MJO-Induced Intraseasonal Mixed Layer Depth Variability in the Equatorial Indian Ocean and Impacts on Subsurface Water Obduction

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ABSTRACT: Change of oceanic surface mixed layer depth (MLD) is critical for vertical exchanges between the surface and subsurface oceans and modulates surface temperature variabilities on various time scales. In situ observations have documented prominent intraseasonal variability (ISV) of MLD with 30–105-day periods in the equatorial Indian Ocean (EIO) where the Madden–Julian oscillation (MJO) initiates. Simulation of Hybrid Coordinate Ocean Model (HYCOM) reveals a regional maximum of intraseasonal MLD variability in the EIO (70°–95°E, 3°S–3°N) with a standard deviation of ~14 m. Sensitivity experiments of HYCOM demonstrate that, among all of the MJO-related forcing effects, the wind-driven downwelling and mixing are primary causes for intraseasonal MLD deepening and explain 83.7% of the total ISV. The ISV of MLD gives rise to high-frequency entrainments of subsurface water, leading to an enhancement of the annual entrainment rate by 34%. However, only a small fraction of these entrainment events (<20%) can effectively contribute to the annual obduction rate of 1.36 Sv, a quantification for the amount of resurfacing thermocline water throughout a year that mainly (84.6%) occurs in the summer monsoon season (May–October). The ISV of MLD achieves the maximal intensity in April–May and greatly affects the subsequent obduction. Estimation based on our HYCOM simulations suggests that MJOs overall reduce the obduction rate in the summer monsoon season by as much as 53%. A conceptual schematic is proposed to demonstrate how springtime intraseasonal MLD deepening events caused by MJO winds narrow down the time window for effective entrainment and thereby suppress the obduction of thermocline water.

KEYWORDS: Indian Ocean; Entrainment; Oceanic mixed layer; Intraseasonal variability

1. Introduction

The oceanic mixed layer, typically a layer with vertically uniform temperature, salinity, and density near the sea surface, is the media between the atmosphere and the ocean interior. The mixed layer depth (MLD), remarking the base of the surface mixed layer, is critical for the heat budget of the surface ocean and modulates sea surface temperature (SST) variability and atmospheric convection (e.g., Shinoda and Hendon 1998; Alexander et al. 2000; Dommenges and Latif 2002; Schott et al. 2009). For example, the tropical cyclone intensity depends critically upon the initial MLD in its formation region (e.g., Chang and Anthes 1978; Bender et al. 1993; Chan et al. 2001; Zhao and Chan 2017). The MLD also represents the volume of seawater over which surface heat, freshwater, and momentum inputs would be distributed (e.g., Chen et al. 1994), with profound impacts on ocean biology (e.g., Fasham 1995; Polovina et al. 1995; Behrenfeld 2010) and near-surface acoustic propagation (Sutton et al. 1993). The change of MLD can cause downward and upward mass fluxes across the mixed layer base, dubbed as detrainment and entrainment,

respectively (e.g., De Szoek 1980; Cushman-Roisin 1987). Detrainment/entrainment, carrying momentum, energy, and carbon dioxide fluxes, drives interior ocean motions and conveys climate change signals between the atmosphere and subsurface ocean. Therein, the mass flux between the mixed layer and the permanent pycnocline is described as subduction/obduction (Huang 1988; Qiu and Huang 1995), corresponding to the formation/erosion of ventilation-related water masses and serving as the vertical limbs of global ocean circulation.

The theory of subduction and ventilation was initially based on the water mass formation framework postulated by Iselin (1939) and Stommel (1979), which were followed by a number of ideal-fluid pycnocline models for the diagnosis of subduction/obduction rate (e.g., Woods 1985; Huang 1990; Williams 1991; Qiu and Huang 1995; Marshall 1997). A widely used diagnosis framework was proposed by Qiu and Huang (1995), in which only the climatological annual cycle of MLD is considered, and subduction and obduction are assumed to occur only in late winter and early winter, respectively. This framework has been widely utilized to estimate the subduction/obduction rate and its variability (e.g., Karstensen and Quadfasel 2002a,b; Qu and Chen 2009; Trossman et al. 2009; Liu and Huang 2012). Recent studies have worked on the refinement of the subduction/obduction diagnosing method by taking into account more realistic MLD variability. It has been demonstrated that the semiannual variability of MLD in regions controlled by

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monsoons can give rise to an additional subduction/obduction window in the summer monsoon season and weaken the subduction/obduction in the winter monsoon season (Liu et al. 2018, 2020).

A remarkable advance in the understanding of subduction/obduction is brought by the finding that the high-frequency MLD variability associated with mesoscale eddies can greatly enhance subtropical subduction, particularly for the mode waters, as revealed by in situ observations (e.g., Uehara et al. 2003; Kouketsu et al. 2011; Xu et al. 2014, 2016) and numerical simulations (Qu et al. 2002; Rainville et al. 2007; Nishikawa et al. 2010). By contrast, there were few studies to appreciate the impact of high-frequency MLD variability on the obduction phase, which mainly takes place in the subpolar and tropical regions. The tropical oceans are subjected to strong modulation effects of atmospheric intraseasonal oscillations. Considering that the Madden–Julian oscillation (MJO; Madden and Julian 1971) is the dominant mode of atmospheric intraseasonal oscillations, we use the “MJO” as a collective representation for convenience in the study. Under strong forcing effects of MJOs, the tropical MLD exhibits prominent intraseasonal variations (e.g., Drushka et al. 2012, 2014; Keerthi et al. 2016; Li et al. 2017) and possibly affects the tropical obduction. It is still unknown whether the MJO-induced MLD variability can significantly enhance the obduction, in analogy to the effect of mesoscale eddies on subtropical subduction. Exploring this problem provides an opportunity to further refine the theoretical understanding and diagnosing methods of the thermocline water obduction in tropics. These efforts may also provide implications for understanding climate change, given that the obduction brings anomalies of temperature, salinity, carbon dioxide, and nutrients to the sea surface and potentially modulates air–sea interactions.

The equatorial Indian Ocean (EIO) is home to most strong MJO events (Zhang 2005) and exhibits strong intraseasonal variability (ISV) in SST (e.g., Duvel et al. 2004; Han et al. 2007; Duncan and Han 2009; Jayakumar et al. 2011), ocean salinity (e.g., Matthews et al. 2010; Grunseich et al. 2011; Li et al. 2015), and upper-ocean circulation (e.g., Schiller and Godfrey 2003; Masumoto et al. 2005; Chen et al. 2015, 2019). Intraseasonal SST variability is actively coupled to the MJO’s atmospheric variabilities (Hendon and Glick 1997; Wang and Xie 1998; Woolnough et al. 2001). The EIO is, therefore, an ideal region to examine the MJO’s possible effect on MLD and subsurface water obduction. Estimations based on in situ observations suggested that in this region, the MLD change during strong MJO events achieves a peak-to-peak difference of more than 15 m in boreal winter (Drushka et al. 2012, 2014). Keerthi et al. (2016) explored large-scale intraseasonal MLD variability in the Indian Ocean and found that the summertime MLD in the EIO primarily responds to the intraseasonal active/break spells of the summer monsoon, while the MJO is the primary cause for wintertime intraseasonal MLD anomalies. Using ocean model simulations, Li et al. (2017) showed that intraseasonal MLD variability achieves amplitudes of ~ 10 m in the EIO and the Bay of Bengal in summer. In spite of existing knowledge, a comprehensive investigation is required to clarify how MJO

events cause MLD variability in the EIO, such as the relative importance of surface wind forcing and buoyancy forcing exerted by MJOs.

Given the prominent intraseasonal MLD variability, it is expected that strong entrainment events can occur during the active phase of the MJO when strong winds and radiation-induced cooling drive dramatic thickening of the mixed layer (Drushka et al. 2012, 2014; Li et al. 2017). These entrainment events may greatly enhance the equatorial obduction of the cold, nutrient-rich thermocline water, which could be fundamental for understanding SST variability, air–sea interactions, and marine biological processes (e.g., Chen et al. 2016). Results of ocean model simulations presented in this study, however, suggest quite the opposite. Taking into consideration the MJO forcing effects actually reduces the annual obduction rate. It is an interesting and debatable question as to how and why MJOs suppress the equatorial obduction of thermocline water.

The overall purpose of the present study is to reveal the key processes through which MJOs cause MLD variability in the EIO and explore the possible impacts of MJO-induced high-frequency MLD variability on the equatorial obduction. This is pursued mainly by performing and analyzing ocean general circulation model (OGCM) experiments. The rest of the paper is organized as follows. Section 2 describes the data and model utilized in this study, especially the OGCM experiments designed for our analysis. Section 3 presents the characteristics and mechanisms of intraseasonal MLD variability. Section 4 provides evaluations for the MJO forcing effects on processes of entrainment and obduction, and a conceptual schematic is postulated to demonstrate how the MJO-induced MLD variability affects obduction. Section 5 provides a summary of the main findings of the paper.

2. Data and models

a. Data

In situ temperature measurements during 2012–15 by two moored buoys of the Research Moored Array for African–Asian–Australian Monsoon Analysis and Prediction (RAMA; McPhaden et al. 2009) at 0° , 80.5°E and 1.5°S , 80.5°E are used to estimate the ISV of MLD and validate the OGCM results. Daily temperature data collected typically at 1-, 10-, 13-, 20-, 40-, 60-, 80-, and 100-m depths are linearly interpolated into 1-m vertical intervals. Because of insufficient sampling coverage of salinity data, MLD is computed as the depth at which the temperature decreases 0.5°C from the surface value based on RAMA temperature data (Monterey and Levitus 1997).

We also utilize the Grid Point Value of the Monthly Objective Analysis (MOAA GPV) data to obtain the climatological structure and seasonal variation of MLD in the EIO. MOAA GPV is based on Argo float, moorings, and available CTD casts (Hosoda et al. 2008). This dataset provides optimal interpolated, $1^\circ \times 1^\circ$, monthly temperature and salinity fields at standard pressure levels between 10 and 2000 dbar. The MOAA GPV is a monthly dataset and is not able to resolve the intraseasonal MLD. Therefore, we employ the Simple Ocean Data Assimilation (SODA), version 3.4.2, to represent the

observed intraseasonal MLD variability. It is built upon the eddy-permitting Modular Ocean Model, version 5, and the details can be found in Carton et al. (2018). The datasets are mapped onto a uniform $0.5^\circ \times 0.5^\circ$ horizontal grid with a temporal resolution of 5 days. Daily 10-m vector wind data from the $0.25^\circ \times 0.25^\circ$ Cross-Calibrated Multi-Platform (CCMP), version 2, ocean wind product (Atlas et al. 2008) for 2000–11 and the $0.25^\circ \times 0.25^\circ$ gridded Advanced Scatterometer (ASCAT) surface wind measurements (Bentamy and Croize-Fillon 2012) for 2012–14 are used to analyze surface wind variability of MJOs. The zonal and meridional surface wind stresses are calculated using the standard bulk formula

$$\tau^x = \rho_a C_d |V_{10}| u_{10} \quad \text{and} \quad \tau^y = \rho_a C_d |V_{10}| v_{10}, \quad (1)$$

where $\rho_a = 1.175 \text{ kg m}^{-3}$ is the air density, $C_d = 0.0015$ is the drag coefficient, and u_{10} and v_{10} are zonal and meridional components of 10-m winds, respectively.

For MOAA GPV, SODA, and OGCM outputs that contain salinity records, MLD is defined as the depth at which potential density is larger than the sea surface value by 0.125 kg m^{-3} (Levitus 1982). Unless otherwise noted, the intraseasonal anomaly of an oceanic or atmospheric variable is obtained through removing the monthly climatology and applying a 30–105-day bandpass Lanczos digital filter. The results are generally insensitive to choices of the upper and lower bounds of the bandpass window.

b. HYCOM and experiments

The ocean model used in this study is the Hybrid Coordinate Ocean Model (HYCOM; e.g., Wallcraft et al. 2009), version 2.2.18, which has been successfully utilized to investigate MJO-related oceanic variations (e.g., Li et al. 2014, 2015, 2017; Chen et al. 2015, 2019). For this study, HYCOM is configured to the Indian Ocean within $30^\circ\text{--}122.5^\circ\text{E}$, $50^\circ\text{S--}30^\circ\text{N}$, with horizontal resolutions of $0.25^\circ \times 0.25^\circ$ and 26 hybrid layers in the vertical direction. Thickness of the top layer is $\sim 2.9 \text{ m}$ near the sea surface. Three 5° sponge layers are applied at the western, eastern, and southern open-ocean boundaries, where the model temperature and salinity are relaxed to monthly climatology. The diffusion/mixing parameters are identical to those adopted in Li et al. (2013).

Surface atmospheric forcing fields include surface winds from CCMP for 2000–11 and ASCAT for 2012–14, $1^\circ \times 1^\circ$ surface shortwave radiation (SWR) and longwave radiation (LWR) from Clouds and the Earth's Radiant Energy System (CERES; Wielicki et al. 1996; Loeb et al. 2001), $0.25^\circ \times 0.25^\circ$ precipitation from the Tropical Rainfall Measuring Mission (TRMM) Multisatellite Precipitation Analysis (TMPA) level 3B42 product (Kummerow et al. 1998; Huffman et al. 2007), and $0.75^\circ \times 0.75^\circ$ 2-m air temperature and humidity from ERA-Interim (Dee et al. 2011). Note that wind speed and wind stress are fed separately to the model. Wind stress affects the model ocean through driving advection, upwelling/downwelling, and turbulent mixing, while wind speed affects the ocean mainly through evaporation and turbulent heat flux. HYCOM computes evaporation and turbulent heat flux with wind speed, air temperature, specific humidity, and surface temperature

using the bulk formula of the Coupled Ocean–Atmosphere Response Experiment (COARE 3.0) algorithm (Kara et al. 2005). For river discharges, we use the satellite-based monthly discharge records of the Ganga–Brahmaputra from Papa et al. (2010) and monthly discharge data of Dai et al. (2009) for other rivers.

The model is spun up from a state of rest for 30 years using monthly climatological forcing, and then it is integrated forward from March 2000 to December 2014. Six experiments are designed to isolate the effects of different processes (Table 1; see also Li et al. 2015). The main run (MR) is forced by original daily forcing fields, and its solution is assumed to contain complete processes and used for validation against observational data. To exclude the MJO-related forcings, we performed the NoMJO experiment in which all atmospheric forcing fields are low-passed with a 105-day Lanczos digital filter. Thus, ISV in NoMJO originates from the oceanic internal processes, such as mesoscale eddies and tropical instability waves. The difference (MR – NoMJO) measures the overall effects of MJOs on the ocean. In the third experiment, NoSWR, only SWR is 105-day low-passed, and therefore its difference from MR, MR – NoSWR, measures the effect of intraseasonal SWR anomalies on the ocean. In NoWND, both wind speed and wind stress are low-pass filtered, whereas in NoSTRS only wind stress is low-pass filtered. The difference between MR and NoSTRS (MR – NoSTRS) measures the effect of wind stress, and NoSTRS – NoWND reflects the effect of wind speed. In the NoPRCP experiment, we use only 105-day low-passed filtered precipitation, and MR – NoPRCP isolates the precipitation effect. Outputs of the six experiments are stored as 3-day mean data from March 2000 to December 2014. The outputs in 2000 and 2001 are discarded to exclude the transient effects from the spinup run, and the 13-yr data of 2002–14 are used for our analysis.

3. Intraseasonal variability of mixed layer depth

Measurements by RAMA buoys at B1 (0° , 80.5°E) and B2 (1.5°S , 80.5°E) show large-amplitude variability of MLD, with evident intraseasonal fluctuations superimposed on the seasonal cycle (Fig. 1). The MLD generally reaches the deepest position in summer, as the response to the mature stage of the Indian summer monsoon. The seasonal difference of MLD achieves 40–60 m (Figs. 1a,b). There are noticeable intraseasonal deepening events, especially during the winter monsoon season. In the strong deepening event during November 2014–January 2015 at B1, the MLD can reach $\sim 100 \text{ m}$, a depth equivalent to the summertime MLD (Fig. 1a). As quantified by the 30–105-day bandpass filtered MLD anomalies, ISV of MLD can reach as large as 15 m in amplitude, with a standard deviation (STD) of 6.7 m at B1 (Fig. 1c). At B2, intraseasonal fluctuations of $\sim 20 \text{ m}$ are found, and the STD is 7.3 m (Figs. 1b,d). It is noted that the intensity of ISV of MLD may be different in different periods. Taking B2 for example, the STD of intraseasonal MLD anomalies between March 2009 and March 2011 achieves 10.4 m (figure not shown). These results derived from RAMA buoys confirm the existence of prominent intraseasonal MLD variability in the EIO region.

TABLE 1. Summary of HYCOM experiments.

Expts	Forcing	Description
MR	Daily forcing	Complete run
NoMJO	105-day low-passed forcing	Remove all MJO effects
NoSWR	105-day low-passed SWR	Remove MJO SWR effect
NoWND	105-day low-passed wind speed/stress	Remove MJO wind stress/speed effect
NoSTRS	105-day low-passed wind stress	Remove MJO wind stress effect
NoPRCP	105-day low-passed precipitation	Remove MJO precipitation effect

Because of the nonlinear nature of ocean stratification, the background mean state and seasonal variation of MLD can affect ISV. Thus, before analyzing the ISV, we first examine the mean state and seasonal variation of MLD in the EIO. As shown in Fig. 2, the spatial structures of climatological MLD in the MR are generally consistent with MOAA GPV, such as the deep MLDs south of 13°N and in the Arabian Sea and the shallow MLDs in the southwest Indian Ocean. There are overly deep MLDs in the Bay of Bengal in HYCOM (Figs. 2a,b). This is likely a common bias in the ocean model simulations (e.g., Duncan and Han 2009; Li et al. 2013, 2014, 2015; Felton et al. 2014) related to the complicated thermal and saline stratification near the sea surface, including the shallow MLD and the thick barrier layer (e.g., Thadathil et al. 2007; Li et al. 2017). Nevertheless, in the EIO on which our analysis focuses, the simulated MLD is close to observation. The monthly anomalies of MLD, relative to the climatological mean MLD, averaged over 3°S – 3°N are used to check the seasonal evolution in the EIO (Figs. 2c,d). The MLD is deeper in summer than in winter, in agreement with RAMA buoy measurements (Fig. 1a). The seasonal variation of MLD reaches ~ 40 m, close to RAMA results, and the largest MLD

occurs in June–August. These seasonal features are simulated well by HYCOM.

The STD of 30–105-day bandpass filtered MLD measures the overall intensity of the intraseasonal MLD variability. The prominent ISV is concentrated in the central-to-eastern portion of the Indian Ocean equator, with the largest STD of ~ 14 m (Fig. 2f). It is consistent with 10–15-m STD of intraseasonal MLD by Keerthi et al. (2016). Moreover, the STD of intraseasonal MLD in HYCOM MR agrees well with SODA data in both spatial distribution and typical intensity in the EIO box (Figs. 2e,f). The following analysis will focus on this EIO box (70° – 95°E , 3°S – 3°N). It is noted that we also calculate the intraseasonal MLD with another definition, as the depth with potential density increase from the sea surface value equivalent to a temperature decrease of 0.5°C (de Boyer Montegut et al. 2004). Over the area of interest marked by the black box in Fig. 2, the STD maps based on two MLD definitions agree well with each other in both spatial distribution and typical intensity (figure not shown). Thus, the ISV of MLD in the EIO is insensitive to the choice of methods diagnosing MLD. In general, HYCOM MR can well represent the large-scale distribution and seasonal/intraseasonal variation of MLD in the EIO.

FIG. 1. Daily temperature (color shading; $^{\circ}\text{C}$) and MLD (black lines; m) measured by RAMA buoys (a) B1 (0° , 80.5°E) and (b) B2 (1.5°S , 80.5°E). Also shown is 30–105-day bandpass filtered MLD anomalies at (c) B1 and (d) B2. The positive MLD anomaly indicates a deeper MLD than the average.

FIG. 2. Annual-mean climatological MLD (m) during 2002–14 from (a) MOAA GPV data and (b) HYCOM MR, the monthly MLD anomaly (m) averaged over 3°S–3°N from (c) MOAAGPV data and (d) HYCOM MR, and the STD of 30–105-day MLD from (e) SODA and (f) HYCOM MR. The black-outlined rectangle defines the equatorial Indian Ocean region of 70°–95°E, 3°S–3°N.

The change of MLD is a complex process, determined by a mutable balance between the destabilizing effect of surface wind stress through turbulent mixing and the stabilizing effect by density stratification that are regulated by surface buoyancy fluxes (heat and freshwater) and ocean dynamics (advection and upwelling/downwelling). It is instructive to clarify the major causes for the intraseasonal MLD variability. Figure 3 shows the intraseasonal MLD anomalies induced by different atmospheric forcing effects and oceanic internal processes in the EIO. The STD of 30–105-day MLD averaged over the EIO region is 6.16 m in the MR (Fig. 3a), roughly consistent in magnitude with RAMA buoy measurements (Fig. 1). The total forcing effect by MJOs, as represented by MR – NoMJO, is a STD of 6.22 m and a correlation of 0.96 with that of MR. The small difference in STD (6.16 vs 6.18 m) is statistically insignificant as suggested by an F test and likely arises from ocean internal instability. On the other hand, the 30–105-day MLD from NoMJO, as the ocean internal processes, has a small STD (1.63 m) and an insignificant correlation with the MR (0.10). The sharp contrasts between MR and NoMJO suggest the dominant role played by the MJO in causing intraseasonal

MLD variability and a minor role played by the oceanic internal processes.

Among all the MJO-related forcing effects, the wind stress effect measured by MR – NoSTRS induces the largest STD of 5.41 m and the highest correlation with the MR (0.95) (Fig. 3b). In our model setting, surface wind stress controls turbulent mixing and ocean dynamics such as advection and upwelling/downwelling. This forcing effect is likely the primary process through which the MJO affects MLD. As an evaluation for both magnitude and correlation, the overall contribution of an effect to the total variability can be estimated through a linear regression (Li et al. 2015),

$$C = \frac{\langle \text{MLD}^{\text{Effect}}(t) \times \text{MLD}^{\text{MR}}(t) \rangle}{\langle \text{MLD}^{\text{MR}}(t) \times \text{MLD}^{\text{MR}}(t) \rangle} \times 100\%, \quad (2)$$

where MLD^{MR} and $\text{MLD}^{\text{Effect}}$ are 30–105-day MLD anomalies of MR and an effect, respectively, and the angle brackets denote the summation over time. In the EIO box, the contribution of wind stress effect reaches 83.7%. The wind speed effect, as measured by NoSTRS – NoWND, has a small STD of

FIG. 3. The 30–105-day MLD (m) averaged in the EIO region (70°–95°E, 3°S–3°N) from the MR (black), (a) NoMJO (cyan) and MJO effect (MR – NoMJO; pink), (b) wind stress effect (MR – NoSTRS; orange), (c) wind speed effect (NoSTRS – NoWND; green), (d) SWR effect (MR – NoSWR; red), and (e) precipitation effect (MR – NoPRCP; blue). STD and linear correlations are listed in each plot.

1.56 m but a significant correlation of 0.70 with the MR (Fig. 3c). Wind speed and wind stress are separately exerted onto the model ocean in HYCOM. It mainly affects the magnitude of surface turbulent heat fluxes (latent and sensible heat fluxes) and operates as surface buoyancy forcing for MLD. The cloud-cover-related surface SWR is another surface buoyancy forcing process that the MJO modulates the oceanic mixed layer (e.g., Duvel et al. 2004; Jayakumar et al. 2011; Li et al. 2014, 2015). The SWR effect produces a STD of 1.97 m, and its correlation with the MR is 0.53 (Fig. 3d). Contributions of the wind speed and SWR effects to the total variability in MR are rather close, and they are 17.7% and 16.9%, respectively. The precipitation effect also shows a small STD of 1.70 m (Fig. 3e).

Different from others, it shows a negative correlation with MR (–0.20). The results shown in Fig. 3 overall suggest a much stronger wind forcing effect than the buoyancy forcing effect of the MJO on intraseasonal MLD variability in the EIO. This can be verified by comparing MLD with the Monin–Obukhov (Monin and Obukhov 1954) length

$$L_{MO} = u_*^3 / (\kappa B_0), \quad (3)$$

where the friction velocity is calculated by $u_* = (\tau/\rho)^{1/2}$ for surface wind stress τ and water density ρ , and $\kappa = 0.4$ is the von Kármán constant. The surface buoyancy flux is calculated by $B_0 = [\alpha g / (\rho c_p)] Q_0 + \beta g S_0 F_0$, where Q_0 is the surface heat flux,

FIG. 4. Time-longitude plots of composite MJO cycle based on RMM index: (a) wind stress magnitude (N m^{-2}), (b) zonal wind stress (N m^{-2}), (c) SLA (cm), (d) MR MLD (m), (e) wind stress effect on MLD (MR – NoSTRS), (f) wind speed effect on MLD (NoSTRS – NoWND), (g) SWR effect on MLD (MR – NoSWR), and (h) precipitation effect on MLD (MR – NoPRCP). All variables are 30–105-day bandpass filtered and averaged over 3°S – 3°N before performing the composite.

F_0 is the net surface freshwater flux, S_0 is the surface salinity, and c_p is the heat capacity of seawater, and α and β are the thermal and haline coefficients of expansion, respectively. Taking the station (2°N , 82°E) as an example, during the time that the buoyancy flux was convection favorable, that is, positive L_{MO} , the STD of L_{MO} calculated on the basis of intraseasonal forcing (wind stress and buoyancy flux) is estimated as 63.1 m, which is greatly larger than that of intraseasonal MLD (8.9 m). According to Lombardo and Gregg (1989), $L_{\text{MO}} > \text{MLD}$ indicates the dominance of wind stress. Using other

locations of EIO generally achieves similar results. Thus, it can be concluded that the wind forcing is dominant over the surface buoyancy forcing for intraseasonal MLD deepening.

To better understand these forcing effects, we compute the MJO composite based on the real-time multivariate (RMM) MJO index (Wheeler and Hendon 2004) for all useful variables (Fig. 4). The index describes the MJO evolution as an eight-phase cycle based on the location of large-scale outgoing longwave radiation and associated changes in vertical wind shear. For a given MJO phase, the composite properties are

formed by averaging the variables over the days with RMM index in that phase and RMM magnitude > 1.5 (indicating significant atmospheric perturbations). In this way, the composite results represent the mean evolution of atmospheric/oceanic conditions during the passage of MJOs, and this method was widely used in previous studies (e.g., Guan et al. 2014; Wang et al. 2016; Zhu et al. 2020). The surface wind stress, which is critical for intraseasonal MLD variability, achieves the maximal magnitude in phase 3 of the composite MJO (Fig. 4a). Phase 3 generally characterizes a stage when the MJO has its fully developed convection center, involving windy and cloudy conditions, lingering over the EIO (Wheeler and Hendon 2004). The forcing effect of surface wind stress on MLD is twofold. The increased wind stress enhances turbulent mixing near the ocean surface through kinetic energy input, while the westerly wind anomalies (Fig. 4b) evoke downwelling equatorial Kelvin waves (e.g., Kessler et al. 1995; Hendon et al. 1998; McPhaden and Yu 1999) which cause upper-ocean convergence (e.g., Somavilla et al. 2017), as indicated by eastward-propagating positive sea level anomalies (SLAs; Fig. 4c). Chen et al. (2015) have pointed out that MJO-driven Kelvin waves can affect the intraseasonal MLD variability in the equatorial Indian Ocean by the adiabatic processes. Likely, the enhanced turbulent mixing and the convergence of surface warm water work mutually to cause the deep MLD in phases 3–5 in MR and MR – NoSTRS (Figs. 4d,e). The westerly wind anomalies exhibit swift eastward propagation in the central and eastern basin. Correspondingly, SLAs and MLD anomalies also show eastward propagation tendencies but with slower transition speed, possibly reflecting the slower phase speed of oceanic Kelvin waves ($\sim 2.6 \text{ m s}^{-1}$ for the heaviest baroclinic mode) than that of MJOs ($\sim 5.0 \text{ m s}^{-1}$). It is noticeable that the SLA amplitude is much stronger near the eastern boundary than in the EIO interior, while the MLD anomalies in the two areas are comparable. This indicates that turbulent mixing by the increased wind stress does contribute to the MLD deepening, in addition to the downwelling effect of westerly winds. Here we are unable to determine the relative importance of the two processes. Moreover, it is unknown why there is seemingly westward propagation of wind stress anomaly west of 60°E (Fig. 4a). However, it has little impact on our results.

The close resemblance in anomaly features between Figs. 4d and 4e confirms the importance of wind stress in causing MLD changes. By contrast, anomalies induced by wind speed, SWR, and precipitation are smaller in magnitude and show apparent differences in the timing. MLD anomalies produced by wind speed and SWR tend to lead those of MR by 1–2 phases (Figs. 4f,g), while the precipitation effect is roughly out of phase with MR (Fig. 4h). The negative contribution by precipitation to MLD change reflects the fact that the enhanced rainfall during the active phase of MJO convection acts to shoal the MLD through surface buoyancy input, and this effect is opposite to the wind-driven MLD deepening in MR and MR – NoSTRS.

Next, we explore how the effects of different processes are superimposed to yield the total MLD variability in MR. Because the precipitation contribution is weak and negative, here we only examine the wind stress, wind speed, and

FIG. 5. MJO composites of (a) MLD (m), (b) SLA (cm), and (c) surface net heat flux (W m^{-2}) derived from MR (black), wind stress effect (MR – NoSTRS; orange), wind speed effect (NoSTRS – NoWND; green), and SWR effect (MR – NoSWR; red). All variables are 30–105-day bandpass filtered and averaged in the EIO box before permitting the composite.

SWR effects. As shown in Fig. 5a, the evolutions of MLD are highly consistent between MR (black) and the wind stress effect (orange). They are deep during phases 3–5, reaching the peak in phase 4, and shallow in phase 8. SLA, representing the anomalous upper-ocean convergence/divergence, is primarily controlled by wind stress (Fig. 5b) through ocean dynamical processes such as equatorial Kelvin waves. The wind stress-driven SLAs also show a peak in phase 4 and a trough in phase 8, largely mimicking the change of MLD. The resemblance between MLD and SLA confirms that the upper-ocean convergence/divergence is a major process through which the MJO wind stress causes the MLD change, in addition to wind

FIG. 6. As in Fig. 3, but for intraseasonal entrainment rate anomaly (10^{-6} m s^{-1}) of the EIO region.

stress-driven turbulent mixing. In comparison, the MLD deepening anomalies caused by the wind speed effect (green) and SWR effect (red) are only $\sim 1 \text{ m}$ and much weaker than the $\sim 4 \text{ m}$ deepening caused by the wind stress effect (orange). They tend to occur in phases 2–3 (Fig. 5a), corresponding to the surface heat loss peaks in phase 2–3 (Fig. 5c). This explains why their effects on MLD lead the total MLD anomaly by 1–2 phases. These buoyancy forcing effects exerted by SWR and wind speed-controlled turbulent heat flux are weaker than wind stress forcing and mainly contribute to the initial deepening of MLD in phase 3.

4. Impacts on obduction

MLD change serves as one of the causes of entrainment, which can be written as (e.g., De Szoeke 1980; Cushman-Roisin 1987)

$$E = w_{\text{mb}} + \mathbf{u}_{\text{mb}} \cdot \nabla h_m + \frac{\partial h_m}{\partial t}, \quad (4)$$

where w_{mb} and $\mathbf{u}_{\text{mb}} = (u_{\text{mb}}, v_{\text{mb}})$ are vertical and horizontal velocities at the mixed layer base, respectively, and h_m denotes MLD. In the entrainment phase, i.e., $E > 0$, water mass goes from the subsurface ocean into the surface mixed layer, whereas in the detrainment phase, water mass is discharged from the mixed layer down to the pycnocline.

The MJO-induced strong intraseasonal MLD deepening events are expected to cause large entrainment events. In the EIO, the intraseasonal fluctuations of entrainment from the MR can reach as large as $3 \times 10^{-5} \text{ m s}^{-1}$, with the STD of $8.82 \times 10^{-6} \text{ m s}^{-1}$ (Fig. 6a). Similar to the case of MLD, the ISV of entrainment rate is predominantly caused by the MJO. The difference, $\text{MR} - \text{NoMJO}$, shows a slightly higher STD ($8.93 \times 10^{-6} \text{ m s}^{-1}$) and a high correlation with the MR (0.98). The sum of entrainment events in one year, dubbed as annual

FIG. 7. (a) Intraseasonal entrainment-rate anomaly (black) and its three components, MLD tendency (orange), vertical pumping at the mixed layer base (cyan), and the lateral induction (pink), averaged in the EIO box. (b) MJO composites of 30–105-day entrainment rate and its three components. (c) Bimonthly STD of intraseasonal entrainment rate (black) and MLD tendency (orange). All results are derived from MR – NoSTRS (10^{-6} m s^{-1}).

entrainment rate, is used to quantify both the frequency and magnitude of the water mass exchange at the mixed layer base. Because of the large entrainment events by MJOs, the annual entrainment rate averaged in the EIO box increases from 435 m yr^{-1} in NoMJO to 582 m yr^{-1} in MR. In other words, taking into consideration MJO-related forcings can intensify the upward mass flux at the mixed layer base by 34%. Again, the wind stress effect is the leading contributor to ISV of E , showing a STD of $7.91 \times 10^{-6} \text{ m s}^{-1}$ and a correlation of 0.97 with MR (Fig. 6b). Its overall contribution to the total ISV of E is 86.8%. The SWR effect is of secondary importance, with a STD of $3.17 \times 10^{-6} \text{ m s}^{-1}$ and a correlation of 0.70, and its contribution to MR is 25.4% (Fig. 6c). In comparison, the effects of wind speed and precipitation are relatively weak, and the precipitation effect is negative (Figs. 6d,e).

Wind stress can affect the entrainment rate through three different physical processes, as represented by the three RHS terms in Eq. (4). First, wind stress causes MLD changes through turbulent mixing and ocean dynamical processes, as discussed in section 3. The MLD change, as quantified in Fig. 7a by the MLD tendency term $\partial h_m / \partial t$ (orange), is of the largest STD ($8.73 \times 10^{-6} \text{ m s}^{-1}$), and its overall contribution to E (black) reaches 92.8%. The positive $\partial h_m / \partial t$ caused by wind forcing in phases 1–3 (Fig. 7b) gives rise to the maximum MLD in phase 4 (Fig. 5a). Second, the vertical pumping term w_{mb} induced by wind stress can directly influence the entrainment by carrying water mass across the mixed layer base. The amplitude of intraseasonal w_{mb} at the mixed layer base accounts

for half of that of E (Fig. 7a), but its overall contribution is only 4.3% because of the mismatch in timing (Fig. 7b). Both $\partial h_m / \partial t$ and E show the peak in phase 2 and the trough in phase 5. However, the peak of w_{mb} occurs in phase 8, and its trough is found in phase 3, which may be related to the downwelling Kelvin waves. Third, the lateral induction term $\mathbf{u}_{mb} \cdot \nabla h_m$ induced by wind stress can also cause entrainment through horizontal currents across MLD slope. The contribution of $\mathbf{u}_{mb} \cdot \nabla h_m$ is the smallest, with the STD of $1.81 \times 10^{-6} \text{ m s}^{-1}$ (Fig. 7a). The intraseasonal entrainment rate variability also shows prominent seasonality, which mainly arises from the MLD tendency term. It achieves the strongest intensity in March–April with the STD of $>10 \times 10^{-6} \text{ m s}^{-1}$ and is weakest in July–August ($<5 \times 10^{-6} \text{ m s}^{-1}$) (Fig. 7c). This seasonality is consistent with MJO seasonality that MJO is more energetic in the winter monsoon season than in the summer monsoon season (Zhang and Dong 2004).

Not all entrainment events can effectively contribute to the obduction of subsurface water. Theoretically, there are two types of entrainment, i.e., ineffective entrainment and effective entrainment. The former type refers to the case that the water entrained into the mixed layer was once within the mixed layer and thereby in contact with the atmosphere during the previous 1-yr period. This amount of water is not originated from the permanent pycnocline, but rather from the recently detrained mixed layer water. The latter type denotes the entrainment of the nutrient-rich, cold permanent pycnocline water (Qiu and Huang 1995), which carries the properties determined by climate conditions of its ventilation region in

FIG. 8. (a) Percentage of effective entrainment relative to the total entrainment, and the (b) annual, (c) summertime (May–October), and (d) wintertime (November–April) obduction rates (m yr^{-1}). All properties are averaged from 2002 to 2014 as derived from HYCOM MR. In (b)–(d), obduction-rate values smaller than 20 m yr^{-1} are left white.

extratropics. The annual obduction rate is defined as the accumulation of effective entrainment over one year,

$$O = \frac{1}{T} \int_{T_s}^{T_e} E dt, \quad (5)$$

where $T = 1$ year, and T_s and T_e are the starting time and ending time of the effective entrainment period that are usually determined by the seasonal cycle of MLD. Considering the ISV induced by MJO, entrainment may operate differently from the traditional view, that is, showing multiple effective entrainment periods throughout a year. Therefore, the annual obduction rate is modified as

$$O = \frac{1}{T} \sum_{n=1}^N \int_{T_{ns}}^{T_{ne}} E dt, \quad (6)$$

where N is the number of effective entrainment periods, and T_{ns} and T_{ne} are the starting time and ending time of the n th period. To ascertain the effective entrainment, we perform reversed trajectory tracing for the entrained water particles for over one year. Particles are released at every 0.25° grid point of the EIO with 3-day intervals in the entrainment phase ($E > 0$). For a given station at a given time with $E > 0$, the particle is released at the local mixed layer base, and then is traced

backward using three-dimensional velocity fields of HYCOM with a time step of 0.5 day, which are obtained through linear interpolation from the 3-day mean HYCOM outputs. In each time step during tracing, if the trajectory is intersected by the mixed layer upstream, it is labeled ineffective and does not integrate into the obduction rate. If the water particle is always below mixed layer throughout the previous 1-yr period, it is classified as effective entrainment.

As shown in Fig. 8a, effective entrainment only accounts for a small fraction ($<20\%$) of the total entrainment events. This indicates that more than 80% of the entrained water was detrained from the surface mixed layer during the previous 1-yr period. In the EIO, obduction of subsurface water mainly takes place in a small region near the equator between 77° and 92°E (Fig. 8b). The maximal obduction rate averaged from 2002 to 2014 is $\sim 120 \text{ m yr}^{-1}$ at $\sim 83^\circ\text{E}$. The annual obduction rate integrated over the EIO region is 1.36 Sv ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) in HYCOM MR. In addition to the annual obduction rate, we also compute the obduction rates in summer and winter separately taking into consideration the effects of monsoon. Here, summer is defined as the summer monsoon period of May–October, while the winter is defined as November–April (Duncan and Han 2009; Han et al. 2007). That is, the accumulated effective entrainment in May–October is calculated as the summertime obduction rate, and the entrainment for November–April is calculated as the wintertime obduction rate. Unexpectedly, we find that 84.6% of obduction occurs during summer monsoon periods (Fig. 8c), with only a small portion occurring during winter monsoon periods (Fig. 8d). This is primarily determined by the seasonal cycle of MLD (Fig. 2c). The deep MLD in response to the summer monsoon is favorable for harvesting water from the permanent pycnocline, while the relatively shallow MLD in the winter monsoon season makes it difficult for effective entrainment to occur.

Next, we examine the impact of MJOs on the obduction of subsurface water in the EIO. Figure 9 compares the annual duration time for effective entrainment and summertime obduction rate averaged over 2002–14 between MR and NoMJO. The two simulations generally achieve similar spatial patterns in the EIO. It is interesting to see that without the MJO-related forcings, the duration time of effective entrainment is extended as compared with the case in MR (Figs. 9a,b). Correspondingly, the summertime obduction is larger in NoMJO (Figs. 9c,d). The EIO-integrated summertime obduction rate derived from NoMJO is 2.45 Sv (Table 2), more than twofold that from MR (1.15 Sv). In other words, our model results suggest that MJOs act to greatly suppress the summertime obduction of subsurface water in the EIO, reducing the summertime obduction rate by as large as 53%. Table 2 specifies the effects of different forcing processes of the MJO on the summertime obduction rate. It is seen that all the MJO-related forcing effects, including wind stress, wind speed, SWR, and precipitation, can suppress the obduction of subsurface water in the EIO, and the wind stress effect is the dominant contributor as expected. The SWR effect plays a secondary role but with a much smaller contribution.

To understand how the ISV of MLD affects obduction in the EIO, we compare the probability density distributions of MLD

FIG. 9. The accumulation time of effective entrainment (days) from (a) MR and (b) NoMJO. Also shown are the summertime obduction rate (m yr^{-1}) from (c) MR and (d) NoMJO. All properties are for 2002–14. Regions with obduction rates smaller than 20 m yr^{-1} are left white.

derived from MR and NoMJO (Fig. 10a). MLD is primarily concentrated in the moderate range regardless of MJO-related forcings. 56% and 62% of MLD values from MR and NoMJO fall into the range of 20–50 m, respectively. By contrasting MR and NoMJO results, it is found that taking into account the MJO forcings can produce more extreme MLD values, i.e., $>55\text{-m}$ and $<15\text{-m}$ values. Figures 10b and 10c suggest that MR and NoMJO show rather similar seasonal variations in MLD, reaching the deepest in July ($\sim 50\text{m}$) and the shallowest in March ($\sim 30\text{m}$). The primary difference arises from the intraseasonal time scale. The intraseasonal MLD variability is much stronger in MR than in NoMJO, owing to the MJO-related forcing effects. The STD of intraseasonal MLD in NoMJO is merely $\sim 1.2\text{--}1.9\text{m}$ (Fig. 10c), much smaller than that of $4.0\text{--}9.1\text{m}$ in MR (Fig. 10b). Moreover, MR shows much stronger intraseasonal MLD variability in spring (April–May) than in other seasons (Fig. 10b), which is consistent with SODA results that the monthly STD of intraseasonal MLD reaches the maximum in April–May and minimum in July–August (figure omitted). The difference in STD of intraseasonal MLD between MR and NoMJO also reaches the maximum in April–May (figure not shown). Therefore, the stronger intraseasonal MLD variability in spring may strongly affect the subsequent obduction in the summer monsoon season by rendering entrainment events ineffective. In the existence of intraseasonal MLD deepening, the possibility becomes much larger that the water entrained into the mixed layer in summer originates from the spring mixed layer (recall Figs. 9a,b).

Based on the analysis presented above, a conceptual schematic is postulated to demonstrate how MJO-induced intraseasonal MLD variability suppresses the summertime obduction in the

EIO. For the cases without the MJO forcing effect, the obduction can be approximately described by the canonical model based on a smooth seasonal cycle (Qiu and Huang 1995), except that obduction occurs in summer featuring a region dominated by the Indian summer monsoon. As shown in Fig. 11a, although mixed layer entrainment takes place over a wide time window of one year, only a fraction of this entrained water, as indicated by the trajectories between A and B, comes from the permanent pycnocline that is not influenced by mixed layer and is labeled as effective entrainment. With the MJO-forcing effect taken into account, strong intraseasonal MLD fluctuations in spring are superimposed on the seasonal cycle, and the trajectories also show strong intraseasonal displacements in the vertical location (Fig. 11b). One can see that detrainment and entrainment take place alternately corresponding to intraseasonal shoaling and lowering of MLD. More importantly, some water trajectories between B and C are intersected by the deepened mixed layer in spring, rendering the entrainment ineffective. In other words, the bound of the effective entrainment window has to move from B to C because of the MJO-driven MLD variability. As a result, only the entrained water between A and C can be considered as effective entrainment and contribute to the summertime obduction rate. Therefore, the overall role played by the MJO forcing is to shorten the time window for effective entrainment and reduce the obduction of subsurface water in the EIO. According to our model simulations, this reduction effect can be as large as 53%.

5. Conclusions

The MLD is a key parameter for air–sea interaction, and its variability gives rise to detrainment/entrainment processes and

TABLE 2. The integrated summertime obduction rate in the EIO averaged from 2002 to 2014 from the MR, NoMJO, MJO effect, wind stress effect, wind speed effect, SWR effect, and precipitation effect. All of the units are Sverdrups ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$).

MR	NoMJO	MJO effect	Wind stress effect	Wind speed effect	SWR effect	<i>P</i> effect
1.15	2.45	−1.3	−1.24	−0.04	−0.09	−0.02

FIG. 10. (a) The probability distribution of MLD of the EIO box from the MR (black) and NoMJO (cyan). Also shown are the monthly climatological MLD (blue) and standard deviation of intraseasonal MLD variability (gray shading) in the EIO box from (b) MR and (c) NoMJO.

contributes to the subduction/obduction of water masses. Strong entrainment usually implies a significant impact of the thermocline water on the mixed layer temperature/salinity and therefore a stronger coupling between the atmosphere and ocean dynamical processes. Particularly, the obduction near the equator represents the upward mass flux from the permanent pycnocline that conveys the climatic signals originating from the subtropics and higher latitudes and modulates the tropical climate on decadal and longer time scales (e.g., Gu and Philander 1997; Zhang et al. 1998; Li et al. 2012). The EIO is a key region of the initiation and development of the MJO, the dominant mode of the tropical atmospheric intraseasonal oscillations, and exhibits strong intraseasonal variability in response to MJOs. However, mechanisms of intraseasonal MLD variability in the EIO and impacts on the obduction of subsurface water were rarely investigated before. This study provides a comprehensive investigation of this issue.

Measurements by RAMA buoys near the Indian Ocean equator have documented prominent ISV of MLD with periods of 30–105 days. Comparisons with in situ observations reveal that HYCOM can well represent the MLD distribution and its variability in the EIO. Intraseasonal MLD variability is prominent in the EIO, where the STD of 30–105-day MLD reaches ~ 14 m. Six parallel HYCOM experiments are performed to assess the relative importance of MJO-related

forcing effects, including wind stress, wind speed, SWR, and precipitation. It is demonstrated that wind stress is the major driver of the intraseasonal MLD variability in the EIO, which can explain 83.7% of the total ISV. In the active stage of MJOs (phase 3 of the RMM index), the increased surface wind stress causes enhanced turbulent mixing, and westerly wind anomalies drive upper-ocean convergence. The two processes work mutually to cause MLD deepening in the EIO. Besides, the wind speed and SWR effects also contribute to the MLD deepening through sea surface buoyancy forcing, but their roles are secondary. In the EIO, the composite MLD is deep during phases 3–5, reaching the largest depth in phase 4, and is shallow in phase 8. MJO-induced intraseasonal MLD deepening gives rise to high-frequency entrainment events of subsurface water, leading to an increase of the annual entrainment rate of the EIO by 34%. The HYCOM results show that MJO-related wind stress is the primary driver for the intraseasonal entrainment, while the contribution of SWR is secondary. Among the three components of entrainment, the wind stress-forced MLD tendency term plays a more important role than the vertical pumping term and the lateral induction term.

In the EIO, only a small percentage of these entrainment events ($<20\%$) can effectively contribute to the annual obduction rate of 1.36 Sv, with the maximal obduction rate $\sim 120 \text{ m yr}^{-1}$. As determined by the seasonal variation of

FIG. 11. The conceptual model of obduction for the cases (a) without and (b) with the MJO effects. The black curves indicate the instantaneous location of the mixed layer base. The slanted and curved thick horizontal arrows in (a) and (b), respectively, denote the trajectories of water mass entering into the mixed layer. The abscissa denotes both time and position of the water parcel entering into the mixed layer. The black dashed lines isolate the time windows of the detrainment and entrainment phases, and the blue and red dashed lines present the time window of effective entrainment for the cases without and with MJOs, respectively. Trajectory A corresponds to the ending of effective entrainment, and trajectories B and C correspond to the starting of effective entrainment for the cases without and with MJOs, respectively.

MLD, 84.6% of the obduction occurs in the summer monsoon season. Taking into account the MJO effects can greatly reduce the summertime obduction in the EIO by $\sim 53\%$. All MJO-related forcing effects act to suppress the obduction of subsurface water, with the wind stress effect as the dominant contributor. Further analysis indicates that MJOs give rise to prominent intraseasonal MLD variability, which achieves the largest intensity in spring (April–May). These strong intraseasonal MLD deepening events in spring greatly affect the subsequent summertime obduction by rendering entrainment events ineffective. A conceptual schematic is postulated to depict how MJO-induced springtime intraseasonal MLD deepening events narrow down the time windows of effective entrainment and thereby suppress the obduction rate. The present study has clarified the mechanisms of intraseasonal MLD variability in the EIO region and highlighted its noticeable influence of suppressing the summertime obduction of subsurface water. There are many interesting problems to be addressed, such as the interannual and event-by-event variabilities of the ISV. These variabilities may cause corresponding modulations of obduction on different time scales and thereby exert impacts on SST. This topic warrants future exploration.

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