The Sloshing and Diapycnal Meridional Overturning Circulations in the Indian Ocean

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ABSTRACT: The meridional overturning circulation (MOC) seasonality in the Indian Ocean is investigated with the ocean state estimate product ECCO v4r3. The vertical movements of water parcels are predominantly due to the heaving of the isopycnals all over the basin except off the western coast. Aided by the linear propagation equation of long baroclinic Rossby waves, the driving factor determining the strength of the seasonal MOC in the Indian Ocean is identified as the zonally integrated Ekman pumping anomaly, rather than the Ekman transport concluded in earlier studies. A new concept of sloshing MOC is proposed, and its difference with the classic Eulerian MOC leads to the so-called diapycnal MOC. The striking resemblance of the Eulerian and sloshing MOCs implies the seasonal variation of the Eulerian MOC in the Indian Ocean is a sloshing mode. The shallow overturning cells manifest themselves in the diapycnal MOC as the most remarkable structure. New perspectives on the upwelling branch of the shallow overturn in the Indian Ocean are offered based on diapycnal vertical velocity. The discrepancy among the observation-based estimates on the bottom inflow across 32°S of the basin is interpreted with the seasonal sloshing mode. Consequently, the “missing mixing” in the deep Indian Ocean is attributed to the overestimated diapycnal volume fluxes. Decomposition of meridional heat transport (MHT) into sloshing and diapycnal components clearly shows the dominant mechanism of MHT in the Indian Ocean in various seasons.

KEYWORDS: Indian Ocean; Ekman pumping/transport; Rossby waves; Vertical motion; Diapycnal mixing; Seasonal variability

1. Introduction

The meridional overturning circulation (MOC) in the Indian Ocean is regarded as a crucial deep upwelling limb of the global MOC, for it takes almost 40% of the global deep upwelling with less than 20% area of the World Ocean (Huussen et al. 2012). It exhibits a remarkable feature of seasonal variability under influence of monsoons. This seasonally varying forcing, together with the limited extent in the Northern Hemisphere, makes a unique MOC pattern in the Indian Ocean compared with the other two major basins. The most remarkable feature of the Indian Ocean MOC (IMOC) is the prevailing cells spanning from the upper ocean to the abyssal ocean with directions reversed in winter and summer (e.g., Gartnericht and Schott 1997; Lee and Marotzke 1998; Schott and McCreary 2001; Wang et al. 2014; Han and Huang 2020). Lee and Marotzke (1998) applied a decomposition on the IMOC streamfunction and concluded that the Ekman flow plus its barotropic compensation explained a large part of the variation. Wang et al. (2014) related the upwelling (downwelling) branch of the IMOC cells in boreal winter (summer) with the negative (positive) wind stress curl in the zonal band around 20°S. However, some fundamental questions remain unanswered in those studies, such as the driving force of the barotropic compensation flow in the deep ocean, and how to

explain the other branch of the overturn in the north Indian Ocean with the same argument (wind stress curl) as for its southern branch.

Schott and McCreary (2001) interpreted this variation as an adiabatic process of sloshing isopycnals. They argued that this seasonal variability “mostly represents an *adiabatic sloshing* back-and-forth of water masses, and not diabatic, across-isopycnal flow. Thus, it is indicative mostly of changes in heat storage, rather than of cell strength.” However, no supporting evidence was provided in their paper. The baroclinic Rossby and Kelvin waves are found active within the latitudes of seasonal IMOC cells (20°S–20°N), and the isopycnals or isotherms are indeed perturbed notably by them (e.g., McCreary et al. 1993; Masumoto and Meyers 1998; Birol and Morrow 2001; Wang et al. 2001, White et al. 2004; Subrahmanyam et al. 2009; Vialard et al. 2009; Gnanaseelan and Vaid 2010; Rao et al. 2010; Johnson 2011; Suresh et al. 2013; Nagura 2018; Menezes and Vianna 2019; Zhang and Han 2020). Moreover, observations by satellite altimetry and Argo floats found semiannual Kelvin waves and Rossby waves dominate the vertical displacement of sea level and isopycnals on the equator, while annual Rossby waves are predominant signals off the equator (e.g., Han et al. 1999; Masumoto and Meyers 1998; Birol and Morrow 2001; Rao et al. 2010; Johnson 2011; Nagura 2018; Huang et al. 2018a,b; Zanowski and Johnson 2019). How these waves are linked with the Eulerian IMOC is another question to be investigated in this study.

As components of IMOC, the shallow overturning cells and their variability in the Indian Ocean have been extensively studied (e.g., Schott and McCreary 2001; Miyama et al. 2003;

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Lee 2004; Schott et al. 2004; Schott 2005; Schoenefeldt and Schott 2006; Schott et al. 2009; Han et al. 2014; Nagura and McPhaden 2018; Meng et al. 2020). However, as was addressed in Schott and McCreary (2001) and has hardly changed for decades, the lack of quantitative evidence from observations has left the science reliant on model results. Regarding the hot spots where the water parcels subduct and upwell, representation first shown by Schott et al. (2002) and later modified (Schott et al. 2004; Schott 2005; Schoenefeldt and Schott 2006; Schott et al. 2009) identified several areas that vary from winter to summer. In this study, based on the analysis of diapycnal vertical velocities, a solution that includes the shallow overturning cells extending to 30°S in the Indian Ocean is obtained, and a new perspective on the upwelling branch of the shallow overturn is offered.

Extensive assessments on the deep IMOC have been conducted across 32°S by the transbasin hydrographic sections (e.g., Fu 1986; Toole and Warren 1993; McDonagh et al. 2008), inverse methods and assimilations (e.g., Toole and Warren 1993; Robbins and Toole 1997; Macdonald 1998; Ganachaud et al. 2000; Bryden and Beal 2001; Sloyan and Rintoul 2001; Ferron and Marotzke 2003; McDonagh et al. 2008; Hernández-Guerra and Talley 2016), and GCM solutions (e.g., Wacongne and Pacanowski 1996; Garternicht and Schott 1997; Lee and Marotzke 1997, 1998; Zhang and Marotzke 1999). Regarding the strength of deep IMOC which is normally defined as the maximum of bottom-up integrated transport across 32°S (Huussen et al. 2012), notable discrepancies exist not only between the hydrographic data and the GCM outputs, but also among the different hydrography and the inverse solutions based on them [see Ferron and Marotzke (2003) for a discussion on the issue]. In this study, discrepancy among hydrographic estimates is interpreted from the perspective of sloshing motions.

The second question associated with the Indian Ocean's deep overturn is the overestimated diapycnal diffusivity in the abyssal ocean (Schott and McCreary 2001). The diffusivity is derived with the vertical advective–diffusive balance and the stratification of the density field, i.e., the “abyssal recipes” by Munk (1966). Ganachaud et al. (2000) estimated an averaged diffusivity of $8.6 \pm 4 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ in the north Indian Ocean box based on a deep inflow of $\sim 10 \text{ Sv}$ ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$). Schott and McCreary (2001) stated that an amount of vertical diffusion of about $4 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ is required to maintain the currently estimated strength of deep circulation. Huussen et al. (2012) obtained a variety of diapycnal turbulent diffusivities from 1 to $10 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ for different estimates on deep overturn. All these estimates are much higher than the value of $1 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ for the global ocean (the so-called “pelagic diffusivity”; Munk and Wunsch 1998). Many researchers attempted to attribute this high diffusivity to the Indian Ocean's rough topography, and much elevated diffusivity was indeed found near the fracture zone (MacKinnon et al. 2008). However, it is still difficult to acquire a basin-averaged diffusivity as high as the abovementioned values. In fact, measurements from lowered ADCP shear and CTD strain profiles found that “average diffusivities are $(0.1\text{--}0.2) \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ in the upper 3000 m of the Indian Ocean” (Kunze et al. 2006).

In a global pattern of diapycnal diffusivity based on in situ measurements, the Indian Ocean, either of the upper 1000 m or of the full depth (particularly north of 8°S, the region Ganachaud et al. (2000) used for their estimate), does not exhibit noticeably larger values than other basins (Waterhouse et al. 2014). From the energy perspective, “there is consistently too little internal wave dissipation to sustain the Indian Ocean MOC at deeper levels,” and a factor of as large as 8 has to be assumed to “achieve consistency with the overturning solutions that have a strength of about 10 Sv” (Huussen et al. 2012). In this study, new insight into understanding this “missing mixing” problem is offered.

Mechanisms of the meridional heat transport (MHT) in the Indian Ocean have been explored in previous literature with various decomposition techniques (e.g., Lee and Marotzke 1998; Ganachaud and Wunsch 2003; Ma et al. 2019). The frequently used one, often referred to as the barotropic–baroclinic–horizontal (BBH) decomposition” (Macdonald and Baringer 2013), separates MHT into components associated with the net volume transport, overturn, and gyre, respectively. The overturning component is found dominant in the MHT of the Indian Ocean (Lee and Marotzke 1998). In a later section, a novel decomposition is proposed from the perspective of sloshing motion. The new method can reveal how much heat is transported meridionally by the adiabatic and diabatic processes of the overturning circulation, respectively.

This study starts from exploring seasonally varying IMOC dynamics from the perspective of isopycnal movement. It then applies the new solutions to interpret a variety of questions concerning IMOC, such as those above. This paper is set out as follows. Section 2 introduces the data this study relies on and the method. In section 3, the Eulerian vertical velocity is compared with the heaving rate of isopycnals. The heaving signals of isopycnals with unique signatures of Rossby and Kelvin waves are analyzed. In section 4, with a new concept of diapycnal or diabatic MOC, applications of the sloshing perspective in the shallow overturn, the missing mixing in the deep Indian Ocean, and the mechanism of MHT in the basin are discussed. A summary is given in section 5. Concerning the terminologies, the vertical undulation of isopycnals is termed “heaving,” and the vertical velocity with which the isopycnals move up and down is referred to as the “heaving rate” [see Huang (2020) for a thorough discussion]. In comparison, “sloshing” refers to water parcels' back-and-forth movement in the horizontal (J. McCreary 2020, personal communication).

2. Data and methodology

The data used in this study are the Estimating the Circulation and Climate of the Ocean State Estimate version 4 release 3 (ECCO v4r3) (Forget et al. 2015; Fukumori et al. 2017). In this ocean reanalysis dataset based on the MITgcm (Marshall et al. 1997; Adcroft et al. 2004), Lagrange multipliers are used (Wunsch and Heimbach 2007) to enforce global and local satisfaction of basic conservation rules (heat, freshwater, momentum, energy, etc.) for the period of 24 years (January 1992–December 2015). In the horizontal plane, ECCO

adopts a nonuniform llc90 (latitude–longitude–cap) gridding system, which maps Earth using five faces including an Arctic face and four mostly latitude–longitude sectors, with a nominal resolution of 1° ; in the vertical direction, ECCO has rescaled height coordinates with 50 vertical levels and partial cell representation of bottom topography (Campin et al. 2008). The K -profile parameterization vertical mixing scheme of Large et al. (1994) is employed for realistic simulation of near-surface mixing processes. The Redi (1982) isoneutral mixing scheme and the Gent and McWilliams (1990) parameterization are used to represent the mixing effects of mesoscale eddies. More information on ECCO can be found at <https://ecco-group.org>. ECCO's low horizontal resolution might be an issue for coastal regions in the Indian Ocean (Webber et al. 2014), but suffices for studies that deal with the large-scale Eulerian-mean motions in the open ocean. The ECCO product and its other version (GECCO) have been adopted in obtaining IMOC streamfunctions (Wang et al. 2012, 2014; Han and Huang 2020). Comparisons of ECCO with various observation datasets in the Indian Ocean have illustrated its fidelity in representing the seasonal variability (Halkides and Lee 2011; Webber et al. 2014). The climatological sea level anomaly (SLA) of ECCO compares favorably with the satellite-altimetry-derived SLAs in the Indian Ocean over 12 months (Rao et al. 2010).

It is common practice to obtain a MOC streamfunction by vertically integrating the zonally integrated meridional flow component across the basin. This method requires the net volume transport across the zonal section to vanish. However, for the Indian Ocean, the existence of an open boundary to the Pacific welcomes a net inflow with significant volume flux (~ 10 Sv). The Indonesian Throughflow (ITF) and its subsequent net southward transport break the condition necessary for a uniquely defined IMOC streamfunction south of $\sim 10^\circ\text{S}$. Attempts have been made to minimize the influence of ITF on the IMOC streamfunction. For example, Wang et al. (2014) integrated the mass streamfunction “from the ocean bottom upward such that ITF's influence is confined within the upper ocean since most of the ITF transport is present in the upper around 400 m.” Han and Huang (2020) used the Helmholtz decomposition to eliminate the divergent component associated with ITF from the zonally integrated flow field and obtained a consistent solution of IMOC.

Besides the common practice of integrating the meridional flow vertically, the streamfunction can also be obtained by integrating the vertical velocity component. After all, the streamfunction ψ is defined as the function of both V and W as follows:

$$V(y, z) = -\frac{\partial \psi(y, z)}{\partial z}, \quad W(y, z) = \frac{\partial \psi(y, z)}{\partial y}, \quad (1)$$

where V and W represent the zonally integrated Eulerian velocity components defined as

$$V(y, z) = \int_{x_w}^{x_e} v(x, y, z) dx, \quad W(y, z) = \int_{x_w}^{x_e} w(x, y, z) dx, \quad (2)$$

in which x_w and x_e indicate the locations of the western and eastern boundaries of the zonal section across the basin, and are functions of y in general; v and w are the 3D Eulerian velocity components in the meridional and vertical directions. The zonally integrated continuity equation gives

$$\frac{\partial V}{\partial y} + \frac{\partial W}{\partial z} = u_w(y) - u_e(y), \quad (3)$$

where u_w and u_e denote the zonal velocity component at x_w and x_e , respectively. It is readily seen that the definition of ψ is meaningful only if the RHS of Eq. (3) vanishes. This is the case for zonal sections with wall boundaries at both ends, but is not the case for those open to ITF. A simple way to identify the uncertainties ITF brings to determining IMOC is comparing the two streamfunctions obtained by V and W , respectively, which are written as ψ^V and ψ^W . The two sets of IMOC solutions for all the climatological months are shown in Figs. 1 and 2. The two solutions' overall patterns resemble each other, both with notable seasonal variations, a finding consistent with previous results (e.g., Lee and Marotzke, 1998; Wang et al. 2012, 2014; Han and Huang 2020). This agreement proves the approximate equivalence of V and W in deriving ψ even with the influence of ITF. Figure 3 shows those places where ψ^V and ψ^W do not agree with each other in different months. The domain where the two solutions most disagree is concentrated in the upper few hundred meters to the south of 10°S . It reflects the influence of ITF, which enters the Indian Ocean from 10° to 20°S with the bulk of its transport present in the upper 400 m (Liu et al. 2005) and heading southward after reaching the western part of the Indian Ocean (e.g., Schott et al. 2009; Wang et al. 2014). Though this study does not concern the uncertainty of IMOC due to ITF, Fig. 3 suggests where the most considerable uncertainties reside. On the other hand, the above comparisons indicate that the two solutions, ψ^V and ψ^W , performs equally well for investigating the seasonal variations of IMOC. It is ψ^W that will be adopted for the analysis in the following sections.

This paper only uses ECCO's climatological mean data because it suffices for studying the dynamics and mechanisms of seasonal variation in the Indian Ocean. The potential densities with a reference pressure of 0, 2000, and 4000 dbar (referred to as σ_0 , σ_2 , and σ_4 , respectively) are calculated with the Gibbs SeaWater (GSW) Oceanographic Toolbox (McDougall and Barker 2011). This toolbox contains the up to date TEOS-10 subroutines for evaluating the thermodynamic properties of pure water and seawater (www.teos-10.org). The depth of a specific isopycnal is determined by interpolation with densities at the 50 standard depth levels of ECCO. The heaving rate, i.e., the vertical displacement rate of isopycnals, written as $w_{\text{iso}}^{\text{heave}}$, is calculated with isopycnal depths as follows:

$$w_{\text{iso}}^{\text{heave}}(\sigma, m) = \frac{1}{T} [h(\sigma, m) - h(\sigma, m-1)], \quad m = 1, 2, \dots, 12, \quad (4)$$

where $h(\sigma, m)$ refers to the depth of the isopycnal σ in the m th month, from January to December, and $h(\sigma, 0)$ is for December; T is taken to be 30 days here. Since the monthly

FIG. 1. The climatological IMOC streamfunctions ψ^V (integrating V bottom-up) for each month calculated by ECCO. The clockwise (anticlockwise) cells in dull red (steel blue) are positive (negative). The two yellow dashed lines indicate the locations of the open eastern boundary where ITF enters the Indian Ocean.

mean diagnostics of ECCO are defined in the middle of each month, w_{iso} in Eq. (4) is thus defined at the first day of each month. The potential density in Eq. (4) may refer to σ_0 , σ_2 , or σ_4 , but it will be shown in later section that the heaving rates derived from σ_0 , σ_2 , and σ_4 are close to one another.

3. Vertical motions in the Indian Ocean

a. Eulerian vertical velocity versus heaving rate of isopycnals

For adiabatic motions of the stratified fluids, density is materially conserved. The isopycnals remain as material surfaces when moving with the fluid, i.e., fluid parcels initially found on an isopycnal will always stay on that isopycnal. This condition yields

$$w_{\text{iso}} = \frac{\partial h}{\partial t} + u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y}, \quad (5)$$

where w_{iso} denotes the vertical velocity of fluid parcels associated with the isopycnals, including the heaving of the isopycnal (first term of RHS) and along-isopycnal movement on the tilted isopycnal (the last two terms of RHS). In this case, the vertical motion is fully explained by w_{iso} . However, when there are diabatic processes, the situation becomes different, which is schematized in Figs. 4a and 4b. Suppose the material surface (marked by a star) whose density is ρ_0 at initial time t_0 moves up to a new level after a time interval Δt when there is a downward heat flux or heating from above (Fig. 4a), the density of this material surface is reduced, which makes the isopycnal of ρ_0 at time $t_0 + \Delta t$ a bit deeper than the material surface which is marked by the star. In this case, when the

heaving rate of the isopycnal ρ_0 , i.e., w_{iso} , is calculated as per the positions of the isopycnal, it is no longer equal to the Eulerian vertical velocity (written as w_{Eul}), which tracks the movement of water parcel. Their difference ($w_{\text{Eul}} - w_{\text{iso}}$) provides a measure of the diabatic processes across this material surface, which is written as w_{dia} . The physical implication of this positive “diapycnal vertical velocity” w_{dia} in the case of Fig. 4a can be interpreted as the upward volume flux “across” a specific isopycnal in a rate proportional to the downward heat flux. If with an upward heat flux (subduction or cooling from above, Fig. 4b), the density of the material surface (also marked with a star) is increased due to cooling, which makes the new depth of isopycnal of ρ_0 shallower than that of the material surface at time $t_0 + \Delta t$. It looks like some volume of water “crosses” the isopycnal ρ_0 downward, which is consistent with a negative w_{dia} . Therefore, this kind of diapycnal or cross-isopycnal processes in the ocean can be measured by the difference of w_{Eul} and w_{iso} . Note, there is no “relative movement” between the material surface (which is tracked by the water parcel only, i.e., the black star in Fig. 4) and the isopycnal. Actually, the “isopycnal” is just different “name” of the material surface at different time. The two types of vertical velocities, w_{Eul} and w_{iso} , track a specific water parcel and a specific name (potential density in this case), respectively.

As shown in Eq. (5), w_{iso} includes two components, associated with vertical displacement of the isopycnals and along-surface flow on tilted isopycnals, respectively, which are written as follows:

$$w_{\text{iso}}^{\text{heave}} = \frac{\partial h}{\partial t}, \quad w_{\text{iso}}^{\text{tilt}} = u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y}. \quad (6)$$

FIG. 2. As in Fig. 1, but for IMOC solution ψ^W (integrating W southward from the northern boundary of the Indian Ocean).

The combined effects of the two components on w_{iso} are illustrated in Figs. 4c and 4d for the tilted isopycnals. To quantify the relative importance of these effects, $w_{\text{iso}}^{\text{heave}}$ is calculated with the method introduced in the previous section, and $w_{\text{iso}}^{\text{tilt}}$ is evaluated with the derived depths of isopycnals and the horizontal flow fields available in ECCO. Two typical months and two isopycnals are chosen to demonstrate the

comparison (Fig. 5). Evidently, the heaving motion dominates the vertical velocity component of flow along isopycnals all over the basin except at the upper ocean near the western boundary.

According to the above discussion, a close relationship between w_{Eul} and w_{iso} can be expected in the largely adiabatic domain of the ocean. The heaving motion of isopycnal,

FIG. 3. As in Fig. 1, but for the difference of Figs. 1 and 2, i.e., $\psi^V - \psi^W$.

FIG. 4. Schematic relationships among various vertical velocities concerning isopycnals. The gray solid lines represent isopycnals, a star represents a water parcel, and a thick dashed line represents the new position of isopycnal ρ_0 subject to combined effect of heaving and diapycnal heat flux. A dashed arrow represents vertical movement of water parcel due to heaving of isopycnals, a solid arrow represents along-isopycnal movement of water parcel, a red arrow shows the direction of heat flux, and a blue arrow shows the direction of horizontal flows. See text for detailed explanations.

$\sigma_0 = 27.4 \text{ kg m}^{-3}$, is chosen for examining this relationship. The depth of the isopycnal is of course, not uniform over the basin. The annual-mean depth of isopycnal $27.4\sigma_0$ north of 20°S of the Indian Ocean ranges from 871 to 1143 m, with a mean of ~ 977 m and a standard deviation of ~ 47 m. The reason to choose this isopycnal is that the upwelling and downwelling branches of the seasonal IMOC are generally stable at that depth range (Fig. 1 or Fig. 2). Figure 6 shows the climatological-mean w_{iso} throughout the year. The Eulerian vertical velocity w_{Eul} , at a standard depth level closest to the mean depth of $27.4\sigma_0$, is shown in Fig. 7. Though the patterns of both w_{iso} and

w_{Eul} change dramatically throughout the year and they are not evaluated at exactly the same depths, their month-to-month resemblance is striking. With the same color scale setting, Fig. 8 shows how generally small their differences are compared to themselves.

b. Heaving isopycnals and planetary waves

A variety of other isopycnals are also examined, and the conclusion remains more or less the same except that the differences become larger at shallower depths. These results imply that the isopycnal heaving motions account for a large

FIG. 5. The two components of the vertical velocity (m s^{-1}) associated with isopycnals, $w_{\text{iso}}^{\text{heave}}$ (first and third columns) and $w_{\text{iso}}^{\text{tilt}}$ (second and fourth columns). Two months [(top) January and (bottom) July] and two isopycnals [(left two columns) $26.5\sigma_0$ and (right two columns) $27.4\sigma_0$] are chosen for demonstration. The mean depths of both isopycnals north of 20°S are indicated in the parentheses. Positive (negative): upward (downward). Data: ECCO r4v3.

FIG. 6. The climatological-mean heaving rate of the isopycnal $27.4\sigma_0$, w_{iso} , through the year. The mean depth of this isopycnal within the domain north of 20°S is ~ 977 m. Data: ECCO v4r3.

portion of the Eulerian vertical movements at least in the open ocean and away from the mixing layers where the adiabatic processes do not dominate. since IMOC streamfunctions are determined by w_{Eul} , as demonstrated in the previous section, and w_{iso} is dominated by the heaving motions of isopycnals, it is sensible to attribute IMOC seasonality to the heaving of isopycnals.

What causes the heaving motions of isopycnals? As briefly introduced in section 1, earlier literature has clarified that isopycnals' perturbations are highly correlated with the Rossby and Kelvin waves that are active in the tropical Indian Ocean. Indeed, signatures of eastward-propagating equatorial Kelvin waves, westward-propagating Rossby waves emanating from

the eastern boundary, and coastal Kelvin waves propagating along the waveguide in the northern Indian Ocean are clearly manifested in Fig. 6.

For revealing the relationship between propagating waves and IMOC seasonality, a series of Hovmöller diagrams are plotted in Fig. 9 along the latitude, 14.9°S , which lies within the band ($20^\circ\text{--}10^\circ\text{S}$) with most substantial vertical motions as indicated in Fig. 1 or Fig. 2. Westward-propagating Rossby waves dominate the signals of sea surface elevation, isopycnal depths, and their heaving motions. It is evident that two downwelling Rossby waves dominate in this latitude from May to September, which results in a deep reaching anti-clockwise overturning cell; at the same time, two upwelling

FIG. 7. The climatological-mean Eulerian vertical velocity w_{Eul} at the standard level 958 m of ECCO.

FIG. 8. The difference of Figs. 6 and 7, i.e., $w_{\text{Eul}} - w_{\text{iso}}$.

Rossby waves start to grow and dominate the vertical motions from December to March, which produces a clockwise overturning cell in the basin. Both upwelling and downwelling phases are characterized by two centers located in the western and eastern half of the basin, respectively. This “double maxima” phenomenon at around 60° and 90°E was first noticed by Masumoto and Meyers (1998) with the altimetry data. Though the presence of the ocean ridge at around 80° – 90°E might block the energy, thus causing the Rossby wave propagating from the east to “break down,” the major mechanisms of this feature are inclined to be the interferences of the localized forcing and free Rossby wave (Wang et al. 2001) or the change in local Ekman pumping across the ridge (Birol and Morrow 2001).

Though IMOC seasonality can be qualitatively explained by the baroclinic planetary waves, those waves are just responses instead of forcings. It is still unclear what factor drives the wave motions of such seasonal pattern.

Before exploring this question with ECCO, it is necessary to examine the wave dynamics ECCO reproduces by analyzing the wave components as revealed by ECCO. Previous studies found the annual and semiannual components dominate the sea level variability in the Indian Ocean (e.g., Clarke and Liu 1993; Morrow and Birol 1998; Schott and McCreary 2001; Birol and Morrow 2001; Polito and Liu 2003; Rao et al. 2010; Johnson 2011; Nagura 2018). The amplitudes and phases of these wave components have been extracted in several ways, such as the least squares fitting with sinusoidal functions (White 1977; Clarke and Liu 1993; Masumoto and Meyers 1998; Polito and Liu 2003), wavelets analysis (Subrahmanyam et al. 2009), Fourier analysis (Rao et al. 2010) and the Finite Impulse Response filter (Polito and Liu 2003; Vaid et al. 2007; Gnanaseelan and Vaid 2010). In this study, the Fourier analysis technique is adopted to process the climatological-mean monthly data of isopycnal depths. It is tested in an idealized experiment in which the signals reconstructed with two

sinusoidal harmonics of annual and semiannual periods agree perfectly well with the original signals.

Figure 10 presents the analysis result for the isopycnal $27.4\sigma_\theta$. The annual component shows largest amplitude in the southern Indian Ocean and the Arabian Sea, exhibiting prominent features of westward-propagating Rossby waves in those regions. The “double maxima” of the Rossby waves in the southern Indian Ocean manifests as two separately propagating Rossby waves with one starting from the eastern boundary and one from the middle of the basin near 70°E (Wang et al. 2001). The semiannual component dominates the equatorial region, especially away from the central basin. The eastern part represents impinging Kelvin waves propagating eastward along the equator and Rossby waves radiating as the Kelvin waves turning poleward and propagating along the Sumatra coast. While the western part reflects the westward propagating Rossby waves impinging on the western boundary and before turning equatorward as coastal, and subsequently the equatorial, Kelvin waves. The pattern of the decomposed wave components inferred from ECCO agrees well with both that from the satellite-altimetry-derived sea level anomaly (Rao et al. 2010), which mirrors the isopycnal depth anomaly (e.g., Gill 1982; Zhang and Han 2020), and that from the depth of the 20°C isotherms derived from XBT data (Masumoto and Meyers 1998).

The vertical structures of the annual and semiannual components are also examined along the 14.9° and 5.4°S transects, respectively (Fig. 11). The westward propagation is prominent for both annual and semiannual components (propagation is directed from higher phase to lower phase). At the same time, in the deep ocean, there is a phase inclination toward the west for both signals. These inclined phases are consistent with those obtained using in situ observations from Argo floats and CTDs (Nagura 2018). It is understood as Rossby wave propagation in the vertical plane, i.e., the phase of the long baroclinic Rossby waves in the deep ocean propagates upward and

FIG. 9. The Hovmöller diagrams at latitude 14.9°S for the surface elevation η and its heaving rate (first row) and depth anomalies of $26.5\sigma_0$ and $27.4\sigma_0$ and their heaving rates (second and third rows). The topography across the basin is appended in the bottom row. The two dashed lines indicate January and July. Data: ECCO v4r3.

westward. In contrast, their energy propagates downward and westward. It is because the vertical propagation of energy and phase is always opposite for the Rossby waves in the stratified fluid (Pedlosky 1987). The downward and westward direction of the wave energy propagation inferred from Fig. 11 is consistent with the ray trajectories obtained by Nagura (2018).

The analysis above examines the climatological wave characteristics ECCO reproduces, and finds fair agreements with results derived from observations. Therefore, we assume that ECCO can be trusted to give a realistic wave field for studying the link between planetary waves and the heaving-associated Indian Ocean overturn. On the other hand, the Fourier analysis on the heaving signals finds that the major limb of the seasonal IMOC (10°–20°S) is primarily associated with the annual component of the Rossby waves. Though Rossby and Kelvin waves well account for the seasonal variation of the Eulerian IMOC, they are just responses to the external forcing instead of the driving mechanism of this seasonality. It is thus fair to ask, what factor drives and determines the strength of the seasonal Eulerian IMOC? Previous studies attributed it to the Ekman transport by zonal winds (Lee and Marotzke 1998) or the localized Ekman pumping (Wang et al. 2014). But we

will show that to a great extent, it is the zonally integrated Ekman pumping anomaly that determines the strength of the seasonal Eulerian overturn in the Indian Ocean.

c. What determines the IMOC seasonality?

The simple 1D linearized wave model has been frequently adopted to study the long baroclinic Rossby waves in the Indian Ocean (e.g., White 1977; Masumoto and Meyers 1998; Birol and Morrow 2001; Wang et al. 2001; Gnanaseelan and Vaid 2010; Watanabe et al. 2016; Menezes and Vianna 2019). It is also used in this study, with a usual form as follows:

$$\frac{\partial h(x, y, t)}{\partial t} + c(y) \frac{\partial h(x, y, t)}{\partial x} = w_{\text{Ekman}}, \quad (7)$$

where $h(x, y, t)$ represents the depth of the isopycnal, and $c(y)$ the phase speed of the baroclinic Rossby wave, which is commonly assumed a function of latitude only. The forcing term on the RHS, w_{Ekman} , refers to the Ekman pumping, which is defined as

$$w_{\text{Ekman}} = \frac{1}{\rho_0} \nabla \times \left(\frac{\boldsymbol{\tau}}{f} \right), \quad (8)$$

FIG. 10. The Fourier fields of (left) amplitudes, (center) percent variance, and (right) phase angle for (top) annual and (bottom) semi-annual components of the depth anomaly of the isopycnal $27.4\sigma_0$. Data: ECCO v4r3.

where ρ_0 is the reference density of water and taken 1000 kg m^{-3} here, τ the wind stress vector, and f the Coriolis parameter, which varies with latitude. The depth of isopycnal can be decomposed into an annual-mean component and a time-varying component (monthly anomaly) as follows:

$$h(x, y, t) = \bar{h}(x, y) + h'(x, y, t), \quad (9)$$

where

$$\bar{h}(x, y) = \int_0^T h(x, y, t) dt, \quad \int_0^T h'(x, y, t) dt = 0, \quad (10)$$

and T is taken as one year. Integrating Eq. (7) over one year, one obtains

$$c(y) \frac{\partial \bar{h}(x, y)}{\partial x} = \bar{w}_{\text{Ekman}}, \quad (11)$$

where the time derivative term vanishes because the isopycnal returns to where it was after one year cycle, and $\bar{w}_{\text{Ekman}}$ denotes the annual-mean Ekman pumping. Substituting Eqs. (9) and (11) into Eq. (7), one can get

$$\frac{\partial h'(x, y, t)}{\partial t} + c(y) \frac{\partial h'(x, y, t)}{\partial x} = w'_{\text{Ekman}}, \quad (12)$$

i.e., the wave equation for monthly anomalies, which was sometimes given directly (e.g., Birol and Morrow 2001; Menezes and Vianna 2019). It is readily seen that the time derivative term on the LHS represents the heaving rate of isopycnal, i.e., w_{iso} .

FIG. 11. The Fourier fields of (left) amplitudes and (right) phase angle for (top) annual and (bottom) semiannual components of the depth anomaly of the isopycnal σ_0 across the latitudinal section of 14.9°S (for annual signal) and 5.4°S (for semiannual signal). The vertical coordinates on the right indicate the approximate mean depth of those densities on the left coordinate. The two horizontal yellow lines indicate $26.5\sigma_0$ and $27.4\sigma_0$, respectively. Data: ECCO v4r3.

FIG. 12. The zonally integrated terms in the linearized Rossby wave model regarding isopycnal $27.4\sigma_0$ in all months. Blue solid: $\sum_x w_{\text{iso}}^{\text{heave}}$, red solid: $\sum_x w'_{\text{Ekman}}$ (not calculated for 5°S – 5°N), yellow solid: $(h'_e - h'_w)/L(y)$, gray solid: $\sum_x w_{\text{Eul}}$ at depth of 958 m (all units are m s^{-1}). Scales in all the subplots of both figures are made the same for a better intercomparisons among months. Data: ECCO v4r3.

If zonally integrate Eq. (12) from the western boundary to the eastern boundary, one obtains

$$\int_{x_w}^{x_e} w_{\text{iso}} dx + c(y)(h'_e - h'_w) = \int_{x_w}^{x_e} w'_{\text{Ekman}} dx, \quad (13)$$

where x_e and x_w denote the longitudinal locations of the western and eastern boundaries, respectively. The terms x_e and x_w are chosen at eastern and western boundaries, respectively. Note that x_e is $\sim 122^\circ\text{E}$ for zonal sections open to ITF. Parameters h'_e and h'_w represent the depth anomalies of the isopycnals at eastern and western boundaries. Because the system [Eq. (13)] is linearized, superposition is feasible for those latitudes along which the integration path crosses land, such as the Indian Peninsula in the north. In the discrete form, Eq. (13) can be written as

$$\sum_x w_{\text{iso}} + c(y) \frac{h'_e - h'_w}{x_e - x_w} = \sum_x w'_{\text{Ekman}}, \quad (14)$$

It is readily seen (Figs. 12 and 13) that at almost all latitudes the term associated with $h'_e - h'_w$ is negligibly small compared with the other two terms. This term slightly deviates from zero only around the northern boundary of the basin, where the effective basin width (thus the length of the path for Rossby wave to propagate) quickly decreases, and coastal Kelvin waves modify the relationship of the linearized wave model, Eq. (12). Here, $c(y)$ is set to unity. Considering the phase speeds of the baroclinic Rossby waves are no greater than unity at all latitudes (e.g., Polito and Liu 2003; Killworth and Blundell 2003a,b), the actual magnitude of this term is even smaller than it appears in Figs. 12 and 13.

Given how small the term $h'_e - h'_w$ is, the zonally integrated heaving rate of the isopycnals is supposed to be approximately equal to the zonally integrated Ekman pumping anomalies as predicted by the linearized Rossby wave model. This is quite true for both isopycnals in all months as indicated in Figs. 12 and 13, especially to the south of 10°S , where there are longer and cleaner lanes for the Rossby waves to run and fewer interruptions by coastal Kelvin waves than in the northern Indian Ocean. While there is a qualitative agreement between $\sum_x w_{\text{iso}}^{\text{heave}}$ (blue solid) and $\sum_x w'_{\text{Ekman}}$ (red solid) in the northern Indian Ocean, quantitatively this agreement is not as good as in the south. This distinction implies the linearized Rossby wave model is not applicable in explaining the heaving of the isopycnals in this region, because the isopycnal movements are also impacted by the active coastal Kelvin waves therein (Vialard et al. 2009; Rao et al. 2010; Suresh et al. 2013).

Also plotted for reference in Figs. 12 and 13 is the zonally integrated Eulerian vertical velocity, $\sum_x w_{\text{Eul}}$. As the close resemblance between w_{Eul} and w_{iso} has been demonstrated in previous section, the agreements between $\sum_x w_{\text{Eul}}$ and $\sum_x w_{\text{iso}}$ in both figures is not surprising. However, one thing worth noting is that the all-year-round downwelling motions relative to the isopycnal (i.e., cross-isopycnal motions) to the south of 20°S in Fig. 13 correspond precisely to the latitudes of the subduction area of the shallow overturning circulation in the Indian Ocean identified in previous researches (Schott et al. 2004; Schott 2005; Schott et al. 2009). This diapycnal downwelling is absent in Fig. 12 because this shallow overturning circulation is found within the upper 500 m only (Miyama et al. 2003; Meng et al. 2020).

FIG. 13. As in Fig. 12, but for the isopycnal $26.5\sigma_0$.

To summarize this section, the Eulerian vertical velocities in the Indian Ocean are largely due to the heaving motion of isopycnals. These heaving motions reflect the propagation of Rossby and Kelvin waves active in the tropics. The largest amplitude of dominant component in the heaving signals, i.e., annual Rossby waves, which leads to the most substantial upwelling/downwelling limb of the seasonal Eulerian IMOC, lies in the latitude 10° – 20° S. Unlike the conclusions made in previous studies, the zonally integrated Ekman pumping anomaly is the major external factor determining the Eulerian IMOC strength.

4. The sloshing MOC and the diapycnal MOC

In section 3, the seasonal variation of the Eulerian IMOC was linked to the motion associated with the isopycnals. Is there a way to separate this isopycnal-heaving mode from the Eulerian IMOC? Projecting MOC streamfunction to the density coordinate is one way to do this job (e.g., Döös and Webb 1994; Lumpkin and Speer 2007). In this paper, as the use of the vertical velocity component was proven to be equivalent to the use of the horizontal velocity component in section 2's MOC streamfunction calculation, it is possible to use the water parcel's vertical velocity component associated with isopycnals, w_{iso} , to derive a new MOC streamfunction depicting the isopycnal-heaving mode of Eulerian overturn in the z coordinate. According to Eqs. (1) and (2), this streamfunction can be defined as follows:

$$\psi_{\text{iso}}(y, z) = - \int_y^{y_N} dy' \int_{x_w}^{x_e} w_{\text{iso}}(x, y', z) dx, \quad (15)$$

where y_N denotes the latitudinal location of the northern boundary of the Indian Ocean. This streamfunction, ψ_{iso} , is

referred to as the “sloshing MOC,” and the Eulerian version is written as ψ_{Eul} hereafter. Reminiscent of the previous introduction, “sloshing” in this paper (as in Schott and McCreary 2001) refers to the back-and-forth horizontal motions due to the movement of the isopycnals. The sloshing MOC is meant to describe the isopycnal-heaving mode in the Eulerian overturning circulation. No diapycnal mixing or water mass transformation is occurring in the sloshing mode. Thus it represents the reversible part of the Eulerian MOC. As a result, the residual part of the streamfunction,

$$\psi_{\text{dia}} \equiv \psi_{\text{Eul}} - \psi_{\text{iso}}, \quad (16)$$

which is defined as the “diapycnal MOC,” represents the diabatic mode of the Eulerian MOC. It refers to water mass transformation and irreversible diapycnal mixing processes.

To estimate ψ_{iso} and ψ_{dia} , the only new thing to do is to interpolate w_{iso} , which is currently defined at the depth of isopycnal, to the standard depth levels of ECCO to align with the Eulerian velocity components. This interpolation is done with $w_{\text{iso}}(\sigma_0)$, $w_{\text{iso}}(\sigma_2)$, and $w_{\text{iso}}(\sigma_4)$, and the median value of them at any specific depth is used to reduce uncertainties. The profiles of $w_{\text{iso}}(\sigma_0)$, $w_{\text{iso}}(\sigma_2)$, $w_{\text{iso}}(\sigma_4)$, and the interpolated w_{iso} are demonstrated in three example sites in Fig. 14. Good convergence among $w_{\text{iso}}(\sigma_0)$, $w_{\text{iso}}(\sigma_2)$, and $w_{\text{iso}}(\sigma_4)$ is reassuring. The heaving rate at the sea surface is taken as zero, i.e., $w_{\text{iso}}(z = 0) = 0$, because the heaving rate of the free surface is negligibly slow (two to three orders of magnitude smaller) compared with the isopycnal surfaces underwater according to the reduced gravity model and also as being illustrated in the right column of Fig. 9. Because the large uncertainties in identifying isopycnals in well-mixed layers may lead to significant errors in w_{iso} , the heaving rates for the upper 100 m are chosen to be all linearly interpolated with $w_{\text{iso}}(z = 0) = 0$ and $w_{\text{iso}}(z = 100 \text{ m})$.

FIG. 14. The interpolation of the heaving rate w_{iso} to the standard depth levels of ECCO. (top left) Three examples of a w_{iso} profile are shown with randomly selected positions that are marked in red dots in the bathymetry map of the Indian Ocean. The blue, red, and yellow lines represent w_{iso} computed with potential densities with respect to different pressures, σ_4 , σ_2 , and σ_0 . The purple circles denote the heaving rates interpolated to the standard depth levels, defined as the median of the three solid lines. The longitude, latitude, and depth of each site are indicated above the three panels. Data: ECCO v4r3.

Higher-resolution density coordinates were tested, but did little to improve results.

With w_{iso} interpolated to the ECCO grid, the sloshing and diapycnal MOC streamfunctions, ψ_{iso} and ψ_{dia} , are calculated and illustrated together with ψ_{Eul} in Figs. 15–17. The striking resemblance between ψ_{Eul} and ψ_{iso} in every month of the year further advocates for the conclusion that the seasonal variation of Eulerian IMOC is primarily a sloshing mode. The deep

ocean becomes much more quiescent in the diapycnal MOC than the Eulerian one. This characteristic can also be seen from the annual mean of the three MOC streamfunctions in Fig. 18, in which the Eulerian MOC is close to those given in the previous results with seasonal or annual mean forcings (Lee and Marotzke 1998). It is believed to result from the canceling of the back-and-forth sloshing motions whose period is one year. On the other hand, a weak but rather persistent clockwise

FIG. 15. The climatological monthly (top) Eulerian IMOC, (middle) sloshing MOC, and (bottom) diapycnal MOC from (left) January to (right) April. Data: ECCO v4r3.

FIG. 16. As in Fig. 15, but from May to August.

diapycnal overturning cell is present in the deep ocean (under 3000 m) in the half-year of boreal spring and summer (bottom rows of Figs. 15–17). This feature is robust in this method as it exists even in calculations using higher-resolution isopycnal coordinates (not shown). However, it has to be determined whether it results from a physical process or just the numerical error arising from the method and ECCO's limited vertical resolution. In any case, this deep diapycnal cell is flagged with a question mark for future investigation and will not be discussed in the rest of this paper. Omitting the uncertain deep cell, the

most remarkable feature in the diapycnal MOC is the persistent shallow overturning cell which is concentrated above 500 m (e.g., Schott and McCreary 2001; Miyama et al. 2003; Lee 2004; Schott et al. 2004; Schott 2005; Schoenefeldt and Schott 2006; Meng et al. 2020). Unlike the seasonally reversing Eulerian cells, this shallow overturning cell has critical implication for the annual-mean MHT in the Indian Ocean. More details on it and its impact in MHT will be discussed at a later stage.

To summarize, a comparison of the streamfunctions of the Eulerian and sloshing MOCs implies that the sloshing is the

FIG. 17. As in Fig. 15, but from September to December.

FIG. 18. As in Fig. 15, but for the annual mean.

dominant mode in the Eulerian meridional overturnings of seasonal scale. This conclusion can provide new insights into understanding some dynamical questions of the Indian Ocean, such as the upwelling/downwelling hot spots of the shallow overturning circulation, the discrepancies in magnitude of the deep inflow into the basin, the missing mixing problem in maintaining the deep overturning circulation, and the mechanism for MHT in the Indian Ocean. These questions are to be discussed separately in the following subsections.

a. The shallow overturning circulation

For a clearer picture of the shallow overturning circulation, another version of the diapycnal MOC streamfunction, ψ_{dia} , is presented in Fig. 19, which is the same as in Figs. 15–17, but zoomed in to upper 500 m. Compared with previous pictures, which only showed the part of the overturn north of ITF (e.g., Miyama et al. 2003; Schott 2005; Schott et al. 2009; Meng et al. 2020), the solutions here give a more complete depiction of the

circulation. Note, however, that ψ_{dia} above 100 m is less reliable because it is obtained entirely by linear interpolation with $w_{iso}(z = 0)$ and $w_{iso}(z = 100 \text{ m})$ due to the ill-defined isopycnals in the mixing layers.

Previous studies have identified several cells in the shallow overturning circulation system, such as the cross-equatorial cell (CEC), the subtropical cell (STC), and the “equatorial roll” (Wacongne and Pacanowski 1996; Schott et al. 2002; Horii et al. 2013), which are all present in Fig. 19. To be specific, the STC refers to the strongest cell in the Southern Indian Ocean, which is anticlockwise all year round. The CEC refers to the shallow cell to the north of the equator. Unlike the STC, the CEC reverses during a year, with notable clockwise overturn from December to March, and anticlockwise the rest of the year. Such characteristics are consistent with the Ekman transports’ direction subject to the monthly surface wind stress off the equator (e.g., Schott and McCreary 2001; Miyama et al. 2003; Schott et al. 2009). In qualitative consistency with the

FIG. 19. The diapycnal MOC streamfunctions in the upper 500 m, which are the zoomed-in version of the bottom rows of Figs. 15–17. The contour lines of -2 Sv are marked in white. Data: ECCO v4r3.

FIG. 20. (left) The Eulerian vertical velocity w_{Eul} , (center) the heaving rate of isopycnals w_{iso} , and (right) their difference for (top) January, (middle) July, and (bottom) annual mean at a standard depth level of ECCO, ~ 110 m. The dark red (dark blue) boxes denote the upwelling (subduction) areas of the shallow overturning circulations identified by Schott et al. (2002).

previous study (e.g., Schott and McCreary 2001), the equatorial roll, situated at the equator above 100 m, is most vigorous during boreal summer and reverses direction during the winter. Note, although both CEC and equatorial roll are manifested in the diapycnal MOC solution, it does not necessarily mean both cells cause diapycnal fluxes because it still depends on the mixed layer depth (Schott and McCreary 2001). The two cells are included in the diapycnal MOC solution only because the depth range (100 m) in which they are accommodated is the blind spot of the sloshing MOC defined in this study.

Another uncertain question of the shallow overturn is the hot spots where the water parcels upwell/subduct. A schematic representation by Schott et al. (2002) has pointed out several subduction/upwelling areas regarding the shallow overturning circulation in winter and summer and modified afterward to have those areas from both seasons merged into one image (e.g., Schott et al. 2004; Schott 2005; Schoenfeldt and Schott 2006; Schott et al. 2009). In this study, new upwelling branches are suggested by virtue of the obtained cross-isopycnal vertical velocity. Upwelling occurs mostly due to the combined effect of thermocline shoaling (adiabatic process) and the entrainment–detrainment processes across the isopycnals (diabatic process). As schematized in Fig. 4, the thermocline shoaling process can be represented by the vertical velocity associated with isopycnals w_{iso} and the diapycnal

entrainment processes can be measured by the diabatic vertical velocity w_{dia} .

Figure 20 presented the w_{Eul} , w_{iso} , and their difference, i.e., the diabatic vertical velocity w_{dia} , at the level of ~ 110 m, a typical depth across the STC center (some other nearby levels are also examined, and the patterns remain qualitatively the same). The subduction (dark blue boxes) and upwelling (dark red boxes) areas for January and July illustrated in Schott et al. (2002) are superimposed for comparison. For January, the upwelling box does not match an upwelling. While in the subduction box, there is downwelling in the Eulerian velocity field, and it is almost all accounted for by the vertical movements of the isopycnals, which gives rise to little cross-isopycnal downwelling (Fig. 20, upper right). For July, the coastal upwelling indicated in the dark red box, again, is mostly accounted for by isopycnal heaving instead of diapycnal processes. Though with a much-reduced strength compared with the Eulerian field, the subduction box does match a large area of downward diabatic motions in the southeastern basin (Fig. 20, middle row), which is known as the “subtropical subduction zone” (Schott et al. 2002; Miyama et al. 2003; Schott et al. 2004; Schoenfeldt and Schott 2006). Moreover, this subduction area is not only active in austral winter, but all year round, as evidenced by both Figs. 13 and 19.

Due to the uncertainty arising from the cancellation of w_{Eul} and w_{iso} in this method, these interior upwelling and

downwelling regions remain uncertain. However, the areas of strong coastal upwellings off Kenya and northeast of Madagascar may serve as the upwelling branch of the STC from 15°S to the equator aside from the area proposed in Schott et al. (2002) (the dark red box in the upper and bottom rows of Fig. 20: 50°–75°E, 12°–5°S). That area, in fact, is known as the Seychelles–Chagos thermocline ridge (SCTR; McCreary et al. 1993), or the southwestern tropical Indian Ocean thermocline ridge (STR; Hermes and Reason 2008; Halkides and Lee 2011), or the western Indian Ocean upwelling (WIO; Zhang and Han 2020). It is characterized by a shallow depth of 20°C isotherm. As is thoroughly reviewed in a recent study (Zhang and Han 2020), this upwelling zone is primarily caused by Rossby waves driven by remote winds with a combination of local Ekman pumping. As being examined with another ocean reanalysis data by Zhang and Han (2020), it is the mean state rather than the heaving rate of the shoaling isopycnals that is essential to the upwelling in this area. As a result, there is not much information on upwelling from w_{iso} , but it is detectable from the diapycnal vertical velocity component, w_{dia} (bottom row, Fig. 20). This is consistent with the outcome by Halkides and Lee (2011) that “the dominant ocean process (in the SCTR) is wind-driven vertical mixing,” while “vertical advection as a direct response to thermocline shoaling has little effect on the mixed layer temperature.” As also revealed by the annual mean diabatic vertical velocity, the diabatic upwelling off Kenya and north of Madagascar is equally prominent as the SCTR region (bottom right of Fig. 20). As found out by potential vorticity (PV) analysis with the Argo floats data, there is no interior way in the Indian Ocean that connects the Southern Hemisphere interior to the equatorial region. Thus the water parcels must transit through the western boundary to reach the equator (Nagura and McPhaden 2018). A GCM Lagrangian tracer study also found that water from the Southern Hemisphere crosses the equator via the western boundary (Miyama et al. 2003). It is very likely that the subducted water in the subtropical subduction zone is advected westward to the western boundary and upwelled along the African coast. However, these upwelling branch findings are no more than qualitative conjectures due to the uncertainties associated with the inferred diabatic vertical velocity.

b. The deep overturning circulation and the missing mixing

As introduced in section 1, the discrepancies are large on estimating the deep IMOC strength. Analysis by Robbins and Toole (1997), Schott and McCreary (2001), and Huussen et al. (2012) suggest an average value of 10 Sv with a range of 3–17 Sv based on the hydrographic sections occupied in June–July of 1965, November–December of 1987, and March–April of 2002 [see Huussen et al. (2012) for a list].

Though an emphasis has not been put on the seasonal variation of the deep IMOC in those estimates based on the hydrographic data, studies with model data do find evident seasonal and interannual variability of it (e.g., Wang et al. 2012, 2014). Unlike the observation-based results, models have found the deep meridional transports across 32°S are not always inflows, but sometimes manifest as outflows (Fig. 4 of Wang et al. 2012). With solutions of the streamfunctions

derived in previous sections, the net volume transports across 32°S of the Indian Ocean under 2000 or 3000 m are presented by month together with the estimates based on hydrographic sections plotted as error bars in Fig. 21. For one thing, transports from different solutions of the Eulerian and sloshing MOCs resemble one another on the annual cycle patterns. This implies that the seasonal variation of the deep circulation is largely due to the sloshing motions. For another, the estimates from hydrographic sections in different seasons, though generally larger, also exhibit an annual cycle in phase with those estimates from ECCO. Therefore, it is suggested that the discrepancies among the observation-based estimates on the bottom transports across 32°S of the Indian Ocean result from the seasonality of deep flows associated with the sloshing motions.

The second question associated with the deep overturn in the Indian Ocean is the overestimated diffusivity in the abyssal ocean (Schott and McCreary 2001). As introduced in section 1, the current estimates based on advective–diffusive balance are an order of magnitude larger than the measured diffusivities. Again, from the perspective of sloshing motions, the missing mixing problem to sustain the deep overturning circulation may become understandable. As argued above, the deep overturn is dominated by the sloshing mode of the water motions. This is to say, though there is significant inflow through the deep levels across the zonal section at, say 32°S, and most of the entering water parcels do ascend, those parcels do not actually cross the isopycnals when ascending, and thus require nothing from the diapycnal diffusivity. The real diapycnal volume flux can be inferred from the diapycnal MOC, as shown in Fig. 18 or the bottom rows of Figs. 15–17. Putting aside the uncertain deep clockwise cell as stated in the previous section, the diapycnal upwelling in the Indian Ocean becomes significantly weaker than 10 Sv and concentrated between 1000 and 3000 m for all months. With the annual mean diapycnal MOC streamfunction (Fig. 18), the averaged diapycnal diffusivities at 1000 and 2000 m can be estimated with the 1D advective–diffusive balance as follows for different subdomains of the basin,

$$\bar{\kappa}(y, z) = \frac{\rho_z}{\rho_{zz}} \frac{V_{\text{up}}}{A(y, z)}, \quad (17)$$

where $A(y, z)$ represents the total area of the surface upon which the averaged diffusivity is estimated, for example, $A(20^\circ\text{S}, 1000\text{ m})$ refers to the total area of the surface at 1000 m and north of 20°S in the Indian Ocean; V_{up} represents the total diapycnal upwelling volume flux across A ; ρ_z and ρ_{zz} denote the first- and second-order vertical derivatives of the density. A typical value of ρ_z/ρ_{zz} for the deep Indian Ocean could be taken as 1000 m (Schott and McCreary 2001); $\bar{\kappa}$ represents the diapycnal diffusivity average over the area A . With the values of V_{up} being read from the diapycnal MOC streamfunction in Fig. 18, $\bar{\kappa}$ could be calculated with results listed in Table 1. The averaged diffusivities are all smaller than $1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$. If considering the vertically averaged diffusivity, each estimate in Table 1 might be further reduced by a factor of 2–3 [the diapycnal upwelling is not present in the whole column, but only

FIG. 21. The northward deep transports obtained with various solutions of the MOC streamfunctions (solid lines) and the estimates from hydrography data collected by three transbasin expeditions in different seasons across 32°S of the Indian Ocean. The meanings of those streamfunctions can be found in previous sections, for example, $\psi_{Eul}^V(32^\circ\text{S}, 3000\text{m})$ refers to the value of the Eulerian IMOC obtained by integrating V bottom-up at 32°S and 3000m , which equals to the net meridional volume flux from the bottom to 3000m across 32°S . Data for streamfunctions: ECCO v4r3. Bars denote the hydrographic section results. The horizontal bars mark the months when the expedition took place, and the vertical bars mark the range of the estimates. These estimates include $10 \pm 2\text{ Sv}$ (McDonagh et al. 2008) for the expedition in March–April 2002, 3.6 Sv (Fu 1986) for the one in June–July 1965, $12 \pm 3\text{ Sv}$ (Robbins and Toole 1997), $17 \pm 5\text{ Sv}$ (Macdonald 1998), $11 \pm 4\text{ Sv}$ (Ganachaud et al. 2000), and $10 \pm 3\text{ Sv}$ (Bryden and Beal 2001) for the one in November–December 1993.

part of it, say the column of $\sim 2000\text{ m}$ ($1000\text{--}3000\text{ m}$) within the total depth of $\sim 5000\text{ m}$ as indicated in Fig. 18], which makes a vertically averaged deep diffusivity closer to $1 \times 10^{-5}\text{ m s}^{-1}$. In any case, these estimates become much more consistent with the observations of mixing than previous ones.

The deficient energy for dissipation found by Huussen et al. (2012) can also be explained in a way that not that much energy is actually required for the diapycnal mixing in the deep ocean. Thus, the view of sloshing provides a possible explanation to reconcile the missing mixing problem in the deep Indian Ocean.

c. Mechanism of meridional heat transport in the Indian Ocean

The MHT mechanism in the Indian Ocean can also be reinvestigated from the sloshing perspective. As introduced in section 1, before studying MHT due to MOCs, ITF's non-volume-balanced heat transport needs to be separated. However, the commonly used BBH decomposition underestimates the temperature transport by ITF by treating the ITF advected

temperature as a section mean, while in reality the ITF water is concentrated in the warmer upper levels. A new method is invented for this question. Due to the complexity of its pathways in the Indian Ocean, the ITF water is assumed to flow southward as soon as its entering the basin. Model researches indicate that some ITF water spreads northward before turning southward (e.g., Song et al. 2004; Valsala and Ikeda 2007), but it does not transform much during its northward trip, thus has little impact on MHT of the concern here. Rather than analyzing the detailed southward pathways, we could estimate the temperature transport by ITF with five different assumptions defining the temperature it advects:

- 1) ITF advected temperature is defined at the open boundary where it enters from the east.
- 2) ITF advected temperature is defined as the zonal mean at corresponding latitude and depth.
- 3) Same as case 2, but the zonal mean is defined west of 60°E .
- 4) Same as case 2, but only the upper 500 m where the bulk of the ITF is found is included in the mean (Liu et al. 2005; Wang et al. 2014).

TABLE 1. Estimates of the area-averaged diapycnal diffusivity derived by the vertical advective–diffusive balance. Several surfaces upon which the averaged diffusivities are estimated are used, with depths at 1000 and 2000 m and the southern boundary at 10° , 20° , and 30°S , respectively. The diapycnal upwelling volume flux values V_{up} are read from the annual-mean diapycnal MOC in Fig. 18.

Depth of the area	1 km	1 km	1 km	2 km	2 km	2 km
Southern boundary of the area	10°S	20°S	30°S	10°S	20°S	30°S
Diapycnal upwelling V_{up} (Sv)	0.33	1.0	2.2	0.83	1.4	2.7
Diffusivity $\bar{\kappa}$ ($10^{-5}\text{ m}^2\text{ s}^{-1}$)	1.9	3.9	6.4	5.1	5.9	8.3

FIG. 22. Components of the BBH decomposition on MHT in the Indian Ocean with the annual mean data of ECCO (blue: the barotropic component depicting the non-volume-balanced MHT; red solid with circles: the overturning component; yellow: the gyre component; black: total MHT of direct estimate). The red line with shade denotes the estimates of the MHT associated with ITF and its uncertainties. The error bars denote the “plain MHT” provided in Forget and Ferreira (2019), also with ECCO.

5) Same as case 4, but the zonal mean is defined west of 60°E.

With the annual mean of ECCO, MHT associated with ITF is shown in Fig. 22, with the shade denoting the range of the estimates with different assumptions. Also plotted in Fig. 22 is the direct estimate of the MHT ($\iint \rho c_p v T dx dz$) and its components obtained by the BBH decomposition as briefly introduced in section 1 (Bryden and Imawaki, 2001; Macdonald and Baringer 2013). Not surprisingly, the barotropic component associated with ITF (blue solid with circles) is significantly smaller than the estimates by the method above (red solid with shade). Regarding the other two components, the total MHT (black solid) is dominated by the overturning component (red solid with circles), while the gyre component (yellow solid with circles) is only of minor importance in almost all latitudes. This decomposition result is consistent with that of Lee and Marotzke (1998) using the GFDL model output, only that the magnitude of the total MHT is notably larger for the ECCO result. Since Forget and Ferreira (2019) also calculated MHT of the Indian Ocean with ECCO and their results (shown as error bars in Fig. 22) are consistent with the result here, this difference in magnitude is attributed to the differences in models. On the other hand, this difference will not impact the conclusion on the MHT mechanism afterward. The product of density and specific heat, ρc_p , is nearly constant in the ocean ($\pm 1\%$) with a mean value of $4.1 \times 10^6 \text{ J m}^{-3} \text{ }^\circ\text{C}^{-1}$ (Johns et al. 2011; Msadek et al. 2013), which is adopted in the MHT calculation here.

Since the overturning MHT (red solid with circles in Fig. 22) is the dominant component in the total, it is thus worthwhile to further separate it into components associated with the sloshing overturn and diapycnal overturn, respectively. Though the 3D flow fields are not solved for either sloshing or diapycnal motions, the algorithm of the baroclinic component in the BBH decomposition only requires the zonally integrated quantities, which are available as MOC streamfunctions as indicated in

Figs. 15–17. The zero-net-transport condition necessary for a volume-balanced heat transport (Montgomery 1974) is also checked for ψ_{Eul} , ψ_{iso} , and ψ_{dia} . For ψ_{iso} , the net transport is zero by definition. While for ψ_{Eul} and thus ψ_{dia} , the net transport is of $O(0.1)$ Sv, nonzero but orders of magnitude smaller than the overturning strengths. The decomposition results for each month are demonstrated in Fig. 23.

It is intriguing to find that the direct estimates (green solid, Fig. 23) become close to the MHT components associated with the Eulerian MOC (black thick, Fig. 23) after removing the ITF component estimated by abovementioned method (the red solid with shade, Fig. 23). It is because ITF water flows through the basin almost horizontally that the Eulerian MOC, ψ_{Eul}^W , is little affected by ITF. That is to say, ψ_{Eul}^W , which is obtained by integrating the vertical velocity component, is virtually unaware of the volume input of ITF.

After separating the MHT by Eulerian MOC into components of sloshing and diapycnal modes, respectively, it becomes immediately apparent that the diapycnal MOC (black thin, Fig. 23) almost always transports heat southward while it is the sloshing mode (dashed, Fig. 23) that produces the seasonal variations of (roughly) the total MHT. For example, when the sloshing MHT gets stronger than the diapycnal one in the boreal winter, the total MHT turns northward all across the equator, consistent with the strong clockwise overturning cells at that time. This result provides a quantitative evidence to the statement made by Schott and McCreary (2001) almost 20 years ago, “this seasonal heat transport anomaly is dominantly 3a ‘sloshing’ motion that increases heat storage in the northern part of the basin during the winter monsoon but decreases it during the summer monsoon.”

MHT by the diapycnal MOC of the upper 500 m (open circles, Fig. 23) only shows a slight difference with full depth (thin solid, Fig. 23). This implies the diapycnal overturning component of MHT in the Indian Ocean is primarily caused by the

FIG. 23. The climatological MHT components in the Indian Ocean: total MHT of direct estimate (green solid), total MHT excluding the ITF component (red solid with shade), MHT by the Eulerian MOC, ψ^w (thick solid), MHT by the sloshing MOC (dashed), and MHT by the diapycnal MOC of full depth (thin solid) and the upper 500 m only (open circles). Data: ECCO v4r3.

shallow overturning circulation discussed in the previous subsection.

The sensitivity of above MHT decomposition to the artificially chosen depth above which w_{iso} is obtained all by interpolation (see the beginning of this section) is also tested with other options (50 and 150 m). The results are not notably altered.

5. Summary and discussion

With an analysis on the climatological fields of an ocean state estimate product, ECCO v4r3, this study finds the vertical movements of the water parcels in the Indian Ocean are

primarily caused by the isopycnal heaving, particularly away from the western coast. The notable seasonal variation of Eulerian IMOC is thus attributed to the heaving pattern of the isopycnals. Aided by the linear Rossby wave model, the zonally integrated Ekman pumping anomaly is found responsible for the strengths of the Eulerian IMOC. This is different from previous conclusion that the Ekman transport plays a major role.

The term introduced by Schott and McCreary (2001), “sloshing,” is used to describe the seasonal-reversing horizontal motions induced by the undulating isopycnals. The concept of sloshing MOC is proposed relative to the conventional Eulerian MOC. The striking resemblance of both MOCs

FIG. 24. The zonal mean depth anomaly of the isopycnal $27.4\sigma_0$ for 12 months in the Indian Ocean. The ones in April (blue) and October (red) are highlighted. Positive (negative) anomaly here means the isopycnal is perturbed shallower (deeper) than its annual-mean position. Data: ECCO v4r3.

FIG. 25. A schematic representation of the IMOC seasonality due to sloshing. The horizontal blue dashed denotes the annual mean position of a certain isopycnal and the thick blue curve its seasonal undulations. Yellow arrows indicate the poleward (equatorward) horizontal sloshing flows induced by stretching (squeezing) of the water columns as dictated by PV conservation. The blue arrows denote the heaving motions of the isopycnals associated with the planetary waves. The red arrows represent the heaving motion on the equator. It is induced by the convergence/divergence of the horizontal sloshing motions, instead of by Ekman pumping since $f = 0$ on the equator. This sloshing convergence/divergence is due to the asymmetric strengths of heaving motions in the two hemispheres. The most quiescent overturn occurs in boreal spring and autumn with the maximum displacements of the isopycnals. The overturning is so weak in October that no organized structure fitted to this schematic can be identified. Black horizontal arrows indicate the deep inflow with their widths representing seasonally varying strengths of the flow (Fig. 21). The upward curly arrows denote the diapycnal upwelling relatively strong in April and October (see bottom rows of Figs. 15–17). Note, this explanation is only applicable to the adiabaticity-dominant motions, so that the isopycnals deviate not far from the material surfaces at all times.

implies that the Eulerian meridional overturn in the Indian Ocean is mostly adiabatic motion, particularly in the deep and abyssal oceans. This sloshing mode is most clearly revealed by the zonally mean isopycnal depth anomaly in Fig. 24. As is known for a sinusoidal oscillation, the displacement has a phase lag of $\pi/2$ with respect to the velocity. This is why the maximum displacements of the isopycnal shown in Fig. 24 are obtained one season (three months) after January and July when the Eulerian IMOC streamfunctions reach their maximum (Fig. 1 or Fig. 2). Based on these results, the IMOC seasonality dynamics due to sloshing is schematized in Fig. 25 for typical months of four seasons, with detailed explanations in the caption.

The difference between Eulerian and sloshing MOCs defines the “diapycnal MOC” (bottom rows of Figs. 15–17). It is intriguing to find that the most remarkable structures in the diapycnal MOC are the shallow overturning cells and even these remain uncertain due to the isopycnal method’s inability to resolve the mixing layer. A solution extending south to 30°S is obtained for the shallow overturning cells, depicting a more complete image than earlier

studies (Fig. 19). The upwelling/subduction zones of the shallow overturn are qualitatively investigated with the diabatic vertical velocity. New perspectives inferred from the analysis include: The downwelling in the northwestern Arabian Sea and upwelling off Somalia are primarily due to isopycnal heaving; while the upwelling in the Seychelles–Chagos thermocline ridge area is more connected with the mean state rather than the heaving motion of isopycnals. Moreover, the upwelling zones off Kenya and northeast Madagascar are suggested to be components of the upwelling branch of those shallow overturning cells (Figs. 19 and 20).

The perspective of sloshing is applied to provide new interpretations to answer questions concerning the deep overturning circulation in the Indian Ocean. It is concluded that discrepancy in estimates of the magnitude of bottom water transport into the Indian Ocean across 32°S has resulted from seasonal sloshing. While the missing deep mixing is believed to be a result of sloshed water parcels undergoing upward motions without crossing isopycnals. New estimates of diapycnal diffusivity based on the diapycnal IMOC provide improved consistency with in situ observations.

The volume-balanced part of MHT in the Indian Ocean is partitioned into sloshing and diapycnal components. It makes clear how this MHT is controlled by the sloshing and diapycnal modes: Its seasonal variation is dominated by the sloshing mode. While the diapycnal MHT component, which is primarily caused by the shallow overturning circulation, is stable throughout the year, transporting heat southward at virtually all latitudes. Thus, for annual or longer time scale, the diapycnal MHT component becomes the most dominant factor.

Limitations in this study include the following aspects. All the conclusions drawn in this study are inferred from ECCO, thus are product dependent. The density coordinate adopted does not provide adequate resolution in the upper ocean where strong mixing occurs nor in the abyssal ocean where stratification is particularly weak, which undermines the results concerning the shallow overturning circulation cells, and to a lesser extent, the deep overturning circulation. The limited spatial resolution in both horizontal and vertical directions might also be a source of uncertainties to the results.

Some questions remain unanswered, such as whether the clockwise deep cell in the diabatic MOC is a real phenomenon or an error. This paper only discusses the seasonal variation of the sloshing motion. How does its longer (interannual or decadal) variability influence the circulation of the Indian Ocean? What are the situations in other basins? In fact, among many studies on the seasonal variability of the Atlantic MOC, a schematic figure did express the concept of sloshing (Fig. 7 of Yang 2015). Given the interpretation of the missing mixing problem in this paper, would the sloshing idea provide any clue to the dichotomy of diffusivities in the global ocean, i.e., the discrepancy between the “pelagic diffusivity” of $10^{-5} \text{ m}^2 \text{ s}^{-1}$ and the inferred diffusivity of $10^{-4} \text{ m}^2 \text{ s}^{-1}$ required to balance the global bottom water formation rate (Munk and Wunsch 1998)?

Other fundamental questions arise if the sloshing mode is indeed dominant globally on basin scales. For example, could the Indian Ocean still be known as “an important deep upwelling limb of the global MOC” (e.g., Huussen et al. 2012)? After all, the upwelling under the Eulerian framework is chiefly an adiabatic process of sloshing. Those deep-water parcels have not been actually brought up across isopycnals and transformed into other water masses. When studying the implications of the meridional overturn on climate scales, the diapycnal MOC may serve as a better candidate than the Eulerian one.

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Data availability statement. The ECCO reanalysis dataset adopted in this study, version 4 release 3, is available at the website ecco.jpl.nasa.gov for open access.

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