Decadal Variability of the Upper-Ocean Salinity in the Southeast Indian Ocean: Role of Local Ocean–Atmosphere Dynamics

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ABSTRACT: Ocean salinity plays a crucial role in the upper-ocean stratification and local marine ecosystem. This study reveals that ocean salinity presents notable decadal variability in upper 200 m over the southeast Indian Ocean (SEIO). Previous studies linked this salinity variability with precipitation anomalies over the Indo-Pacific region modulated by the tropical Pacific decadal variability. Here we conduct a quantitative salinity budget analysis and show that, in contrast, oceanic advection, especially the anomalous meridional advection, plays a dominant role in modulating the SEIO salinity on the decadal time scale. The anomalous meridional advection is mainly associated with a zonal dipole pattern of sea level anomaly (SLA) in the south Indian Ocean (SIO). Specifically, positive and negative SLAs in the east and west of the SIO correspond to anomalous southward oceanic current, which transports much fresher seawater from the warm pool into the SEIO and thereby decreases the local upper-ocean salinity, and vice versa. Further investigation reveals that the local anomalous wind stress curl associated with tropical Pacific forcing is responsible for generating the sea level dipole pattern via oceanic Rossby wave adjustment on decadal time scale. This study highlights that the local ocean–atmosphere dynamical adjustment is critical for the decadal salinity variability in the SEIO.

KEYWORDS: Indian Ocean; Atmosphere-ocean interaction; Ocean dynamics; Salinity; Decadal variability

1. Introduction

The upper ocean absorbs a large portion of solar radiation and presents marked dynamic and thermodynamic variations compared with the deep ocean. Upper-ocean salinity is an important footprint of the hydrological cycle and climate variability (Roemmich et al. 1994; Durack and Wijffels 2010; Cheng et al. 2020). For example, salinity changes the upper-ocean stratification (Sprintall and Tomczak 1992; de Boyer Montégut et al. 2007; Zhang et al. 2018) and further influences ocean circulation (Fedorov et al. 2004; Menezes et al. 2014; Nagura and Kouketsu 2018). Thus, it is essential to study the variation of upper-ocean salinity to understand the ocean environment.

The upper-ocean salinity is mainly controlled by the freshwater exchange at the air–sea interface and ocean dynamics (Yu 2011). Previous research demonstrated that sea surface salinity is mainly controlled by the surface freshwater flux, while salinity below the air–sea interface is also affected by horizontal advection, mixing, entrainment, and nonlinear ocean dynamics (Nagura and Kouketsu 2018; Qu et al. 2019). Particularly, the advection of freshwater is more important for the tropical oceanic salinity than the mid- and high-latitude regions (e.g., Ponte and Vinogradova 2016). Previous studies pointed out that upper-ocean salinity in the Indo-Pacific Ocean shows significant decadal variability (e.g., Zhang et al. 2013; Du et al. 2015; Nie et al. 2020), overwhelming the long-term anthropogenic trend (Sun et al. 2021). Based on Argo observations, the upper-ocean salinity was found to increase in the western tropical Pacific and decrease in the southeast Indian Ocean (SEIO; 12°–31°S, 90°–110°E) during 2004–13 (Du et al. 2015). The freshening trend of salinity can be found from the ocean surface to the subsurface area in the SEIO (Hu et al. 2019; Nie et al. 2020).

Different hypotheses were provided to explain the strong decadal salinity variability in the SEIO. Previous studies suggested freshwater flux change, mainly due to the variation in precipitation, should be responsible for the marked decadal salinity variation in the SEIO (e.g., Zhang et al. 2013; Du et al. 2015; Zhang et al. 2016; Hu and Sprintall 2017). The freshwater flux change is remotely induced by the tropical Pacific climate variability. On interannual time scales, El Niño–Southern Oscillation (ENSO) changes the mixed layer salinity off the coasts of Java/Lesser Sunda and northwest Australia via modulating the local precipitation (Zhang et al. 2016). On decadal time

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scale, a contrasting salinity pattern is characterized by freshening in the SEIO and salinification in the western tropical Pacific Ocean from the mid-1990s (Du et al. 2015). This salinity pattern is attributed to a strengthening of the Indo-Pacific Walker circulation with increased precipitation around the Maritime Continent (MC) and decreased precipitation over the western-central tropical Pacific. The Pacific climate variability can also influence the SEIO salinity through ocean dynamics. For example, Zhang et al. (2013) pointed out that the mean meridional currents transported the anomalous salinity from the equatorial Indian Ocean into the SEIO thereby changing the local salinity. Also, Hu et al. (2019) indicated that the Indonesian Throughflow (ITF; Gordon 1986) transported the anomalous salinity from the western Pacific into the SEIO. Furthermore, the interdecadal Pacific oscillation (IPO; Power et al. 1999) induced oceanic Rossby waves, which propagate westward and modulate the thermocline depth off the coast of northwest Australia (Feng et al. 2004; Wijffels and Meyers 2004), can change the SEIO salinity through geostrophic current adjustment (Zhang et al. 2018a).

Although previous studies have shown that precipitation and ocean dynamics are both responsible for salinity variability in the SEIO, most of them attributed the salinity variability to the remote forcing from the tropical Pacific and paid less attention to the role of the local air–sea interactions in the salinity change. We use three Argo-based gridded products—Roemmich and Gilson mapped Argo (RG Argo), the Grid Point Value of the Monthly Objective Analysis (MOAA GPV), and the global ocean Barnes objective analysis Argo gridded dataset (BOA)—to show that the linear trend of salinity in the upper ocean from 2005 to 2013 (Fig. 1a). There is a significant freshening trend (−0.08 psu yr$^{-1}$) in the SEIO. Here we define the upper ocean as the upper 200 m of the ocean, since the salinity anomaly mainly occurs above this depth (Figs. 1b,c). This salinity change not only appears along the coast of northwest Australia, but also extends westward to 90°E. The maximum salinity decrease is located around 20°S, whereas the rainfall variation concentrates over the MC, indicating that the SEIO salinity change is not due to anomalous freshwater flux. Additionally, the salinity trend may not be explained by the ITF transport due to the location of salinity change being south of the ITF outflow region (Gordon 1986). Thus, the trend of upper-ocean salinity in the SEIO may not only be directly influenced by the remote forcing from the tropical Pacific through the atmospheric and oceanic pathways, but may also be related to the local dynamical processes in the tropical Indian Ocean.

The present study investigates the role of local ocean–atmosphere interactions on decadal salinity variability in the SEIO based on observations and assimilation products. We find that the local wind-driven anomalous geostrophic current controls the SEIO salinity variation by transporting the
freshwater from the warm pool region. Further investigation suggests that the geostrophic transport induced by the local oceanic Rossby wave adjustment is mainly affected by the tropical Pacific climate variability through interbasin teleconnection.

The paper is organized as follows. Section 2 introduces data and methods. The decadal variability of upper-ocean salinity in the SEIO and salinity budget analysis during the salinification and freshening periods is shown in section 3. Section 4 analyzes the sea level and wind anomalies related to meridional salinity transport. Section 5 investigates the linkage between the SEIO salinity and large-scale interbasin climate variability from the tropical Pacific. Finally, a discussion and summary are presented in section 6.

2. Data and methods

a. Observational datasets

Different Argo-based gridded products are used to reveal three-dimensional salinity distribution in the Indian Ocean: RG Argo from the Scripps Institution of Oceanography (Roemmich and Gilson 2009), MOAA GPV from the Japan Agency for Marine-Earth Science and Technology (Hosoda et al. 2008), and BOA (Li et al. 2017). These monthly datasets have a horizontal resolution of 1° × 1° and are available from surface to 2000 dbar. The RG Argo and BOA only use Argo profiles to map the ocean state, while MOAA merges Argo profiles, conductivity–temperature–depth profilers, and Triangle Trans Ocean Buoy Network data. The BOA uses slightly different objective interpolation techniques based on Barnes (1964), while the RG Argo and the MOAA use the implemented optimal interpolation method.

In addition, we use daily multisatellite merged sea level anomalies (SLA) from the French Archiving, Validation, and Interpolation of Satellite Oceanographic Data (AVISO) project to analyze the effect of ocean dynamic processes on salinity variation. The AVISO sea level data are available from October 1992, with a horizontal resolution of 0.25° × 0.25°. To analyze the freshwater exchange and atmospheric forcing, we analyze the evaporation, precipitation, and surface wind from the European Centre for Medium-Range Weather Forecasts interim reanalysis (ERA-Interim) product, which has a horizontal resolution of 1° × 1° and spans from 1980 to the present.

b. ECCO data

The Estimating the Circulation and Climate of the Ocean version 4.3 (ECCO v4.3) is a global ocean model product from 1992 to 2015, simulated by the Massachusetts Institute of Technology General Circulation Model (MITgcm; Marshall et al. 1997a,b). The horizontal resolution of this model is 1° × 1°. In the meridional direction, the resolution gradually increases to 1/3° within 10° of the equator. The model has 50 vertical layers, with an interval of 10 m from the surface to 150 m. ECCO v4.3 is forced by the ERA-Interim atmospheric dataset (Dee et al. 2011) and assimilates multiple sources of oceanic observations (Wunsch et al. 2009; Forget et al. 2015) including Argo data and various satellite measurements with the Soil Moisture and Ocean Salinity (SMOS) (Reul et al. 2014), National Aeronautics and Space Administration Aquarius/Satellite de Aplicaciones Científicas (SAC-D) (Lagerloef 2012), and the Soil Moisture Active Passive (SMAP) missions. Another remarkable advantage of the ECCO data is that its variables exactly obey known conservation rules and are free of artificial internal heat and freshwater sources/sinks, allowing closed salinity tracer budget diagnostics. This feature enables the term-by-term salinity budget diagnosis (Ponte and Vinogradova 2016; Vinogradova and Ponte 2017; Zhang et al. 2018a; Hu et al. 2019). The freshwater fluxes (evaporation and precipitation) were adjusted to satisfy the conservation laws in ECCO v4.3 (Forget et al. 2015; Liu et al. 2019), resulting in these variables deviating from the observations. Therefore, in this study we use freshwater flux from the ERA-Interim, which is ECCO’s forcing data.

3. Salinity budget analysis

A salinity budget is evaluated by the methodology similar to previous studies (Qu et al. 2013; Gao et al. 2014; Hu et al. 2019; Li et al. 2019):

\[
\frac{\partial [S]}{\partial t} = \text{ADV} + F + \text{Res}, \tag{1a}
\]

\[
\text{ADV} = \frac{1}{V} \int \int \left[ -\nabla \cdot (S\mathbf{v}) \right] dV, \tag{1b}
\]

\[
F = \left[ S \right] \frac{\int \int (E - P) dA}{V}, \tag{1c}
\]

where \([S]\) is the salinity averaged in the upper 200 m (S200); \(t\) denotes the time; \(V\) and \(A\) are the volume and surface area of the domain, respectively; \(\nabla\) is the horizontal gradient operator; and \(v\) is the horizontal velocity, including zonal and meridional velocity. The term ADV is the salinity advective flux. The term \(F\) is net air–sea freshwater flux, mainly controlled by the difference between evaporation and precipitation \((E - P)\) over the ocean surface; Res is the residual of the salinity flux, representing diffusive processes at the bottom and sides of a control volume.

To diagnose the effect of ocean dynamics on salinity variability, the salinity advection term can be further decomposed into four parts:

\[
\frac{1}{V} \int \int \left[ -\nabla \cdot (S\mathbf{v}) \right] dV = \frac{1}{V} \int \int \left[ -\nabla \cdot (S\mathbf{v}) \right] dV + \frac{1}{V} \int \int \left[ -\nabla \cdot (S'\mathbf{v}) \right] dV + \frac{1}{V} \int \int \left[ -\nabla \cdot (S\mathbf{v}') \right] dV + \frac{1}{V} \int \int \left[ -\nabla \cdot (S'\mathbf{v}') \right] dV. \tag{2}
\]

In Eq. (2), the overbars and primes denote climatological mean and deviation from the climatological mean, respectively. The first and fourth terms on the right-hand side are the mean advection and the nonlinear term, respectively. The second and third terms denote the salinity gradient advected by the ocean current, which contains the effect of mean salinity gradient

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advected by the anomalous currents \( ((\partial \bar{S}/\partial x)u') \); \( ((\partial \bar{S}/\partial y)y') \) and the anomalous salinity gradient advected by the mean currents \( ([\partial \bar{S}/\partial x]u); ([\partial \bar{S}/\partial y]y) \), respectively.

3. Upper-ocean salinity variability in the SEIO

a. The features of the decadal variability

Previous studies found that the upper-ocean salinity in the SEIO experienced a freshening trend during 2005–13 (Du et al. 2015; Nie et al. 2020), which is well captured in Fig. 1. The salinity decreasing trend markedly appears in the upper 200 m, with a slight increasing trend \((-0.01 \text{ psu yr}^{-1}\) in 200–500 m (Fig. 1b). The vertical structure of the salinity trend indicates that the anomalous freshening took place at the area where the climatological salinity front is (i.e., \(12^\circ\text{S–30}^\circ\text{S}; \text{Fig. 1c}\). The spatial distribution of this decadal salinity variation is distinct from the long-term trend (Helm et al. 2010) and interannual variability of salinity (Zhang et al. 2013) in this region. Therefore, the mechanism of this decadal salinity variability is unique and worth exploring.

To further investigate the decadal salinity variation in the SEIO, we extend the time series of S200 from 1992 to 2015 based on the ECCO data. Figure 2a shows a clear quasi-decadal variability of the salinity, featuring prominent salinification periods (1992–98, 2002–05, and 2013–15) and freshening periods (1998–2001 and 2005–13). By contrast, the amplitude of the interannual variation is much smaller. The decrease in salinity during 2005–13 is in one of the declining periods of this cycle. Besides, the decadal anomalies of the SEIO salinity are associated with the preceding meridional ocean currents in the SEIO (Figs. 2b,c). The positive phase (1994–2000 and 2004–11) and negative phase (2000–04 and 2011–15) of the SEIO salinity decadal variability correspond to the anomalous northward and southward ocean currents during salinification periods (1992–98, 2002–05, and 2013–15) and freshening periods (1998–2001 and 2005–13), respectively. These anomalous currents cross the strong zonal salinity front in the SEIO region and may play an important role in the decadal variability of salinity in the SEIO.

b. Salinity budget analysis

To understand what dominant factors control the decadal salinity variability in the SEIO, we perform the salinity budget analysis to quantify the contribution of each term (Fig. 3). The result shows that the salinity tendency is almost overlapped with the sum of all terms, indicating that the result from the budget analysis is valid. Compared to the freshwater flux, the advection term dominates the salinity variation during the freshening and salinification periods (Figs. 6a,e,i,m,q), with the meridional advection contributing larger than the zonal advection (Figs. 6c,g,k,o,s). For the freshwater flux, the role of precipitation is more important than evaporation (Figs. 6b,f,j,n,r).

Composite maps during the freshening and salinification periods clearly show the different contributions between advection and freshwater flux. The spatial distributions of
the S200 tendency term (Figs. 4a,b) are quite similar to the results from advection term (Figs. 4c,f) during the salinification and freshening periods. This result indicates the dominant role of the advection in the decadal salinity variability in the SEIO. The advection includes zonal and meridional advection. To distinguish the contribution of the two terms, we decompose the advection into the zonal and meridional advection following Eqs. (1b) and (2). The result illustrates that the meridional advection term is dominant in the salinity variation (Fig. 3c). Further analysis suggests that the \( \frac{\partial S}{\partial y} \) term, denoting the mean salinity gradient advected by anomalous meridional current, contributes to the major part of the meridional advection (Fig. 5). Indeed, there is a prominent meridional salinity gradient in the SEIO (contours in Fig. 2b), which is associated with the high-salinity band induced by intense evaporation in the subtropical high-salinity water regions (Toole and Warren 1993; Pokhrel et al. 2012) and the low-salinity water mass induced by local precipitation and freshwater transport by the ITF in the ITF outflow region (Qu and Meyers 2005).

Previous studies suggested that the anomalous meridional salinity gradient modulates the SEIO salinity through the mean meridional current [i.e., the \( \frac{\partial S}{\partial y} \) term; Zhang et al. 2013; Du et al. 2015]. In addition, the \( \frac{\partial S}{\partial x} \) term appears to be an important component of the horizontal advection term in the SIO (Figs. 5e,f). The terms \( \frac{\partial S}{\partial y} \) and \( \frac{\partial S}{\partial x} \) indicate the meridional and zonal salinity advection by the Leeuwin Current and the ITF, respectively (Zhang et al. 2016; Zhang et al. 2018b; Hu et al. 2019). Both of them, however, cannot effectively affect the SEIO salinity. Considering that the Leeuwin Current is a boundary current, the \( \frac{\partial S}{\partial y} \) term is only evident in the coastal region (Figs. 5a,b). The ITF brings the salinity anomalies from the Indonesian Seas into the central Indian Ocean instead of SEIO. Thus, the \( \frac{\partial S}{\partial x} \) term can only affect the salinity in the north of the SEIO (Figs. 5e,f). Due to the weak mean zonal salinity gradient, the anomalous zonal advection associated with \( \frac{\partial S}{\partial x} \) does not significantly impact the SEIO salinity, whose contribution is limited within a narrow band around 10°S (Figs. 5g,h). In addition, the nonlinear advection terms [i.e., \( \frac{\partial S}{\partial y} \) and \( \frac{\partial S}{\partial x} \)] have considerable influences in the tropical and northern Indian Ocean but not in the SEIO region (Figs. 6d,h,l,p,t).

To reveal the detailed reason responsible for the cause of decadal salinity variability in the SEIO, we calculate the salinity budget in each salinification and freshening period.
Fig. 6. During the freshening (salinification) period, advection term accounts for \(-90\%\) \((-91\%)\) of the salinity decrease (increase) in the SEIO. The advection term is dominated by the meridional advection (Figs. 6c,g,k,o,s). Further decomposition of horizontal advection confirms that the \((\partial S/\partial y)u'\) term dominates the total variability of the advection term. Other advection terms \([((\partial S/\partial y)\eta), (\partial S/\partial \alpha)\eta, and (\partial S/\partial \alpha)u']\) are relatively small during all periods (Figs. 6d,h,l,p,t). Overall, \((\partial S/\partial y)u'\) dominates the variability of the salinity tendency in the SEIO.

In comparison with the contribution of the advection terms, the effect of freshwater flux term is secondary, although its contribution is also positive (Du et al. 2015; also see Figs. 3a and 4c,d). The freshwater flux term is mainly dominated by precipitation in SEIO (Figs. 3b and 6b,f,j,n,r), as well as other tropical oceans (Schmitt 2008). The contribution of freshwater flux to the salinity is considerable around the MC instead of the SEIO region.

4. The local oceanic dynamics associated with the salinity decadal variability

The salinity budget diagnosis suggests that the mean meridional salinity gradient advected by the anomalous meridional currents \([(\partial S/\partial y)u']\) dominates the quasi-decadal variability in the SEIO. The tendency of salinity is significantly correlated with the meridional current anomaly \((r = 0.9)\) as well as the anomalous geostrophic meridional velocity \((r = 0.8)\), indicating that the anomalous meridional velocity may be induced by the local geostrophic adjustment (Fig. 7a). Previous studies (Lee 2004; Lee and McPhaden 2008; Meng et al. 2020) pointed out that zonal sea surface height (SSH) gradient in the SIO is key in interannual/decadal variability of meridional geostrophic transport, especially meridional geostrophic transport anomalies in the upper ocean (Nagura 2020). Here we define an SLA dipole pattern index as the SLA difference between the western \((H_W; 5^\circ-20^\circ S, 52^\circ-90^\circ E)\) and eastern \((H_E; 5^\circ-20^\circ S, 110^\circ-120^\circ E)\) southern Indian Ocean region. Both \(H_W\) and \(H_E\) are adjacent to the SEIO. The correlation coefficient between the anomalous meridional velocity in the SEIO and the SLA dipole pattern index is 0.88 (Fig. 7b), which is significant at 95% confidence level. The correlation coefficient between the SLA dipole pattern index and \(H_W\) \((H_E)\) is 0.97 \((-0.96)\). The partial correlation analysis further confirms that \(H_E\) and \(H_W\) both contribute to the correlation with meridional currents within the SEIO, with correlation coefficients of \(-0.6\) and 0.5, respectively.
The zonal SLA pattern is well captured in Fig. 8. During the salinification (freshening) period, the SLA is positive (negative) in the southwestern Indian Ocean but negative (positive) in the Indonesian Seas and the west coast of Australia.

The local atmospheric circulation is distinct during the salinification and freshening periods. In the SIO, an anomalous anticyclone (cyclone) appears in the east of 70°E during the salinification (freshening) periods (Figs. 4c,d). Thus, the HW is forced by the wind stress curl in the SIO (e.g., Zhang et al. 2013; Volkov et al. 2020). Previous studies pointed out that HW is induced by the westward propagating Rossby wave forced by the wind stress curl over the SIO from the seasonal to interannual time scale (Xie et al. 2002; Masumoto and Meyers 1998; Trenary and Han 2013; Zhuang et al. 2013). Indeed, the wind stress curl in the central-to-eastern SIO (CESIO; 5°–20°S, 90°–120°E; red box in Fig. 9a) is significantly correlated with HW with a correlation coefficient of 0.86. A lead–lag correlation analysis indicates that the wind stress curl over the CESIO leads HW by 2 months with a maximum correlation coefficient of 0.88 (Fig. 9c). It is worth noting that this lag time is much less than that of the typical Rossby wave at these latitudes (5°–20°N; see the appendix), because the wind stress curl can also simultaneously influence HW anomaly by forcing local Ekman pumping velocity (Fig. 9a).

Next, we use a simple 1.5-layer reduced-gravity model governed by the linear vorticity equation (see the appendix) to simulate HW due to the Ekman upwelling from local wind forcing and the westward propagating oceanic Rossby wave forced by wind stress curl from the east. This simulation can reproduce the observed HW variation with $r = 0.76$ (Figs. 9b,c). Meanwhile, we calculate the explained variance of the simulated HW for wind stress curl at different longitudes (Fig. 9f).

The result reveals that the HW variability is mainly associated with the wind forcing between 75° and 110°E. To demonstrate the role of wind stress curls over the SIO in changing HW, we further decompose the simulated HW into three parts: the
effects of the local wind stress, the westward propagating Rossby waves from CESIO, and the Rossby waves from the eastern boundary (Fig. 9d). The variability of HW is mainly contributed by the local wind stress and westward propagating Rossby waves from CESIO, while the signal from the eastern boundary is minor. The lead–lag correlations show that the simulated HW from local wind is simultaneously correlated with wind stress curls, while the simulated HW associated with the Rossby waves from CESIO delays wind stress curl for 6 months (Fig. 9e). The combined effect of them results in the wind stress curl over the SIO leading HW by 2 months. We also calculate the meridional geostrophic current in the SEIO by using the SLA obtained from the 1 1/2-layer reduced-gravity model. The simulated meridional geostrophic current is also highly correlated with the zonal gradient of SLA and meridional geostrophic velocity anomaly calculated by the oceanic

FIG. 6. S200 budget (10^-3 psu month^-1) in the SEIO: (a) the S200 tendency term (St), freshwater flux term (E – P), salinity advection term (ADV), and residual term (Res); (b) the surface freshwater budget including freshwater flux term (E – P), evaporation term (E), and precipitation term (P); (c) the salinity advection term (ADV) with zonal (ADVU) and meridional (ADVV) advection terms; and (d) further decomposition of advection term into (∂S/∂x)u (UaS), (∂S/∂y)y (VaS), (∂S/∂x)u (UsaS), (∂S/∂y)y (VsaS), (∂S/∂x)u (UsaS), and (∂S/∂y)y (VsaS) during 1992–97. (e)–(h), (i)–(l), (m)–(p), and (q)–(t) As in (a)–(d), but for 1998–2001, 2002–05, 2005–13, and 2013–15, respectively. The error bars represent standard deviation of different terms.
density (Fig. 7b). Overall, through oceanic dynamics process, the wind stress curl over the SIO controls $H_W$ variation, which is important for the salinity in the SEIO by modulating the anomalous meridional current.

By contrast, $H_E$ variability does not correspond to the surface wind over the IO and related Ekman pumping velocity. The $H_E$ is highly correlated with the tropical western Pacific (Wijffels and Meyers 2004; Feng et al. 2010, 2011; Li and Han 2015; Wang et al. 2015). Feng et al. (2004) suggested that the coastal sea level at Fremantle along western Australia is strongly influenced by trade wind variations in the tropical Pacific via equatorial and coastal waveguides. Furthermore, previous studies suggested that the tropical Pacific variability can induce the wind stress curl variability over the SIO, indicating that the SLA dipole pattern can be related to the interbasin teleconnections (Cai et al. 2019; Wang 2019; Sprintall et al. 2020). Nagura and McPhaden (2021) used a 1.5-layer model to examine SSH variability in the SIO. Their result showed that winds in the tropical Pacific Ocean are the main driver of SSH variability at midlatitudes of the SIO. The meridional geostrophic current induced by SLA dipole, which is essential for the decadal salinity variability in the SEIO, is connected with the climate variability in the tropical Pacific.

5. The linkage with interbasin teleconnection

The interbasin climate variability in the tropical Pacific, such as the ENSO and IPO, dominates the tropical atmospheric circulation on the interannual and decadal time scales (Folland et al. 2002; Dai 2013). Sea surface temperature (SST) anomalies in the central-to-eastern Pacific modulate the Walker circulation via zonally shifting the convective center over the Indo-Pacific Ocean. Warm (cold) SST anomalies in the eastern equatorial Pacific lead to westerly (easterly) wind anomalies in
the equatorial Pacific and easterly (westerly) wind anomalies over the equatorial Indian Ocean.

The $H_E$ and $H_W$ are strongly associated with the zonal wind responses in the equatorial Pacific and the Indian Ocean, respectively. Strengthening (weakened) trade winds in the Pacific lead to the high (low) SLA in the northwest Australia coast via the coastal Kelvin wave propagation (Feng et al. 2004). In contrast, anomalous westerly (easterly) winds over the equatorial Indian Ocean and positive (negative) wind stress curl over the CESIO lead to low (high) $H_W$ via the oceanic Rossby wave adjustment (Masumoto and Meyers 1998). The opposite SLA distribution between the $H_E$ and $H_W$ induces anomalous meridional currents in the SEIO region, which are crucial for the local salinity variability.

FIG. 9. (a) Correlation coefficients of wind stress curl (shaded) and surface wind (vectors) with $H_W$ in the black box. (b) Normalized $H_W$ from AVISO (blue) and 1.5-layer reduced-gravity model (black). (c) Lead–lag correlation coefficients of observed $H_W$ with simulated $H_W$ from 1.5-layer reduced-gravity model (solid line) and averaged wind stress curl anomaly over the CESIO (red box in Fig. 9a) (dashed line). The red line denotes 95% confidence level based on a two-tailed Student’s $t$ test. (d) Simulated $H_W$ (red) and its decomposition into the effects of the local wind stress (blue), the westward-propagating Rossby waves from CESIO (black), and the Rossby waves from the eastern boundary (green). (e) Lead–lag correlation coefficients of averaged wind stress curl anomaly over the CESIO with simulated $H_W$ and its decompositions from 1.5-layer reduced-gravity model. (f) Explained percent of variance [$S(X)$] of the modeled $H_W$ by the cumulative wind forcing from 52°E to longitude $X$. See appendix for details.
The tendencies of S200 and the zonal SLA dipole pattern index are highly correlated with the Niño-3.4 index (5°N–5°S, 170°–120°W) \( (r = 0.75 \text{ and } 0.72, \text{ respectively}; \text{Fig. 10a}) \). Lead–lag correlation coefficients between the zonal SLA dipole pattern index and Niño-3.4 index indicate that the central-to-eastern Pacific SST leads the SLA dipole pattern in the SIO by 2 months with a maximum correlation coefficient of 0.82 (Fig. 10b). This reflects that the wind anomalies from the Pacific result in the ocean dynamical adjustments in the SIO. To examine the SEIO salinity variability associated with the Pacific climate variability on different time scales, we apply an 8-yr Butterworth filter to separate the interannual and decadal variations of the zonal SLA dipole pattern index, the tendency of S200, and the Niño-3.4 index. On interannual time scale, the SST regression upon the zonal SLA dipole pattern index shows an El Niño–like warming pattern in the Pacific (Fig. 11a). In the Indian Ocean, SST warming mainly emerges in the west of the SEIO, indicating the crucial role of the downwelling Rossby wave forced by the Pacific variability in modulating the local thermocline (Xie et al. 2002, 2010; Zhang et al. 2020). It is notable that this SST pattern is different from the IOD, which has little effect on the SLA dipole pattern and salinity variability in the SEIO (not shown). This is because the atmospheric variations associated with IOD are much closer to the equator, while the atmospheric variations related to El Niño in the southeastern tropical Indian Ocean are located at higher
latitudes (Yu et al. 2005). The different atmospheric responses to the ENSO and the IOD lead to different SLA patterns and ocean currents in the Indian Ocean.

On the decadal time scale, correlation coefficients of the low-pass filtered Niño-3.4 index with the low-pass filtered zonal SLA dipole pattern index and SEIO S200 tendency is 0.7 and 0.73, respectively, which are both significant at 95% confidence level (Fig. 10c). The SST regression upon the low-pass filtered SLA dipole pattern index, which could lead to the quasi-decadal variation of the SEIO salinity, shows the significant SST warming over the central Pacific (Fig. 11b). Stuecker (2018) connected the quasi-decadal SST variability in the central Pacific with the North Pacific meridional mode, which is also well captured in Fig. 11b.

Although the dipole patterns of SLA are quite similar on interannual and decadal time scales, the latitude of decadal dipole mode is around 15°S, more southward than that of the interannual dipole mode (Fig. 11). The decadal SLA dipole pattern can induce geostrophic currents across the climatological salinity front to efficiently change the SEIO salinity, which is centered around 20°S.

Overall, the basin-scale climate variability in the Pacific causes a series of local ocean–atmospheric responses in the tropical Indian Ocean. On decadal time scale, the central Pacific SST variability forces the dipole-like SLA pattern in the SIO via interbasin teleconnections, leading to meridional salinity transport and the salinity variation in the SEIO.

6. Conclusions and discussion

There is significant decadal salinity variability in the SEIO. We use the salinity budget analysis to reveal that the decadal variability of the salinity in the upper 200 m is mainly contributed by the meridional salinity advection, and the freshwater exchange at the surface plays a secondary role. Further investigation reveals that the anomalous meridional current, a geostrophic response to the zonal SLA dipole pattern in the SIO, dominates decadal salinity variability in the SEIO. The tropical Pacific decadal variability forces this SLA dipole pattern via interbasin teleconnection. The related mechanisms are summarized in Fig. 12. During the negative phase of the tropical Pacific decadal variability (e.g., 2003–15), the cool SST in the eastern-to-central Pacific modulates the Walker circulation over the Indo-Pacific Ocean, leading to westerly wind anomalies over the equatorial Indian Ocean and easterly wind anomalies over the equatorial Pacific. On one hand, the local Ekman upwelling and the westerly wind over the equatorial Indian Ocean generate anomalous cyclonic wind over the CESIO, which forces upwelling Rossby waves to propagate westward and further decreases the $H_{100}$. On the other hand, the easterly wind over the equatorial Pacific leads to high SLA in the northwest Australia coast due to the equatorial and coastal waveguides. Overall, the zonal SLA dipole pattern in the SIO corresponds to an anomalous southward oceanic current, transporting the fresher seawater from the equatorial Indian Ocean to the SEIO, which decreases the local salinity.

This pronounced salinity variability can further influence oceanic stratification. Temperature and salinity are two major factors affecting oceanic stratification (Li et al. 2020). Over most of the Indian Ocean, the upper-ocean temperature is highly correlated with buoyancy frequency ($N^2$) (Fig. 13a), indicating its dominant role in oceanic stratification. However, in the SIO, the upper-ocean salinity is more important (Fig. 13b). Particularly in the SEIO, the correlation coefficient of $N^2$ with S200 is much higher than that with upper-ocean temperature (Fig. 13c). Since the change of oceanic stratification can further influence upper-ocean environment and ecology (Fu et al. 2016; DeVries et al. 2017; Breitburg et al. 2018), the climate impact of the SEIO decadal salinity variability is worthy of being studied in the future.

There are still several issues that need investigation in the future. Limited by the length of datasets, this study focuses on the decadal variability of salinity in the SEIO, while the salinity change on longer time scales cannot be investigated here. In particular, the long-term trend associated with anthropogenic warming is another important signal of oceanic salinity change (Sun et al. 2021). In the tropical Pacific, an El Niño–like warming pattern emerges under global warming (Zheng 2019). Whether this warming pattern can change the SEIO salinity via interbasin teleconnection is still unknown. In addition, eddy-induced meridional salinity flux plays an essential role in the salinity climate pattern in the SIO (Qu et al. 2019). It is admitted that this mesoscale process on the decadal salinity variability cannot be properly resolved by the current
observational and reanalysis datasets. High-resolution models should be used to investigate this issue further.

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APPENDIX

1.5-Layer Reduced-Gravity Model

Previous studies have shown that large-scale SLA variations in the open ocean can be dominated by the local wind-driven Ekman pumping and Rossby wave (Meyers 1979; Masumoto and Meyers 1998). The 1.5-layer reduced-gravity model can be used to quantify the wind-induced SLA variability (e.g., Meyers 1979; Kessler 1990; Capotondi and Alexander 2001; Capotondi et al. 2003; Qiu and Chen 2010). Under the long-wave approximation, the linear vorticity equation governing the 1.5-layer reduced-gravity model is given by

$$\frac{\partial h'}{\partial t} - C_R \frac{\partial h'}{\partial x} = -\frac{g}{\rho_0 g'} \nabla \times \tau - \varepsilon h', \quad (A1)$$

where $h'(x, y, t)$ is the SLA; $C_R$ is the zonal phase speed of long baroclinic Rossby wave, and we use the values derived...
by Chelton et al. (1998). In the 5°–20°S band, \( C_R \) has values ranging from 40 cm s\(^{-1} \) (5°S) to 8 cm s\(^{-1} \) (20°S). Also, \( g' \) is the reduced gravity, which is 0.06 m s\(^{-2} \); \( \rho_0 \) is the reference density; \( \tau \) is the anomalous wind stress; and \( \epsilon = 3 \) years is the Newtonian damping coefficient. The wind stress data are derived from ERA-Interim, and \( h'(x, y, t) \) can be solved by integrating Eq. (A1) from the eastern boundary \( (x = x_e) \) along the baroclinic Rossby wave characteristic. The solution to this equation yields

\[
h'(x, y, t) = \frac{g'}{\rho_0 g} \int_{x}^{x_e} \frac{1}{C_R} \nabla \times \left( x', y, t + \frac{x - x'}{C_R} \right) \times \exp \left( \frac{\epsilon}{C_R} (x - x') \right) dx' + h \left( x_e, y, t + \frac{x - x_e}{C_R} \right) \times \exp \left( \frac{\epsilon}{C_R} (x - x_e) \right),
\]

(A2)

where \( h'(x, y, t) \) means the SLA deduced from the wind-driven, 1.5-layer reduced-gravity model. In Eq. (A2), we use observed SSH as the eastern boundary forcing \( h(x_e, y, t) \). With this solution, we compute SLA by integrating the Ekman pumping forcing from the ERA-Interim monthly wind stress data, along with the Rossby wave characteristics.

To quantify the contribution of wind stress curl at different longitudes to \( H_W \), we calculate the explained variance (Qiu and Chen 2010), which is defined as

\[
S(X) = 1 - \frac{\langle (H'_W(x, t) - \overline{H'_W(X, t)})^2 \rangle}{\langle H'_W(X, t) \rangle^2},
\]

(A3)

\[
\overline{H'_W(X, t)} = \frac{g'}{\rho_0 g} \int_{x}^{x_e} \frac{1}{C_R} \nabla \times \left( x', y, t + \frac{x - x'}{C_R} \right) \times \exp \left( \frac{\epsilon}{C_R} (x - x') \right) dx',
\]

(A4)

where \( \overline{H'_W(X, t)} \) denotes the averaged \( H_W \) forced by the wind stress curl from 52°E to the west of longitude \( X \). The longitude of the eastern boundary \( (x_e) \) is 120°E.

REFERENCES


