Pathways to Better Prediction of the MJO: 2. Impacts of Atmosphere-Ocean Coupling on the Upper Ocean and MJO Propagation

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Abstract This study investigates effects of atmosphere-ocean coupling on MJO precipitation and eastward propagation, and upper ocean conditions during and after MJO passage. To explore pathways for improving MJO prediction, three model experiments are conducted using the Unified Wave Interface-Coupled Model at convection-permitting (4 km) resolution: (a) uncoupled atmosphere-only, (b) coupled atmosphere-ocean, and (c) coupled atmosphere-ocean with improved air-sea flux algorithm simulations. The model simulations are compared with observations from the DYNAMO field campaign in 2011. Both coupled atmosphere-ocean simulations produced eastward propagation of the MJO where the uncoupled, atmosphere-only simulation did not. The uncoupled model overestimates both precipitation and surface winds associated with the MJO, while coupled model simulations substantially reduce model bias. Improved air-sea fluxes lead to systematic improvements in precipitation, winds, sea surface temperature, and the ocean mixed layer when compared to the original coupled simulation. This leads to further improvement of the MJO's eastward propagation speed compared with observations. Despite these improvements, the regional coupled simulations still have difficulty representing the extent of convectively suppressed conditions in the Indian Ocean after MJO passage, which indicates the importance of the large-scale environment from lateral boundary conditions. Coupled model simulations also reveal some issues in the representation of upper ocean stratification in the ocean model, especially errors in salinity, which result in overestimation of the mixed layer depth after MJO passage.

Plain Language Summary Although the Madden-Julian Oscillation (MJO) has been recognized as a coupled atmosphere-ocean phenomenon in some previous studies, systematic investigations of the air-sea interaction processes and how they affect the MJO prediction are still lacking. This study focuses on better understanding of the atmosphere-ocean coupling and their impacts on two key characteristics of the MJO: precipitation and eastward propagation. A novelty of this study is that we provide quantitative measure of the impacts by comparing coupled atmosphere-ocean simulations with uncoupled atmosphere-only simulation against observations from the DYNAMO field campaign in 2011. All model simulations are conducted using a cloud-permitting high resolution (4 km grid spacing) that can better represent moist physics and atmospheric boundary layer processes as a baseline. In the atmosphere-only simulation with sea surface temperature held constant, the model produced excessive, nearly stationary large-scale precipitation that persisted and failed to propagate eastward and leave the Indian Ocean (IO). In contrast, the MJO-induced upper ocean cooling in the coupled atmosphere-ocean model helped dissipate convection over the IO and propagate the MJO eastward. An improved air-sea flux algorithm helped reduce model biases in air-sea fluxes, precipitation, winds, and upper ocean temperature, further improving the MJO's eastward propagation.

1. Introduction
The Madden-Julian Oscillation (MJO) is an atmosphere-ocean coupled phenomenon characterized by an alternating pattern of enhanced and suppressed precipitation initiating in the Indian Ocean (IO) and propagating eastward across the Maritime Continent (MC) and western Pacific (Madden & Julian, 1971, 1972; Zhang, 2005). The precipitation pattern is accompanied by changes in surface winds and upper ocean temperature. Although it is the leading source of tropical intraseasonal variability, its prediction remains a major challenge in numerical weather prediction (Hung et al., 2013; Ling et al., 2014; Zhang, 2005, 2013). Some studies show that the inclusion of atmosphere-ocean interaction improves the prediction of the MJO, especially its eastward propagation (DeMott et al., 2015; Klingaman & Woolnough, 2014; Wang et al., 2015; Zhang et al., 2006).
Observational studies of the MJO using satellite data and observations from two major field campaigns, the Tropical Ocean-Global Atmosphere—Coupled Ocean-Atmosphere Response Experiment (COARE) in 1992–1993 (Webster & Lukas, 1992) and the Dynamics of the MJO in 2011–2012 (Yoneyama et al., 2013), have led to better understanding of the MJO. They have shown that the MJO convection and atmospheric boundary layer have multiscale variability (e.g., Chen et al., 1996; Johnson & Ciesielski, 2017) and air-sea interaction occurs from the diurnal to intraseasonal time scales (Chen & Houze, 1997; Chen et al., 2016; de Szoeke et al., 2015; Fairall et al., 2003; Feng et al., 1998; Shinoda & Hendon, 1998; Weller & Anderson, 1996). During the convectively suppressed periods of the MJO, increased insolation and weak winds lead to a warm and stably stratified ocean mixed layer (Anderson et al., 1996). During the convectively active periods of the MJO, intense convection, precipitation and strong winds develop, deepening the oceanic mixed layer and cooling the upper ocean (de Szoeke et al., 2015; Moun et al., 2014) while providing moisture to sustain atmospheric precipitation.

Numerical modeling studies have revealed various aspects of the MJO convection and air-sea interaction and their impacts on the MJO. Models representing convection using super-parameterization in global models (Benedict & Randall, 2009) or explicitly using cloud-permitting resolution (e.g., Savarin & Chen, 2022) can improve model simulations of the MJO. So does incorporating prescribed varying sea surface temperature (SST; Miura & Randall, 2004), their impacts on the MJO. Models representing convection using super-parameterization in global models (Benedict & Randall, 2011), or a fully coupled dynamic ocean model (DeMott et al., 2014; Shinoda et al., 2013). Some early studies used a fixed-depth mixed layer model to incorporate intraseasonal SST anomalies on top of a seasonal cycle and found that even intraseasonal temperature changes of 0.1°C–0.15°C improved MJO variability and propagation (Waliser et al., 1999). In simple mixed-layer models, the SST cooling is mostly determined by the thickness of the slab ocean, with shallower mixed layers producing more SST variability, and thicker mixed layers producing less (Marshall et al., 2008). Observed SST variations induced by the MJO can exceed 1°C (Lau & Sui, 1997). Over the course of an MJO event, wind mixing from strong surface westerlies can rapidly deepen the ocean mixed layer, and entrainment of cooler water from below can further modify the mixed layer and SSTs (Lucas & Lindstrom, 1991), making the case for use of dynamic ocean models.

Air-sea interaction can have a significant influence on atmospheric moisture and convection associated with the MJO. Marshall et al. (2008) found that the inclusion of atmosphere-ocean coupling with a slab ocean model can improve MJO representation through enhancing the moisture-convergence feedback already present in the atmosphere. The additional variability provided by ocean models results in increased SST ahead of MJO precipitation, where clear skies permit warming, and weak winds induce surface heat fluxes, warming and moistening the atmospheric boundary layer. This leads to increased moistening in the atmosphere, convergence, and results in increased surface westerlies and precipitation to the east of MJO convection, promoting eastward propagation of the MJO. The SST cooling induced by the MJO due to increased cloudiness, strong surface winds, and high air-sea fluxes can help shut down MJO convection to the west. This also explains why models that rely on convective parameterizations (CPs) tend to produce better MJO events when their convection is triggered by moisture convergence (e.g., Liu et al., 2022; Savarin & Chen, 2022). A study by Maloney and Sobel (2004) showed that by eliminating the MJO-induced surface heat flux variability, MJO precipitation becomes decoupled from the surface wind signal. Thus, an accurate representation of air-sea heat fluxes is important for accurate simulations of MJO propagation and interactions between its precipitation and circulation. But air-sea fluxes are hard to represent in models, and even reanalysis products contain significant biases when compared to in-situ observations, such as those measured during the Dynamics of the MJO field campaign (Gao et al., 2016).

Studies of upper-ocean processes and variability associated with the MJO are limited by observations, in part due to the difficulty of observing upper ocean structures from satellites and in situ observations at sparse mooring locations. Nevertheless, there are a number of studies that delve deeper than the surface properties of the ocean, such as SST, currents, and air-sea fluxes, for example, from the RAMA mooring array in the IO (McPhaden et al., 2009) and ship observations during field campaigns (e.g., Moun et al., 2016). Studies using observations and models show that enhanced zonal advection, vertical mixing, and formation of barrier layers in the upper ocean can significantly affect the near-surface temperature and SST evolution (Halkides et al., 2015; Jensen et al., 2015; McPhaden & Foltz, 2013). In addition to understanding ocean processes alone, Shinoda et al. (2013) find that the enhancement of eastward upper-ocean currents by the MJO is strongly dependent on the strength of the surface westerly winds; lower-resolution simulations produce weaker winds, weaker upper-ocean currents and weaker horizontal advection during active MJO. For improved forecasting of MJO events without prior
knowledge of SST variability, it is imperative to better understand, and accurately represent processes in the upper ocean that can help cool or warm SST in different phases of the MJO.

In this study, we investigate impacts of the atmosphere-ocean coupling on the MJO prediction using the Unified Wave Interface-Coupled Model (UWIN-CM) at convection-permitting (4 km) resolution. We compare three model experiments: (a) uncoupled atmosphere-only, (b) coupled atmosphere-ocean, and (c) coupled atmosphere-ocean with improved air-sea flux algorithm using observations from DYNAMO. We analyze the response of MJO precipitation and its eastward propagation, as well as post-MJO ocean recovery, to the representation and biases in air-sea heat fluxes. In Section 2, we describe the atmosphere-ocean coupled model used for the study, the configuration of experiments, the method for identifying and tracking the MJO, and the data used for model evaluation. The large-scale MJO precipitation, surface wind structure, and SST cooling are described in Section 3. Section 4 focuses on the analysis of air-sea fluxes and their biases over the equatorial IO, and how local air-sea flux differences can lead to basin-wide changes in the MJO. Section 5 focuses on the upper-ocean evolution during the MJO event, and the processes through which surface conditions affect SST cooling induced by the MJO. Summary and discussion of the results are presented in Section 6.

2. Model and Data

2.1. Model Configuration

The numerical model used in this study is the UWIN-CM (Chen & Curcic, 2016; Chen et al., 2013), which allows for interactive coupling between the atmosphere and ocean circulation models. The coupled model and configuration are the same as described in Savarin and Chen (2022).

The atmosphere component of UWIN-CM is the Weather Research and Forecasting (WRF) model v3.6.1 with the Advanced Research WRF dynamical core (Skamarock & Klemp, 2008). WRF is a non-hydrostatic atmospheric model that we configure with 44 terrain-following levels in the vertical, and three nested model grids shown Figure 1. The outermost domain (D01) ranges from 15.4° to 174.6°E in longitude, and from 32.0°S to 32.0°N in latitude, with a grid spacing of 36 km. Two inner nests are centered on the equator, covering the areas from 36.0° to 165.2°E and 15.3°S to 15.3°N at a grid spacing of 12 km (D02), and from 54.0° to 154.6°E and 10.6°S to 10.6°N at a grid spacing of 4 km (D03). The highest-resolution domain is convection permitting, with explicit moist physics and no CP, while D01 and D02 use CP. The surface layer parameterization is based on the Monin-Obukhov similarity theory. The atmospheric boundary layer parameterization is the Yonsei University scheme (Hong et al., 2006). The cloud microphysics parameterization used in all three domains is the single-moment five-species scheme in WRF (WSM5; Hong et al., 2004). Initial and boundary conditions of UWIN-CM come from the European Centre for Medium-Range Weather Forecast (ECMWF) operational forecast fields as described in Kerns and Chen (2014), with boundary conditions updated every 12 hr.

The ocean circulation component of UWIN-CM is the HYbrid Coordinate Ocean Model (HYCOM) v2.2.99 (Metzger et al., 2014). HYCOM is a hydrostatic ocean circulation model with a vertical coordinate system that transitions between layers of constant density in deep water, layers of constant depth (z-layers) in shallow water and near the ocean surface, and terrain-following layers in intermediate waters. HYCOM is configured with a single domain with uniform grid spacing of 0.08°, ranging from 9.8 km at the equator to 7.7 km at the domain's northern and southern boundaries, and spans from 30.0° to 172.0°E and from 33.0°S to 33.0°N. Out of 41 vertical layers, 14 are z-layers, 5 of which are located within 15 m of the surface (centered at depths of 0.5, 1.88, 4.33, 8.16, and 12.79 m). Initial and boundary conditions come from daily mean fields of the global HYCOM analysis (Cummings, 2005; Cummings & Smedstad, 2013).

Figure 1 shows the UWIN-CM nested model grids and the HYCOM SST analysis at the model initial time.
2.2. Air-Sea Flux Algorithm

To investigate the effects of the air-sea fluxes on the coupled model simulation of the MJO, we implemented a modification to the parameterization of air-sea latent and sensible heat fluxes in the WRF's surface layer parameterization. The air-sea fluxes in WRF are calculated using the bulk air-sea flux equations (Jiménez et al., 2012):

\[
\text{LHF} = L_e \rho C_h U (q_e - q_a) \tag{1}
\]
\[
\text{SHF} = \rho C_v U (\theta_e - \theta_a) \tag{2}
\]

LHF and SHF are latent and sensible heat fluxes, respectively. \(C_h\) and \(C_v\) are the dimensionless bulk transfer coefficients for moisture and heat (Stull, 1988), \(\rho\) is the surface layer density, \(M\) is the soil moisture availability, and \(c_v\) is the specific heat capacity at constant pressure. \(q_e\) and \(q_a\) are the saturated ground surface specific humidity and the specific humidity of the surface layer, \(\theta_e\) and \(\theta_a\) are the ground surface and surface layer potential temperature. \(U\) represents the wind speed in the surface layer, which is a combination of three factors:

\[
U = \sqrt{U_s^2 + U_{sg}^2 + U_{cv}^2} \tag{3}
\]

\(U_s\) is the surface wind speed, \(U_{sg}\) is the constant sub-grid velocity (0.61, 0.43, and 0.30 m s\(^{-1}\) in the 36-, 12-, and 4-km domains, respectively) (Mahrt & Sun, 1995). \(U_{cv}\) is the convective velocity, which is parameterized based on Beljaars, (1995) as:

\[
U_{cv} = \begin{cases} 
\sqrt{\theta_v - \theta_e}, & \theta_v > \theta_e \\
0, & \text{otherwise}
\end{cases} \tag{4}
\]

\(\theta_v\) and \(\theta_e\) are the virtual potential temperatures of the sea surface and surface layer, respectively. When the vertical gradient of \(\theta_e\) is negative across the air-sea interface, a positive convective velocity due to surface buoyancy is added to the surface layer wind speed.

The data for the original parameterization was collected over land-locked Oklahoma, USA, during the Boundary Layer interaction experiment (BLX83; Beljaars, 1995). Most common \(U_{cv}\) magnitudes fall between 0.5 and 4 m s\(^{-1}\), though they can reach up to 10 m s\(^{-1}\). There are times at which \(U_{cv}\) is higher than \(U_s\), and the contribution of \(U_{cv}\) to air-sea fluxes is substantial, but this mainly occurs at \(U_s < 5\) m s\(^{-1}\). Over ocean, \(U_{cv}\) is likely smaller that over land, and we modified Equation (4) by reducing \(U_{cv}\) to half if its original magnitude (0.5 × \(U_{cv}\)). The other components of surface layer winds (Equation 3) remain unchanged, and it is important to note that this change is not reflected in the model output winds—it only exists within the surface layer parameterization for the purpose of calculating air-sea fluxes. Given identical environments, this modification only results in reduced air-sea latent and sensible heat fluxes.

2.3. Model Experiments

All experiments include a high-resolution (4-km grid spacing) inner domain in which convection is explicitly resolved by the microphysical parameterization, while the Tiedtke CP (Tiedtke, 1989) is used in the two outer domains in addition to microphysics. This convection-permitting 4-km resolution is shown to best represent the magnitudes fall between 0.5 and 4 m s\(^{-1}\), though they can reach up to 10 m s\(^{-1}\). There are times at which \(U_{cv}\) is higher than \(U_s\), and the contribution of \(U_{cv}\) to air-sea fluxes is substantial, but this mainly occurs at \(U_s < 5\) m s\(^{-1}\). Over ocean, \(U_{cv}\) is likely smaller that over land, and we modified Equation (4) by reducing \(U_{cv}\) to half if its original magnitude (0.5 × \(U_{cv}\)). The other components of surface layer winds (Equation 3) remain unchanged, and it is important to note that this change is not reflected in the model output winds—it only exists within the surface layer parameterization for the purpose of calculating air-sea fluxes. Given identical environments, this modification only results in reduced air-sea latent and sensible heat fluxes.

To investigate the effects of atmosphere-ocean coupling and air-sea fluxes on the MJO simulations, we conduct three model experiments. The first model experiment is an uncoupled, atmosphere-only simulation (abbreviated as UA4). The second experiment is the coupled atmosphere-ocean model simulation (AO4-CTRL). In the third experiment, we use the modified air-sea flux algorithm described in Section 2.2, which is abbreviated as AO4-FLX. The initial SST is identical in all simulations (Figure 1), but remains constant throughout the simulation in UA4, while it evolves in the coupled AO4-CTRL and AO4-FLX simulations.

All simulations are initiated at 00 UTC on 22 November 2011, and span 15 days, capturing the time frame of the second MJO event observed during the intense observation period of DYNAMO. At the initial time, observations
show that convection is beginning to organize over the equatorial IO, while by the end of the period, the MJO has propagated over the eastern MC.

2.4. Ocean Mixed Layer

The method of calculating ocean mixed layer depth and ocean barrier layers is based on the variable density criterion introduced by Sprintall and Tomczak (1992), with minor modifications. The barrier layer is defined when the ocean mixed layer depth is smaller than the isothermal layer depth, and it is the ocean layer encapsulated between them. When the isothermal layer is shallower than the mixed layer, a barrier layer does not exist. To determine mixed layer depth and barrier layer thickness, we first designate a reference depth of 10 m to remove effects of diurnal variability. After this, a temperature threshold of 0.2°C is chosen for determining mixed layer properties. A temperature difference of 0.2°C from the temperature at reference depth defines the depth of the isothermal layer. Then, the coefficient of thermal expansion is calculated for that temperature difference, based on the TEOS-10 definition (the Python package gsw v.3.3.1 is used for the calculation). The coefficient of thermal expansion and the temperature threshold are then used to calculate the density threshold corresponding to the water column properties. The depth at which this density threshold is met is the ocean mixed layer depth. The barrier layer thickness is the difference in depths of the ocean mixed layer and the isothermal layer (when greater than zero).

2.5. MJO Tracking

The large-scale precipitation tracking algorithm (LPT), developed by Kerns and Chen (2016, 2020), is used to track the precipitation envelope associated with the MJO. The algorithm tracks a spatially smoothed 3-day rainfall accumulation that exceeds a chosen threshold over an area larger than $3 \times 10^5$ km$^2$. In the studies by Kerns and Chen, the rainfall accumulation threshold of 12 mm day$^{-1}$ was found to accurately capture MJO events during DYNAMO. In this study, a threshold of 15 mm day$^{-1}$ is used instead, as model simulations tend to over-produce precipitation, compared to Tropical Rainfall Measuring Mission (TRMM) 3B42 observations. After precipitation is tracked, additional constraints are introduced to separate MJO precipitation from other systems. These constraints ensure that the LPT-tracked feature exhibits consistent eastward propagation and persists for at least 7 days to eliminate synoptic scale features. Using LPT to track MJO precipitation provides a major advantage over traditional MJO indices—it allows us to visualize both the spatial and temporal evolution of MJO precipitation.

2.6. Observational Data

The simulated MJO event occurred in late November and early December 2011, during the intense observing period of DYNAMO, while many observational platforms were operating simultaneously. This provides a unique opportunity to evaluate the model from different perspectives—including ship measurements and moorings, as well satellite data to provide large-scale context.

The data used in this study includes:

1. In-situ observations:
   1.1 Air sea fluxes at R/V Revelle (at 80.50°E, 0.10°N between 22 November and 1 December 2011 available every minute; R in Figure 1)—calculating using the COARE 3.0 bulk flux algorithm (Fairall et al., 2003);
   1.2 Upper-ocean temperature and salinity from DYNAMO moorings (at 79°E, 0°N, and 79°E, 1.5°S, hourly), and upper-ocean and surface observations from an equatorial RAMA mooring (80.5°E, 0°N, hourly; McPhaden et al., 2009).
2. Satellite products:
   2.1 Precipitation: TRMM 3B42, available 3-hourly at 0.25° resolution (Huffman et al., 2007) and the Global Precipitation Measurement (GPM) Integrated Multi-Satellite E Retrievals for GPM v6B, available half-hourly at 0.1° resolution (Huffman et al., 2019);
   2.2 Surface winds: Cross-Calibrated Multi-Platform (CCMP), available 6-hourly at 0.25° resolution (Atlas et al., 2011);
2.3 **SST**: Global High Resolution Sea Surface Temperature Level 4 MUR Global Foundation Sea Surface Temperature Analysis, available daily at 9 km resolution (Chin et al., 2017);

2.4 **Air-sea fluxes**: Objectively Analyzed Air-Sea Fluxes (OAFlux), available daily at 1° resolution—using the COARE bulk flux algorithm 3.0 (Yu et al., 2008).

3. **Model initial and boundary conditions:**

   3.1 ECMWF operational analysis for the atmosphere component (updated 12-hourly at 0.25° resolution), and daily mean HYCOM global analysis (updated daily at 0.08° resolution).

### 3. MJO Eastward Propagation and Its Induced Upper Ocean Cooling

The traditional MJO indices such as the Real-Time Multivariate MJO Index (Wheeler & Hendon, 2004) cannot be used directly in regional model analysis. Time-longitude diagrams are a useful way of examining large-scale characteristics of MJO precipitation and wind patterns. Figure 2 shows Hovmöller diagrams of rainfall rate (left), surface zonal wind (middle), and SST (right), averaged between 5°S and 5°N, for observations and model simulations. Observations are shown in the top row (Figures 2a, 2e and 2i), using TRMM 3B42 precipitation, CCMP...

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**Figure 2.** Hovmöller diagrams of rainfall rate averaged over 5°S–5°N (left), surface zonal wind (middle), and SST (right) in observations (top, from left to right: Tropical Rainfall Measuring Mission 3B42, Cross-Calibrated Multi-Platform, and Global High Resolution Sea Surface Temperature) and model simulations as follows: UA4, AO4-CTRL, and AO4-FLX. Brown shading on bottom panels shows the height of maximum topography in the averaging region, separating the longitudes of the Indian Ocean from those of the Maritime Continent (black vertical line at 97°E).
surface zonal winds, and OISST. The eastward propagation of MJO precipitation occurs in two distinct bursts which are part of the same convective envelope but are separated by local suppression near the equator. The leading edge of precipitation is accompanied by zonal surface wind convergence, with westerly winds extending across the IO to the west, and easterly winds to the east. The westerly winds over the equatorial IO persist, but weaken, after active MJO convection has propagated eastward into the MC, and conditions over the IO are convectively suppressed. The SST in satellite observations is reduced by up to 1°C, then begins to recover after the MJO propagates out of the IO. This cooling range agrees with observations from R/V Revelle, a ship that was stationed at 80.5°E, 0.1°N for the first 9 days of the modeled period (marked with R in Figure 1), and DYNAMO and RAMA mooring observations (Figure 13).

Compared to observations, all UWIN-CM simulations overproduce precipitation (Figures 2a–2d), but there is a clear distinction between the uncoupled (UA4) and coupled (AO4) simulations. UA4 produces intense precipitation, which is present throughout the simulation over the IO and, to a lesser extent, over the western MC. The precipitation over the IO does not propagate eastward, while AO4 simulations show a clear eastward tendency in both precipitation and surface zonal winds, with some precipitation suppression over the IO in the last days of the simulations. All simulations reproduce broad regions of equatorial westerlies throughout the IO to the west of MJO precipitation. The strongest westerlies in UA4 remain confined to the IO, and in AO4 simulations, the leading edge propagates eastward across the MC, alongside the leading edge of precipitation. Surface westerly winds in AO4 simulations are weaker than in UA4, and weaker in AO4-FLX than in AO4-CTRL. AO4-FLX winds show least bias when compared to satellite observations.

Coupled simulations also produce significant ocean cooling in the IO and western MC. AO4-FLX SST cooling is weaker compared to AO4-CTRL, but AO4-CTRL cools too much compared to observations to begin with, so AO4-FLX shows improvement. The IO basin-wide SST change from the beginning to the end of the simulation, averaged within 5° of the equator is −1.13°C in AO4-CTRL and −0.91°C in AO4-FLX. For the same 15 days, the satellite-derived SST change is −0.47°C, but that includes the period after MJO passage, when the observed SST begins to recover. At the time when the IO-basin SSTs were lowest (30 November), the observed SST change from initial conditions was −0.58°C. In UA4, SST remains constant and identical to the initial condition depicted in Figure 1, which is on average 0.5°C warmer than observations in the equatorial region. The model SSTs in AO4 simulations level out by the end of the simulations, but there is little sign of post-MJO SST recovery. Compared to observations, the reduction in air-sea fluxes in AO4-FLX improves on the large-scale features of the MJO produced in AO4-CTRL—such as the high bias in precipitation and surface winds, and the exaggerated SST cooling.

The high precipitation bias in model precipitation is evident in LPT tracking shown in Figure 3 as is the MJO’s eastward propagation. The time- and space-evolution of large-scale precipitation is shown for TRMM (top) and UWIN-CM simulations, with colored contours outlining the area over which the mean 3-day precipitation exceeds 15 mm day$^{-1}$. Colors represent the MJO convective envelope through time. As a 3-day accumulation is required for tracking, the observed track before 25 November 2011 is shown in gray, identifying the observed MJO’s initiation area near 60°E and 7°S. The eastward propagation of the MJO in TRMM is smooth, with the precipitating area periodically alternating between the Northern and Southern hemispheres, with a mean eastward propagation speed of 4.76 m s$^{-1}$. The modeled MJOs are slow compared to observations. The eastward propagation of the
MJO in AO4 simulations is clear, at 4.03 m s\(^{-1}\) in AO4-FLX, and 3.93 m s\(^{-1}\) in AO4-CTRL. In UA4, large-scale precipitation over the IO never ceases (Figure 3b), and the apparent eastward propagation of the MJO centroid (0.72 m s\(^{-1}\)) comes from the eastward expansion of the MJO convection between 30 November and 1 December, and not from the propagation of precipitation away from the IO.

Unlike in Savarin and Chen (2022), all simulations produce trackable large-scale precipitation that lasts through the entire simulation, with some characteristics summarized in Figure 4. The LPT centroid in UA4 remains in the IO for the entire duration of the simulation, while AO4 simulations propagate eastward relatively smoothly, producing smaller positional errors. The inclusion of air-sea coupling (UA4 vs. AO4-CTRL) reduces the MJO area and integrated precipitation by nearly a factor of two, and an additional similar improvement is produced when comparing AO4-CTRL to AO4-FLX (Figure 4b). We see that relatively small changes in the air-sea flux parameterization can reflect significantly on the MJO, both in terms of the amount of produced precipitation, and its eastward propagation. Both AO4 MJO centroids propagate out the IO on 1 December, slightly before the observed MJO does, but the MJO centroid in AO4-CTRL hangs at the western edge of the MC for about 3 days (with precipitation extending into the IO), while the centroid in AO4-FLX propagates eastward more smoothly, and with fewer discrete jumps (Figure 4c).

Within the MJO, SST cools on average by \(~0.5^\circ\)C, which is well-matched in AO4 simulations (Figure 5). To construct this estimation, we look at time series of SST at each point inside the MJO LPT (Figure 3) over the IO and combine the time series of temperature difference from when a point first enters the LPT (time 0), until the
LPT—and the MJO influence—move away. Figure 5 shows the mean (solid), and the 10th and 90th percentiles (bottom and top dashed lines, respectively) of the change in SST as a function of time spent inside the MJO. Based on this, we can estimate that the observed MJO-induced SST cooling after 9 days under an MJO can be as strong as 0.9°C, or as weak as 0.1°C. There are regions where SST increases during active MJO, but they are mainly located on the edges of the MJO convective envelope.

Overall, the modeled distributions match well with the observations in reproducing MJO-induced SST cooling, with stronger SST decrease in AO4-CTRL than in AO4-FLX. At day 9, both the 10th percentile and the mean of the modeled SST cooling agree remarkably well with observations, especially in AO4-FLX. The modeled changes in the top and bottom 10% of the SST changes are more gradual compared to observations, which show both rapid warming and cooling (±0.3°C) within the first day inside the MJO convective envelope. Some of this could be related to the daily resolution of observational data, and to the fact that the observed MJO area is much smaller than in the models, so there are fewer time series to create the distribution from. This can also be inferred because the observed distribution exists for up to 9 days inside the MJO convective envelope, while the modeled distributions contain locations that remain inside the MJO convective envelope for up to 12 days.

In summary, all UWIN-CM simulations at convection-permitting resolutions produce large-scale equatorial convection and strong surface westerly winds. Only coupled simulations produce significant eastward propagation; the uncoupled simulation constantly produces and reinforces convection over the IO. Between coupled simulations, the eastward propagation is better defined in AO4-FLX, as the amount of precipitation and the size of the MJO convective envelope are less exaggerated. This implies that air-sea interaction provides important insights into the MJO's eastward propagation over the IO. Upper-ocean cooling induced by the MJO (which is absent in UA4) acts as a push for the MJO's eastward propagation, and a smaller bias in air-sea fluxes (corrected in AO4-FLX) reduces the amount of precipitation and speeds up the MJO's eastward propagation over the IO and into the MC.

**Figure 5.** The range of Madden-Julian Oscillation-induced sea surface temperature cooling as a function of time spent inside large-scale tracked precipitation. The solid lines show the median values in for observed SST (black) and AO4-CTRL and AO4-FLX simulations (colors), while the dashed lines denote the 10th and 90th percentiles.
4. Air-Sea Fluxes, Surface Winds, and Precipitation

The atmosphere and ocean interact through air-sea fluxes of momentum, trace elements such as sea salt, and sensible and latent heat. Of those, SHF and LHF are important sources of heat and moisture for the atmosphere, respectively, and a misrepresentation of their magnitude is partly responsible for the high precipitation bias in UWIN-CM simulations. The AO4-FLX experiment was designed with the awareness that the LHF in UA4 and AO4-CTRL were biased toward high values. Figure 6 shows the distributions of LHF (top) and SHF (bottom) in observations (gray) and model simulations (colored), as a function of the 10-m wind speed. Observations come from R/V Revelle (R in Figure 1), which are available every 10 min (gray circles), and the thick black line represents the mean flux value at a given wind speed. To compensate for the model’s less frequent output and possible misplacement of features, the model distributions include all points within 1° of R/V Revelle, after which each distribution is normalized by its maximum value. Orange squares show the daily flux values from the OAFlux product at the location R/V Revelle, which are in good agreement with ship observations.

The observed LHF ranges from 20 to 400 W m\(^{-2}\), with a relatively linear increase with surface wind, while the SHF trend is less linear and ranges from −5 to 110 W m\(^{-2}\). The mean air-sea fluxes and their biases compared to R/V Revelle observations are listed in Table 1. The mean LHF in UA4 is double that of the observed values at R/V Revelle, with the mean bias of 73.63 W m\(^{-2}\), and the mean of the distribution lying above even the highest observed fluxes for a given wind speed (Figure 6a). With air-sea coupling, the entire SHF and LHF distributions are shifted toward lower values, and the LHF bias is reduced by over 60% in AO4-CTRL, with an additional reduction of over 30% in AO4-FLX. The range of air-sea flux values in AO4-FLX is reduced (we get tighter distributions), but LHF still exhibits a high bias, and the lowest observed fluxes at a given wind speed are not reproduced.

| Product      | LHF (W m\(^{-2}\)) | LHF bias (W m\(^{-2}\)) | LHF bias (%) | SHF (W m\(^{-2}\)) | SHF bias (W m\(^{-2}\)) | SHF bias (%) |
|--------------|---------------------|--------------------------|--------------|---------------------|--------------------------|--------------|
| UA4          | 253.74              | 128.84                   | 103          | 30.55               | 16.45                    | 117          |
| AO4-CTRL     | 174.92              | 50.02                    | 40           | 17.26               | 3.16                     | 22           |
| AO4-FLX      | 158.32              | 33.42                    | 27           | 15.90               | 1.80                     | 13           |
| R/V Revelle  | 124.90              | –                        | –            | 14.10               | –                        | –            |

Table 1: Mean Air-Sea Heat Fluxes and Their Biases as a Function of Wind Speed Compared to Measurements From R/V Revelle

Figure 6. Distributions of observed air-sea latent (top) and sensible (bottom) heat flux at R/V Revelle (80.5 E, 0.10 N, gray circles). Model simulated air-sea fluxes (colored contours) come from within 1° of R/V Revelle and are normalized by the highest value in each distribution. The contour levels plotted are at the 1st, 10th, 30th, 50th, and 70th percentiles of the distributions. Yellow squares show the daily OAFlux fluxes at the location of R/V Revelle.
As OAFlux values agree with the ship observations, we use the product to compare air-sea flux distributions on a larger scale—over the entire equatorial IO, a region outlined in black in Figures 3, 55°–97°E and 10°S–10°N. The SHF and LHF distributions in model and OAFlux are shown in Figure 7, along with the distribution of wind speed over the same region (observations from CCMP). Comparison over the larger region allows us to generalize the results from R/V Revelle data. The mean values and biases of wind speed, SHF, and LHF distributions over the equatorial IO are recorded in Table 2.

The LHF biases at R/V Revelle are higher than over the entire equatorial IO, which is reasonable considering that the ship is located right at the equator and experiences the strongest MJO winds, while the equatorial IO fluxes from OAFlux encompass a significant area not in contact with the MJO convective envelope tracked by LPT (Figure 3). Again, we see the strongest improvement from UA4 to AO4-CTRL (45% in LHF), with an additional smaller reduction in LHF bias in AO4-FLX. The biases in SHF are improved as well, though the SHF magnitude is small to begin with. The coupled AO4-CTRL and AO4-FLX improve the surface wind and fluxes, especially in higher wind speed conditions, compared with observations (Figure 7) similar to that documented in Chen et al. (2001).

The improvements in air-sea flux representation (specifically the LHF) from AO4-CTRL to AO4-FLX are larger than what was expected from a small change to the flux calculation parameterization (Equation 4). The relationship between LHF, surface winds, and precipitation is shown in time series of IO-averaged quantities in Figure 8. In all panels, observations or analysis data are shown in black, and model simulations in color; thick lines show data smoothed using a 24-hr running mean. The reduction in LHF between UA4, AO4-CTRL, and AO4-FLX is clear from the beginning of the simulation, while the differences in surface winds and precipitation take some time to develop. This could be related to the upscaling of the change in local air-sea fluxes to the larger basin scale. We see that a reduction in air-sea fluxes results in an overall weakening of surface winds, as well as reduced precipitation over the entire equatorial IO. The impact of LHF on precipitation is relatively straightforward, as reduced LHF means there is less moisture in the atmosphere that is available to precipitate. But the surface winds are affected indirectly, most likely through a CISK-like mechanism, by way of which weaker convection will induce a weaker circulation.

From Figure 8 we can see that the unexpectedly large reduction in LHF (and LHF bias) in AO4-FLX is initially due to the implemented reduction in flux magnitude, but that it is reinforced by weaker surface winds—which will additionally lower air-sea fluxes. However, there are also physical mechanisms that counteract this reinforcement. Specifically, reduced air-sea fluxes and weaker winds lead to less upper-ocean cooling in AO4-FLX compared to the other models.

### Table 2

| Product   | $U_*$ (m s$^{-1}$) | $U_*$ bias (m s$^{-1}$) | LHF (W m$^{-2}$) | LHF bias (W m$^{-2}$) (%) | SHF (W m$^{-2}$) | SHF bias (W m$^{-2}$) (%) |
|-----------|------------------|-------------------------|------------------|---------------------------|----------------|----------------------------|
| UA4       | 7.61             | 2.20                    | 188.71           | 77.19                     | 18.35          | 10.09                     | 122                        |
| AO4-CTRL  | 6.74             | 1.33                    | 138.08           | 26.56                     | 4.20           | 51                        |
| AO4-FLX   | 6.54             | 1.13                    | 127.19           | 15.67                     | 3.62           | 44                        |
| OAFlux    |                  |                         | 111.52           | 8.26                      |                |                           |
| CCMP      | 5.41             |                         |                  |                           |                |                           |

Figure 7. Distributions of (a) wind speed, (b) sensible heat flux, and (c) latent heat flux over the equatorial Indian Ocean. The black line in panel (a) shows the distribution of Cross-Calibrated Multi-Platform surface winds, while the black lines in panels (b and c) show the distributions of daily OAFlux air-sea fluxes. Model fluxes have been re-gridded to match the resolution of observations.
to AO4-CTRL (Figures 2k and 2l), which acts to increase LHF for a given wind speed. We can see evidence of this in the last days of the simulation, when surface winds in AO4 simulations remain distinct (Figure 8b), but LHF magnitudes begin to converge (Figure 8a). This is the result of AO4-FLX SSTs beginning to recover after MJO passage, while the SSTs in AO4-CTRL do not. This mechanism is addressed in more details in the following section.

Figure 8. Time series of Indian Ocean-averaged (a) rain rate (mm hr$^{-1}$), (b) latent heat flux (W m$^{-2}$), (c) surface wind speed (m s$^{-1}$), and (d) Sea Surface Temperature (SST) ($^\circ$C) in model simulations (color) and observations/analysis (black). Precipitation observations come from Tropical Rainfall Measuring Mission 3B42 (solid) and GPM IMERG (dashed), LHF observations come from OAFlux, surface winds from Cross-Calibrated Multi-Platform, and SST from Global High Resolution Sea Surface Temperature. Thin lines represent data at its native resolution, while thick lines show the 24-hr running mean of the data.
5. Upper Ocean Conditions During and After MJO Passage

The modification to the formulation of air-sea fluxes in the atmosphere portion of UWIN-CM results in an unexpectedly marked difference in SST compared to AO4-CTRL, and the processes behind it are addressed in this section. We focus on the equatorial IO (55°−97°E, 10°S−10°N; outlined in black in Figure 3), which coincides with the regions of MJO initiation and eastward propagation along the equator, and our convection-permitting domain. Though the model is good at reproducing the MJO-induced SST cooling (Figure 5), the coupled simulations tend to over-cool the IO (Figure 2, right), partly because the modeled MJO persists longer and over a larger area than what was observations show. However, SSTs in AO4-FLX are on average 0.1°C higher than in AO4-CTRL.

This systematic change in SSTs can result from a few different processes, some of which have been explored in the previous section. First, ocean can lose heat through positive SHF and LHF. Second, surface winds apply stress to the ocean surface, which contributes to the mechanical mixing of cooler water toward the surface. Third, cold water can be advected from a different region. And fourth, the ocean surface can cool by precipitation, creating a cool freshwater lens. Additionally, the structure of the ocean mixed layer itself can affect the change in SST, with effects like the MLD itself (the temperature would be more difficult to change with a deeper mixed layer than a shallower one) and the formation of barrier layers. Since the effect of precipitation is not represented in UWIN-CM, it cannot be the cause for the difference in this case. And though we see a difference in surface winds between AO4-CTRL and AO4-FLX (Figures 7 and 8), we only see a slight weakening of upper-ocean currents in the latter, which also cannot explain the consistent SST differences. As for the effects of heat loss from the ocean to the atmosphere through SHF, we see in the previous section that the difference in SHF between the simulations is very small, both right at the equator (Figure 6) and throughout the equatorial IO (Figure 7).

The mechanism for SST cooling that we focus on here is related to the ocean forcing by surface winds and air-sea fluxes, their effect on the upper ocean, and consequently, on SST. We know from the previous sections that AO4-CTRL produces stronger surface winds than AO4-FLX, which is an indirect dynamical response to the change in air-sea flux parameterization. Figure 9 shows the relationships among three sets of variable pairs: SST and ocean mixed layer depth (MLD; Figure 9a), SST and surface wind speed (Figure 9b), and surface wind speed and MLD (Figure 9c), establishing a relationship among the three quantities. MLD is defined as described in Section 2.4. The colored contours show the difference between the normalized joint distributions in AO4-FLX and AO4-CTRL (red means higher frequency in AO4-CTRL), and the black contours outline the AO4-CTRL joint distributions for reference.

Among each pair of variables, the change from AO4-CTRL to AO4-FLX is seen as a simple shift: higher SSTs are associated with shallower MLD (Figure 9a) and weaker surface winds (Figure 9b), while weaker surface winds are associated with shallower MLD (Figure 9c). Considering these relationships, we can identify a physical mechanism that can explain the differences in SST between AO4 simulations and identify surface winds as the main driver behind MLD and SST changes. Compared to AO4-CTRL, the weaker surface winds in AO4-FLX force less stress on the ocean surface, contribute to (slightly) slower ocean currents and induce less upper-ocean shear. Reduced shear leads to less mixing between the warm mixed-layer waters and the colder water below, resulting in shallower MLD and less SST cooling. With a shallower mixed layer, suppressed conditions after MJO passage (which include weak
winds and clear skies) can work to efficiently recover the SSTs back toward a pre-MJO state. Conversely, stronger surface winds (in AO4-CTRL) induce more cold water mixing into the surface ocean, deepening the MLD, cooling the SST further, and inhibiting its recovery.

To further examine the model simulated upper ocean conditions during and after MJO passage, we compare the model simulations with DYNAMO and RAMA mooring observations at three locations: 79°E and 0°N, 80.5°E and 0°N, and 79°E and 1.5°S (Figures 10–12). During the convectively active MJO period from 22 to 30 November, observations show near-surface (∼10 m) ocean temperature cooling, and a deepening of the ocean mixed layer, especially at the equator. Both the temperature and mixed layer depth are relatively well simulated by the coupled model. Strong surface winds are responsible for upper-ocean mixing.

From 1 to 6 December, after MJO passage, the observed mixed layer shoals and near-surface temperatures begin to increase due to the formation an oceanic barrier layer driven by reduced salinity (fresh water) near the surface (Figures 10 and 11). Both AO4 simulations deviate from observations, missing the near-surface fresh water, the barrier layer formation, and the consequent shoaling of the mixed layer and near-surface ocean temperature (and SST) increase. Instead, the simulated mixed layer after MJO passage remains deep, any surface warming is distributed through a thicker layer, and near-surface ocean temperatures remain steady in both simulations instead of recovering. Fresh water input from precipitation is not coupled to ocean salinity in the current model simulations. However, the ∼20 m deep layer of fresh water in the observations after 30 November (Figure 10) and 2 December (Figure 11) are not associated with local precipitation. Instead, they occur during the post-MJO phase, while there is no precipitation. The observed fresh water may be a result of advection according to the satellite surface salinity data (not shown).

Conditions are different south of the equator, where the upper ocean is generally fresher. At the mooring located at 79°E, 1.5°S, first there is a deepening of the mixed layer (and SST cooling) that lasts for ~4 days, then a very thick barrier layer forms due to strong near-surface freshening (Figure 12). These features are relatively well
simulated. But the modeled surface winds post-MJO remain biased high, the MLD deepens, and the near-surface ocean temperature continues to decrease beyond what is observed.

The overall evolution of SST and MLD from both observations and model simulations are summarized in Figure 13. At all three locations, the observation and model simulations track relatively close during the active MJO phase, when the strong surface-wind-induced mixing is dominant (22 to 30 November). However, they

Figure 11. Same as Figure 10, except for the RAMA mooring located at 80.5°E, 0°N. The observed surface wind and precipitation measurements were recorded by instruments on the RAMA mooring.

Figure 12. Same as Figure 10, except for the mooring located at 79°E, 1.5°S.
differ significantly during the post-MJO phase from 1 to 6 December. In all cases, AO4-FLX improves on the AO4-CTRL model simulation and reduces model SST bias by ∼50% and MLD bias by ∼16% (Figure 13).

6. Summary and Conclusions

This study investigates impacts of atmosphere-ocean coupling on MJO precipitation and eastward propagation, and upper ocean conditions during and after MJO passage. Three model simulations of the MJO event observed during DYNAMO in 2011 at convection-permitting (4 km) resolution with uncoupled atmosphere only and coupled atmosphere-ocean model are conducted to explore pathways for improving MJO prediction. A revised air-sea flux algorithm is implemented to further improve the coupled model simulation. The combination of ship and mooring observations over the central IO during the DYNAMO field campaign, satellite-derived precipitation and air-sea fluxes, and the high-resolution model data provide a unique opportunity to examine how air-sea interaction processes are represented in numerical models, and how they affect precipitation.

This study exposes three main findings:

1. Coupled atmosphere-ocean model reproduces the MJO eastward propagation whereas uncoupled atmospheric-only model fails
2. Upper-ocean cooling induced by the MJO is essential in reducing excessive and prolonged precipitation over the IO and aiding in MJO's eastward propagation
3. Improved air-sea flux algorithm reduces model biases in rainfall, surface winds, air-sea fluxes, as well as the bias in upper ocean temperature, all of which improve the MJO prediction

Atmosphere-ocean interaction is essential for the MJO's eastward propagation from the IO, as it creates local conditions unfavorable for sustained large-scale convection. In the coupled simulations (AO4-CTRL and AO4-FLX), strong surface winds associated with the MJO promote evaporation and provide fuel for precipitation, but they also cool the upper ocean, both through heat loss to the atmosphere, and cold upwelling due to

Figure 13. Comparison of observed and modeled 10-m ocean temperatures (top) and mixed layer depths (bottom) at three mooring locations: (a and e) 79°E, 0°N; (b and f) 80.5°E, 0°N; and (c and g) 79°E, 1.5°S. The rightmost column (d and h) shows the scatterplot of all three locations combined. The color change indicates time progression, with the red and blue colors for the AO4-CTRL and AO4-FLX simulations, respectively.
enhanced upper-ocean mixing. Cooler SSTs then suppress further evaporation and limiting moisture supply from the underlying ocean, reducing local precipitation; the MJO convection shifts eastward over warmer SSTs, where moisture is more readily available. The case is similar in both coupled simulations, with the main difference being that SST cooling, and subsequent warming are better captured in AO4-FLX, as is the MJO’s eastward propagation. In the atmosphere-only simulation, a positive feedback is established between the SST, winds, and precipitation. Warm SSTs are conducive to strong evaporation, providing an efficient moisture source for precipitation. But unlike in the coupled simulations, the atmospheric forcing does not affect SST, and, together with strengthening surface winds, evaporation remains high, as does the amount of precipitation. With a convenient endless moisture source, precipitation is ubiquitous, and we see no eastward propagation of large-scale precipitation or surface westerly winds.

The coupled atmosphere-ocean simulations reduce the overall precipitation bias over the IO by 40% compared to TRMM observations, but a high bias is still present in the second half of the simulations, when the observed state is strongly suppressed. This is linked to a reduction of moisture availability in AO4-CTRL due to a cooling ocean, but also to a high-biased surface air-sea flux parameterization in the model, which is on average ~25% higher than OAFlux over the equatorial IO in AO4-CTRL. The high LHF bias is present at all wind speeds and is thus not only a result of the models’ overproduction of strong winds. The AO4-FLX simulation was designed to attempt to reduce this bias at low winds through a reduction of convective velocity, the data for the parameterization of which was mostly obtained over land (Beljaars, 1995). Ultimately, the modification resulted in the reduction of the LHF bias over the IO from ~25% to ~15% over the equatorial IO and brought with it a reduction in total precipitation (down 30% from AO4-CTRL bias), and improved MJO propagation, surface winds, and SST.

These results are supported by previous studies that have been able to accurately reproduce an MJO event only when either using a coupled atmosphere-ocean model, or updating SSTs with observed values (Hung et al., 2013; Miura et al., 2009; Zhang et al., 2006), and very few that were able to reproduce an event when using fixed SSTs (Holloway et al., 2013, 2015). However, Drushka et al. (2012) found that during active MJO, the shear-induced cold upwelling during active MJO conditions represents half of the observed temperature change in the mixed layer, which underlines the importance of dynamical ocean models in lieu of using constant-depth slab ocean models.

Despite these improvements, the coupled simulations still show difficulty representing the convectively suppressed conditions after the MJO passage to the extent seen in observations. There is still excessive precipitation over the IO (Figure 2), part of which could be linked to the models’ slightly high-bias in surface winds and air-sea fluxes (especially the LHF) injecting too much moisture into the atmosphere (Figure 8). We also note the difficulties the coupled simulations have in representing the recovering SST through the inadequate formation of barrier layers after MJO passage, in the second half of the simulations. While MJO convection is over the IO, strong winds induce high heat fluxes at the surface, and with entrainment cooling from mixed layer deepening, both factors act to reduce SST by 0.7°C on average—this agrees with previous observational studies (e.g., Lau & Sui, 1997; Maloney & Kiehl, 2002). After active MJO convection passes observed winds weaken, leading to a reduction in air-sea fluxes and upper-ocean mixing, and with increased insolation, we see the formation of oceanic barrier layers near the equator (mooring observations in Figures 10 and 11). Barrier layer formation rapidly shoals the mixed layer, and as surface fluxes are now distributed over a thinner layer, it makes it easier for SSTs to recover. But in coupled model simulations, convection persists a bit longer, and surface winds to not weaken enough to allow for barrier layer formation—thus, any reduction in surface fluxes and increased insolation keep being distributed through a very thick layer, making post-MJO SST recovery slower and weaker than observed.

It has been noted that air-sea interaction is more important for some MJO events than it is for others (Fu et al., 2015; Wang et al., 2015), and that the importance of air-sea interaction depends on the underlying ocean (Drushka et al., 2014; Fu et al., 2015; Moum et al., 2016). For example, if an MJO is passing over an ocean with a deep, warm mixed layer, the surface fluxes and wind stress would be distributed through a thicker layer, and any changes to SST would be small compared to an MJO passing over a shallower mixed layer (Fu et al., 2015; Marshall et al., 2008). While most current studies focus on the role of SST in providing an environment for moisture-convergence to the east of the MJO convection, many studies have studied the processes that lead to the relatively rapid SST recovery after MJO passage, and we have identified the lack of barrier layer formation as one of the culprits in the UWIN-CM simulations. Due to the demonstrated importance of upper-ocean mixing and
air-sea fluxes, our results indicate that upper ocean processes and their two-way interaction with the MJO over the Indian Ocean should be examined closely by further studies.

**Data Availability Statement**

Observations from the DYNAMO field campaign used for the study are available on the DYNAMO legacy data portal (https://orca.atmos.washington.edu/dynamo_legacy/). RAMA mooring observations were obtained through https://www.pmel.noaa.gov/iao/drupal/disdelf/. Satellite precipitation estimates are available at https://doi.org/10.5067/TRMM/TMPA/3H/7 (Precipitation Processing System (PPS) At NASA GSFC, 2018) (Tropical Rainfall Measuring Mission 3B42) and https://doi.org/10.5067/GPM/IMERG/3B-HH/06 (Precipitation Processing System (PPS) At NASA GSFC, 2019) (Global Precipitation Measurement GPM Integrated Multi-Satellite Retrievals for GPM), respectively. Cross-Calibrated Multi-Platform Version-2.0 vector wind analyses are produced by Remote Sensing Systems, and data is available at www.remss.com. Foundation SST is produced by the Group for High Resolution Sea Surface Temperature and is available at https://doi.org/10.5067/GHGMR-4FJ04. (NASA/JPL, 2015) OAAFlux is available at https://doi.org/10.5065/0JDQ-FP94 (Physical Oceanography Department/Woods Hole Oceanographic Institution, & Goddard Institute For Space Studies/Earth Sciences Division/Science And Exploration Directorate/Goddard Space Flight Center/NASA, 2006).

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