CGCM MJO minus non-MJO) and contours (CGCM minus AGCM) 391 are plotted using the same contour interval to emphasise the relative magnitudes of the 392 two differences. Except for CNRM, MJO minus non-MJO CWV differences are much 393 smaller than coupled minus uncoupled CWV differences. Pattern correlations between 394 the two fields for Indian Ocean and western Pacific regions are listed in each panel. Neg-395 ative or small positive correlations for the Indian Ocean suggest that the MJO either weakly 396 disperses moisture away from the Equator, or has little net effect on the CWV distri-397 bution. 398

In the western Pacific, MJO minus non-MJO differences (shading) for ERAI and 407 CNRM-CPL bear a strong resemblance to El Niño conditions, which are known to in-408 fluence MJO propagation (Pohl & Matthews, 2007; DeMott et al., 2018). To remove ENSO 409 influences, we repeated the analysis with 100 day high-pass filtered data (Figure 6f-j), 410 which reduces MJO minus non-MJO differences. Although the western Pacific and In-411 dian Ocean correlations increase for SPCAM3-CPL and CNRM-CPL, respectively, the 412 small magnitude of MJO minus non-MJO changes in these regions ($<1 \text{ kg m}^{-2}$) is likely 413 insufficient to explain the CGCM minus AGCM CWV differences. We therefore reject 414 the hypothesis that the improved MJO in the CGCMs causes enhanced equatorward mois-415 ture gradients. 416

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Figure 6. November–April CWV differences for CGCM MJO periods (RMM > 1) and non-399 MJO periods (RMM < 1) (shading; shaded values correspond to 95% or greater significance over 400 most of domain) and CGCM minus AGCM CWV differences (contours; negative values dashed 401 and zero contour omitted; 95% significance stippled). Shading and contours are plotted with the 402 same contour interval. CGCM MJO minus non-MJO CWV differences for unfiltered (100 day 403 high-pass filtered) input are plotted on the left (right). Pattern correlations for Indian Ocean and 404 western Pacific Ocean regions (magenta boxes in panel a; rIO and rWP, respectively) are listed 405 on each panel. 406

⁴¹⁷ 5 How do ocean feedbacks rectify onto mean state CWV?

418

5.1 Evaluation of processes that could affect CWV differences

Understanding how coupling leads to mean-state moisture changes in the four models is key to understanding how coupling improves MJO propagation. While the models differ in physical parameterizations and ocean models (Table 1), they uniformly exhibit sharpened meridional moisture gradients and improved MJO propagation with coupling. This raises the question of whether the models arrive at their coupled mean states through a common process or a unique process in each model.

We examined several processes that might be linked to the mean-state moisture 425 differences. The processes, and the models whose process change is correlated with the 426 CWV change (shown in parentheses, plus any relevant references) are: enhanced Hadley 427 circulation over the Warm Pool (SPCAM3-CPL, MetUM-CPL); reduced vertical com-428 ponent of gross moist stability (ECHAM-CPL; Neelin, Held, and Cook (1987); Benedict, 429 Maloney, Sobel, Frierson, and Donner (2013); enhanced support of convection by sur-430 face latent heat fluxes (SPCAM3-CPL, CNRM-CPL; Riley Dellaripa and Maloney (2015); 431 DeMott et al. (2016)); and enhanced longwave radiative feedbacks (none; Del Genio and 432 Chen (2015)). These findings would support the "model-dependent pathway" paradigm 433 for mean state CWV changes with coupling, but remain unsatisfying since they gener-434 ally offer few insights into how ocean feedbacks affect a given process. 435

436

5.2 Insights from Q_2 profiles for understanding CWV differences

A new method for analyzing the response of convection to coupling reveals a com-437 mon process that may explain the CGCM-minus-AGCM CWV differences. First, we con-438 structed CGCM rainfall rate probability distribution functions (PDFs) for warm and cold 439 daily mean SSTA periods for the domain 15°S–15SoN and 50°E–180°E (Figure 7a; re-440 sults for 20°S–20SoN and 30°E–240°E are virtually identical). Non-rainy days are ex-441 cluded by estimating the minimum allowable rainfall rate ($\approx 0.05 \text{ mm day}^{-1}$) that yields 442 a first-bin frequency that matches an estimated first-bin frequency obtained by extrap-443 olating the PDF curve from the second-through-fourth bins. The warm-minus-cold PDF 444 differences (Figure 7c) reveal model-dependent changes to the PDFs: ERAI, SPCAM3-445 CPL, and CNRM-CPL shift toward heavier rainfall rates during warm SSTA periods, 446 while MetUM-CPL and ECHAM-CPL contract toward moderate rainfall rates. Rain-447

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Probability distribution functions (PDFs) of a) $log_{10}R$ for ERAI and CGCM warm Figure 7. 451 (solid) and cold (dashed) SST anomalies (SSTA) and b) CGCM (solid) and AGCM (dashed) sim-452 ulations. PDF differences for c) ERAI and CGCM warm - cold SSTA ($\Delta f = 100 \cdot (f_+ - f_-)/f_-$ 453 where f is the fraction of observations per rainfall rate bin and + and - subscripts refer to 454 warm and cold SSTA, respectively) and d) CGCM - AGCM simulations (($\Delta f_{CPL-ATM}$) 455 = $100 \cdot (f_{CPL} - f_{ATM})/f_{ATM}$ where CPL and ATM subscripts refer to CGCM and AGCM simu-456 lations, respectively). Data are drawn from ocean-only points from 15°S-15S°N and 50°E-180°E. 457 All distribution differences in c) and d) were tested with a Kolmogorov-Smirnov test and are 458 significant at the 95% confidence level. 459

fall rate PDFs for CGCMs and AGCMs, and their intra-bin differences, are shown for
comparison in Figures 7b and 7d, respectively. In general, CGCM-minus-AGCM PDF
differences (Figure 7d) are smaller than the warm-minus-cold SSTA differences (Figure 7c).

We next analyzed Yanai's "apparent moisture sink," Q_2 (e.g., subgrid-scale or un-460 resolved moistening by convection; Yanai et al., 1973) conditioned by rainfall rate and 461 SSTA for ERAI and the models. The results, plotted as $-Q_2/L_v$ with units of g kg⁻¹ day⁻¹ 462 so that positive values represent moistening, are shown in Figure 8. November-April mean 463 $-Q_2/L_v$ profiles from the CGCMs (gray contours) demonstrate the broad similarity of 464 convective moistening characteristics for ERAI and all models: convection (via param-465 eterized physics) moistens low levels and dries upper levels at low rainfall rates, while 466 the opposite is observed at high rainfall rates. The warm-minus-cold $-Q_2/L_v$ difference 467

⁴⁶⁸ profiles (shading in Figure 8; left column), however, are not consistent across all mod-

- els. We are especially interested in low-level convective moistening (i.e., below about 700 hPa)
- since MJO propagation via horizontal moisture advection is most sensitive to lower-tropospheric
- ⁴⁷¹ moisture patterns (e.g., Gonzalez & Jiang, 2017). For ERAI, SPCAM3-CPL, and CNRM-
- 472 CPL, low-level convective moistening above the atmospheric mixed layer (above 925 hPa)
- ⁴⁷³ is enhanced for nearly all rainfall rates during warm SSTA periods, but is reduced for
- ⁴⁷⁴ MetUM-CPL and ECHAM-CPL. This would appear to reinforce the model-dependent
- ⁴⁷⁵ pathway for CWV changes with coupling, but as we demonstrate next, it is the combi-
- nation of warm-minus-cold rainrate PDFs and warm-minus-cold $-Q_2/L_v$ profiles that
- explains the uniform increase in equatorial CWV in all four models.

The warm-minus-cold $-Q_2/L_v$ differences (Figure 8; left column) weight equally 485 the warm and cold SSTA $-Q_2/L_v$ averages at each rainfall rate bin. However the dis-486 tribution of rainfall rates for warm and cold SSTA periods is unequal (Figure 7b). To 487 understand how coupling affects convective moistening, we must account for both dif-488 ferences, by multiplying the warm-minus-cold $-Q_2/L_v$ difference profiles (Figure 8; left 489 column) by their respective intra-bin rainfall rate PDF differences (Figure 7b); Figure 8 490 (center column) shows the product. For ERAI, SPCAM3-CPL, and CNRM-CPL, more 491 frequent heavy rainfall and less frequent light rainfall during warm SSTA periods yields 492 less low-level moistening at low rainfall rates, and more low-level moistening at high rain-493 fall rates. In contrast, for MetUM-CPL and ECHAM-CPL, the reduced frequency of heavy 494 rainfall rates for warm SSTA reduces low-level drying during warm SSTA periods and 495 increases low-level moistening during cold SSTA periods. The net effect for all four mod-496 els is an *increase* in low-level convective moistening at high rainfall rates (i.e., $R \ge 5 \text{ mm day}^{-1}$). 497 Note that we would get the same result had we instead analyzed cold-minus-warm SSTA 498 differences. These results, therefore, summarize the net effect of coupling, not simply the 499 effect of warm SSTAs. 500

While the warm-minus-cold SSTA-weighted $-Q_2/L_v$ differences (Figure 8; center column) illustrate the essential effects of ocean coupled feedbacks on convective moistening, the results do not account for the overall distribution of rainfall rates. The final step in the analysis is to multiply the warm-minus-cold SSTA-weighted $-Q_2/L_v$ differences (Figure 8; center column) by the normalized ERAI or CGCM rainfall rate PDF (Figure 7c), where the PDF is normalized by its maximum value. This step reduces the magnitude of $-Q_2/L_v$ differences at very low and very high rainfall rates (Figure 8; right



Figure 8. Average $-Q_2/L_v$ vertical profiles binned by rainfall rate (solid) and left) the $-Q_2/L_v$ difference for warm-minus-cold SSTA (shading; $-\Delta Q_2/L_v$); center) $-\Delta Q_2/L_v$ (left column) multiplied by Δf (Figure 7c); and right) $\Delta f \cdot (-\Delta Q_2/L_v)$ (center column) multiplied by the normalized f_{CPL} (Figure 7b), f_N . Warm-minus-cold $-\Delta Q_2/L_v$ differences in center and right columns are plotted only where statistically significant at the 95% confidence interval. Magenta boxes in right column highligt moistening at $R > 5 mm \ day^{-1}$ and below 600 hPa. Data are drawn from same domain as in Figure 7.

column), but otherwise does not change the rainfall rates associated with the largest moistening differences. Figure 8 (center and right columns) illustrates the rectification of ocean

feedbacks onto moistening by parameterized physics.

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- 512

5.3 The relationship between SST- and rainfall rate-conditioned $-Q_2/L_v$ and mean state CWV differences

The above suggests two criteria for coupled feedbacks to enhance local low-level convective moistening: (a) the region should be dominated by rainfall rates greater than about 5 mm day⁻¹; (b) these heavy rainfall rate regions should exhibit more frequent SSTAs whose sign (positive or negative) is consistent with enhanced low-level moistening for that model.

To assess whether the above criteria apply to these models, we plot maps of the 518 frequency of R > 5 mm day⁻¹ for each CGCM (left column of Figure 9). The frequency 519 of $R > 5 \text{ mm day}^{-1}$ maximizes near the Equator, satisfying the first criterion for en-520 hanced low-level moistening by unresolved convective processes. Whether these regions 521 experience enhanced or reduced moistening, however, depends on whether they are dom-522 inated by warm SSTAs (SPCAM3-CPL and CNRM-CPL), or cold SSTAs (MetUM-CPL 523 and ECHAM-CPL). Maps of the frequency difference for warm and cold SSTAs during 524 heavy rain condition (center column of Figure 9) show that for SPCAM3-CPL and CNRM-525 CPL, the heavy rain-dominated regions are dominated by warm SSTAs, consistent with 526 enhanced low-level moistening by convection. Although the regions of frequent heavy 527 rainfall and frequent warm SSTA do not everywhere agree with CWV differences, this 528 analysis does not consider circulation effects on CWV that may modulate the overall CWV 529 distribution. Nevertheless, the spatial distribution of the frequency of heavy rainfall and 530 warm SSTAs suggest that changes in low-level convective moistening are responsible for 531 the changes in CWV between CGCM and AGCM simulations for these two models. For 532 MetUM-CPL and ECHAM-CPL, the frequent heavy-rain regions are dominated by cold 533 SSTAs, which is consistent with the enhanced equatorial low-level moistening and CWV 534 changes for these two models. Daily SST variability is generally large in the regions of 535 frequent heavy rainfall (e.g., DeMott et al., 2016), which may also help localize the ocean-536 to-convective moistening feedback. 537

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 5 mm day^{-1} ; center) warm-minus-Figure 9. November–April frequency of left) R \geq 538 cold SSTA frequencies of R \geq 5 mm day $^{-1}$; and right) CGCM-minus-AGCM frequencies of 539 $R \geq 5 \text{ mm day}^{-1}$. The coupled minus uncoupled CWV difference (spatially smoothed) is con-540 toured (interval is 1 kg m^{-2} ; negative values are dashed and zero contour is omitted). In center 541 and right columns, frequency differences are plotted only where rainfall distributions are sta-542 tistically significant at the 95% confidence interval, as determined by a Kolmogorov-Smirnov 543 test. 544

The imprint of coupled feedbacks on convective moistening is considered for CGCM-545 minus-AGCM CWV differences. The difference in heavy rain frequency between CGCMs 546 and AGCMs (right column of Figure 9) and CWV differences are well-correlated ($r \ge$ 547 0.7 for the plotted domain). The close relationship between tropical rainfall and CWV 548 is well known (e.g., Bretherton et al., 2005; Thayer-Calder & Randall, 2009; Neena et 549 al., 2014), but it can be difficult to infer if convection is responsible for CWV patterns 550 or vice versa: convection that results in moderate-to-high rainfall rates moistens the at-551 mosphere, but a moist atmosphere favors development of more intense convection. Our 552 analysis, however, suggests that the response of convection to ocean feedbacks plays a 553 role in shaping the coupled minus uncoupled CWV differences. 554

555

6 Synthesis and Discussion

We have demonstrated that the common factor for improved MJO simulation skill 556 with coupling is the sharper equatorward moisture gradients across the Warm Pool, which 557 enhances tropospheric moistening by meridional moisture advection east of MJO con-558 vection. The sharper moisture gradients in CGCMs arise from coupled feedbacks that 559 enhance low-level convective moistening for rainfall rates greater than about 5 mm day $^{-1}$; 560 this enhancement is most frequently observed near the Equator. Next, we consider how 561 atmospheric and oceanic model physics jointly regulate tropospheric moistening, and whether 562 these interactions are consistent with those inferred from observation-based estimates. 563

564

6.1 Ocean feedbacks and convective activity

We first consider interactions among the ocean, convection, and free-tropospheric 565 moisture in MetUM-CPL and ECHAM-CPL. The cumulus parameterization in each model 566 employs a CAPE-based closure assumption: when temperature and moisture perturba-567 tions near the surface and aloft generate positive CAPE, convection is initiated to con-568 sume CAPE and maintain a neutrally stable environment. Because of weak tropical tem-569 perature gradients, CAPE in the tropics is highly sensitive to surface temperature per-570 turbations (Williams, 1994). Rules for convective initiation may also play role. For ex-571 ample, in MetUM, convection is initiated when the temperature profile at the lifted con-572 densation level becomes unstable, regardless of conditions aloft. In MetUM-CPL and ECHAM-573 CPL, compared to cold SSTAs, warm SSTAs generate larger CAPE, requiring more or 574 stronger convection to neutralize the instability. We hypothesize that this increased "de-575

-26-

576 577 mand" for deep convection above warm SSTAs is the cause for relatively stronger upperlevel moistening and weaker low-level moistening for warm SSTAs at all rainfall rates.

MetUM-CPL and ECHAM-CPL are both coupled to 1D ocean mixed layer mod-578 els with fine vertical resolution (≈ 1 m in the upper 10 m) and sub-daily coupling designed 579 to represent the large SST diurnal cycle during suppressed conditions (e.g., Kawai & Wada, 580 2007). The SST diurnal cycle rectifies onto subseasonal scales (Bernie et al., 2005; Shin-581 oda, 2005), so that the Warm Pool daily SSTA distributions in MetUM-CPL and ECHAM-582 CPL are positively skewed (Figure 10). The transition from convectively suppressed to 583 disturbed conditions in MetUM-CPL and ECHAM-CPL would coincide with rapid SST 584 warming in response to strong surface heating, reduced evaporative cooling (from reduced 585 wind speed), and ocean mixed layer shoaling (Figure 11). The rapid SST warming would 586 quickly generate CAPE and initiate convection, even though the mid-troposphere is typ-587 ically too dry to support deep convective moistening at these rainfall rates (Thayer-Calder 588 & Randall, 2009; Kim et al., 2009). As large-scale circulations moisten mid-levels via mois-589 ture advection, increasing MJO-associated surface wind anomalies and, potentially, con-590 vectively driven wind gusts will cool the upper ocean, so that by the time convection has 591 organized into a large-scale system with heavy rainfall, SSTs will have cooled. In these 592 cold SSTA conditions, plumes originating from the boundary layer will be less buoyant 593 and detrain moisture at lower levels than their warm SSTA counterparts, thereby en-594 hancing low-level moisture. 595

SPCAM3-CPL and CNRM-CPL differ from MetUM-CPL and ECHAM-CPL in that 601 neither employ CAPE-based closure assumptions in their cumulus parameterizations, 602 and both are coupled to 3D ocean models with coarse (≈ 10 m) upper-ocean vertical res-603 olution. Convection in SPCAM3-CPL is explicitly simulated with a continuously run-604 ning two-dimensional cloud-permitting model embedded in each GCM grid column. The 605 convection scheme in CNRM-CPL employs a moisture convergence closure assumption. 606 These treatments may reduce the sensitivity of convective initiation to SST perturba-607 tions, allowing CAPE to build up before convection is initiated, thus favoring the shift 608 toward higher rainfall rates during warm SSTA periods (Figure 7c). It is also possible 609 that the coarse upper-ocean vertical resolution in these models yields a smaller SST ten-610 dency than that in MetUM-CPL or ECHAM-CPL. This reduced warming rate could sub-611 tly shift the phasing of maximum SST with respect to convection so that it more closely 612 aligns with heavier rainfall rates. 613

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SSTA skewness, Nov-Apr

- ⁵⁹⁶ Figure 10. SSTA skewness, defined as the third moment of the SSTA distribution, for
- 597 CGCMs.



Figure 11. a) Mean SST as a function of rainfall rate for CGCMs and b) mean SST differ ences for CGCMs minus AGCMs (dots indicate differences are significant at the 95% confidence
 interval.

614

6.2 Coupled feedbacks to the convection-column humidity relationship

The above discussion implicitly invokes our understanding that rainfall intensity 615 is regulated by the mixing of cloud and environmental properties through detrainment 616 and entrainment (e.g., de Rooy et al., 2012). Because bulk mixing rates for these pro-617 cesses are difficult to measure, in GCMs they are often tuned to achieve realistic cloud-618 top height distributions, temperature and humidity profiles, or rainfall climatology. While 619 knowledge of entrainment processes in a GCM can help diagnose the spectrum of con-620 vective variability (DeMott et al., 2007; Thayer-Calder & Randall, 2009; Klingaman & 621 Woolnough, 2014), other factors, such as stability, moisture advection, and the interac-622 tions of cloud microphysical properties with convective heating also regulate the sensi-623 tivity of rainfall intensity to column humidity ([e.g., Klingaman et al., 2015). 624

To better understand the net effect of these interactions, we examined how rain-625 fall rate depends on the CWV saturation fraction (Bretherton et al., 2004; Neelin et al., 626 2009) in each GCM. In the tropics, the frequency of heavy rainfall increases non-linearly 627 with the column saturation fraction, defined as the ratio of column water vapor to sat-628 uration column water vapor. CGCM joint PDFs of CWV fraction and rainfall rate are 629 plotted with contours in Figure 12. The "precipitation uptick" (i.e., where rainfall rate 630 increases rapidly as a function of CWV fraction) in SPCAM3-CPL, MetUM-CPL, and 631 ECHAM-CPL occurs near the 0.8 CWV fraction, consistent with analysis of observa-632 tions (Bretherton et al., 2004; Neelin et al., 2009). For CNRM-CPL, the uptick is de-633 layed until 0.95 CWV fraction, suggesting that deep convection in that model is overly 634 sensitive to column moisture. 635

The effect of coupling on the convection-column humidity relationship is shown by 641 plotting the joint PDF difference for warm and cold SSTA periods (shading in the left 642 column of Figure 12). Compared to cold SSTA PDFs, warm SSTA PDFs in SPCAM3-643 CPL and CNRM-CPL are shifted toward lower saturation fractions, suggesting a reduced 644 critical saturation fraction for heavy rainfall for warm SSTA periods. In MetUM-CPL 645 and ECHAM-CPL, the warm SSTA PDFs shift toward lower saturation fractions at light 646 rainfall rates, but also shift away from heavy rainfall rates for most saturation fraction 647 bins (consistent with the rainfall rate PDF difference shown in Figure 7c). This suggests 648 that convection initiation in MetUM-CPL and ECHAM-CPL may be overly sensitive to 649 SSTAs, while rainfall intensity remains sensitive to saturation fraction as seen in the re-650

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Figure 12. November-April joint PDFs of daily column saturation fraction and daily mean rainfall rate for CGCMs (contours; see text for definition) and left) the joint PDF difference for warm-minus-cold SSTA periods (shading) and right) the joint PDF difference for CGCM minus AGCM simulations (shading). Ocean-only data are compiled over the Warm Pool (20°S-20°N; 30°E-240°E).

duced occurrence of high rainfall rates for warm SSTA periods. This is consistent with our argument that convection initiates too readily (i.e., in a too-dry environment) with warm SSTAs in these models, and the convective plumes quickly lose buoyancy as they entrain dry environmental air.

The PDF differences for coupled minus uncoupled simulations (Figure 12; right column) are difficult to interpret. At a minimum, they illustrate that the net effects of coupled feedbacks on the convection-column humidity relationship are highly model dependent. This sensitivity is likely rooted in large-scale circulation changes or other feedbacks beyond those directly related to coupling.

660

6.3 The role of surface fluxes

The atmosphere senses SST perturbations through their effects on surface fluxes. 661 These fluxes are sources of atmospheric boundary-layer buoyancy, which regulates the 662 depth of convective plumes originating from the boundary layer. Although the latent heat 663 flux is an order of magnitude larger than the sensible heat flux, its contribution to boundary-664 layer buoyancy (via the effect on water vapor density) is roughly comparable to that of 665 the sensible heat flux. For a given wind speed, the latent heat flux is primarily governed 666 by the specific humidity of near-surface air, while the sensible heat flux is more sensi-667 tive to the SST (DeMott et al., 2014; Yokoi et al., 2014). From the perspective of boundary-668 layer buoyancy budgets, SST perturbations are more directly communicated to the at-669 mosphere through the sensible heat flux (e.g., Yang, 2018). Both fluxes contribute to the 670 vertical ascent of convective plumes through converting boundary-layer available poten-671 tial energy to kinetic energy. Once the plumes reach their lifted condensation level, the 672 release of energy by condensation of oceanic water vapor increases plume buoyancy. Both 673 the boundary-layer buoyancy and the latent heat release from the condensation of pre-674 viously evaporated surface water will be greater for plumes originating over warm SSTAs 675 than over cold SSTAs. Warm SSTA plumes will rise farther, and detrain their moisture 676 higher, than cold SSTA plumes. 677

⁶⁷⁸ DeMott et al. (2016) and Gao, Klingaman, DeMott, and Hsu (2018) found that ob-⁶⁷⁹ served tropical intraseasonal SST perturbations to surface fluxes directly contribute up ⁶⁸⁰ to 1–2% of $\langle m \rangle$ and 10–20% of $\frac{\partial \langle m \rangle}{\partial t}$ across the Warm Pool. Our findings herein that cou-⁶⁸¹ pled surface fluxes increase the efficiency of convective moistening at high rainfall rates,

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and that this more efficient moistening occurs near the Equator, offer a process-level perspective of the SST *direct* effects reported in DeMott et al. (2016) and Gao et al. (2018). The more important feedback for MJO propagation, however, is the *indirect* effect of SST perturbations on MJO propagation through its regulation of the mean-state moisture distribution.

Another point worth discussing is the role of low-level moistening from vertical mois-687 ture advection driven by boundary-layer convergence. Our results clearly demonstrate 688 that equatorial boundary-layer moisture export does not uniformly increase with cou-689 pling. The decrease in boundary-layer moisture pumping with coupling seen in MetUM-690 CPL and ECHAM-CPL may be linked to the rapid decline in SSTA with heavy rain-691 fall in those models (Figure 11b), which could limit warm SST contributions to bound-692 ary layer convergence during convective build-up periods. For ECHAM-CPL, the "flat-693 tening" (vs sharpening) of the meridional moisture gradient in the western Pacific (Fig-694 ure 1k), and subsequent reduction of meridional moisture advection, may compound the 695 issue. SPCAM3-CPL and CNRM-CPL exhibit coupled moistening characteristics more 696 similar to those observed in ERAI, and they each exhibit increases in boundary layer mois-697 ture export with coupling. This provides a modicum of support for the idea that increased 698 boundary-layer moisture export is one reason why MJO propagation improves with cou-699 pling, but this topic requires further investigation using observations and coupled reanal-700 yses. Applying the methods developed herein to a larger collection of models with var-701 ious ocean model configurations could help clarify this issue. 702

703

6.4 Ocean coupling-mean state feedbacks in models and ERAI

The final point to consider is whether the ocean feedbacks to convective moisten-704 ing and mean-state moisture patterns in models are consistent with those in observations. 705 We address this point by revisiting Figure 8. Warm-minus-cold SSTA $-Q_2/L_v$ profiles 706 for SPCAM3-CPL and CNRM-CPL are most similar to those for ERAI (left column), 707 but also exhibit larger SSTA-weighted low-level moistening and drying differences (cen-708 ter and right columns). Low-level moistening differences in MetUM-CPL and ECHAM-709 CPL are even larger. Are the larger coupled feedbacks in models an indication that ocean 710 coupling acts a crutch for MJO similation? The muted coupled feedback signature in ERAI 711 in the right column of Figure 8 is largely a consequence of the rarity of high rainfall rates 712 diagnosed by ERAI which, as previously discussed, may be underestimated. Another con-713

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⁷¹⁴ sideration is that ERAI is an uncoupled data assimilation product, and not an obser-

vational product. While $-Q_2/L_v$ can be estimated from oceanic in situ observations col-

⁷¹⁶ lected with triangular or rectangular sounding arrays, these arrays are usually only avail-

⁷¹⁷ able during relatively short field campaigns and do not collect enough samples to allow

subsetting of $-Q_2/L_v$ by rainfall rate and SSTA. Repeating this analysis with other re-

analysis products would help characterize the sensitivity of SSTA-conditioned $-Q_2/L_v$

to the assimilating model.

721 7 Summary and Conclusions

The role of intraseasonal SST perturbations for MJO eastward propagation was 722 studied with 20–25 year coupled and uncoupled simulations of four different GCMs. Monthly 723 mean SST from each coupled model was prescribed to its respective uncoupled simula-724 tion to ensure identical SST mean state and low-frequency variability for each coupled-725 uncoupled simulation pair. As expected, coupling improved MJO eastward propagation 726 beyond the Maritime Continent in all four models, demonstrating the non-negligible role 727 of sub-monthly SST perturbations for MJO simulation fidelity. The challenge for under-728 standing this result requires reconciling the following: SST perturbations are commu-729 nicated to the atmosphere through their effects on surface fluxes, yet surface fluxes play 730 only a minor role in the maintenance and propagation of MJO convection. As in obser-731 vations, MSE budget analyses of the simulated MJO reveal that, to a first order, MJO 732 maintenance and propagation in all eight simulations are maintained by longwave ra-733 diative heating and horizontal moisture advection, respectively. 734

Further analysis revealed that, despite the identical SST climatology in each cou-735 pled and uncoupled simulation pair, coupling improves MJO simulation by uniformly sharp-736 ening mean state zonal and meridional moisture gradients to enhance advection of the 737 mean-state moisture by the anomalous wind. Improved MJO fidelity in the coupled sim-738 ulations is uniformly the result of sharper meridional moisture gradients driven by en-739 hanced equatorial CWV (or relatively smaller CWV reduction on the Equator, as in SPCAM3-740 CPL). The sharper CWV equatorward gradient yields enhanced moistening by poleward 741 flow east of MJO convection and extended eastward propagation of MJO convection com-742 pared to uncoupled simulations. 743

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Daily mean CWV averaged over all eight phases of the Wheeler-Hendon MJO RMM 744 index (Wheeler & Hendon, 2004) during MJO active periods (i.e., RMM amplitude > 1) 745 confirms that the MJO does not sharpen meridional CWV gradients. Mechanisms that 746 might explain the sharper meridional gradients in the CGCMs, such as enhanced Warm 747 Pool Hadley circulation or surface flux feedbacks to equatorial convection, are not uni-748 formly observed across models. Instead, SST perturbations in coupled simulations yield 749 different rainfall rate PDFs during warm and cold SSTA periods whose net effect is en-750 hanced low-level convective moistening at high rainfall rates. The intersection of regions 751 dominated by moderately high rainfall rates $(R > 5 \text{ mm day}^{-1})$ and SST anomaly pat-752 terns that favor enhanced low-level moistening in each model is found near the Equa-753 tor. Hence, the sharper moisture gradients that facilitate MJO propagation in coupled 754 simulations are a consequence of oceanic regulation of low-level moistening by unresolved 755 convective processes. 756

The changes in low-level moistening with ocean feedbacks are sensitive to param-757 eterized processes that regulate convective initiation and entrainment of environmental 758 air. In our study, the CAPE-based closure schemes in MetUM-CPL and ECHAM-CPL 759 yield *reduced* low-level moistening (i.e., relative drying) during warm SSTA periods at 760 all rainfall rates. In contrast, super-parameterized convection in SPCAM3-CPL and the 761 moisture-convergence closure assumption in the CRNM-CPL convective parameteriza-762 tion yield enhanced low-level moistening for all rainfall rates during warm SSTA peri-763 ods, which is consistent with the results from ERAI. These differences illustrate the de-764 pendence of GCM mean state moisture distributions on parameterized physics that lift 765 water vapor from the ocean surface and moisten the free troposphere (Randall, 2013). 766

We began this study to understand how SST-modulated surface fluxes improve MJO 767 simulation. We learned that coupling improves MJO propagation by sharpening merid-768 ional moisture gradients across the Warm Pool, thereby enhancing column moistening 769 by meridional moisture advection east of MJO heating. In an effort to understand why 770 coupling changes the mean-state moisture, we composited Warm Pool $-Q_2/L_v$ profiles 771 conditioned by rainfall rate and SSTA. This yielded unexpected new insights into how 772 parameterized convection and upper-ocean heating together influence tropical mean-state 773 moisture patterns. This framework is potentially useful for understanding a broader ar-774 ray of observed phenomena, such as large-scale convective aggregation and two-way feed-775 backs between convection and SST perturbations on interannual-to-decadal scales. It could 776

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also be leveraged as a process-oriented diagnostic to assess the fidelity of simulated ocean-

- atmosphere or land-atmosphere interactions in a hierarchy of model configurations, and
- to study causes of persistent model biases, such as the poor representation of tropical
- convectively coupled equatorial waves, SST mean-state cold biases, or the double ITCZ.

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