Decadal SST Variability in the Southeast Indian Ocean and Its Impact on Regional Climate

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ABSTRACT

The southeast Indian Ocean (SEIO) exhibits decadal variability in sea surface temperature (SST) with amplitudes of ~ 0.2 – 0.3 K and covaries with the central Pacific ($r = -0.63$ with Niño-4 index for 1975–2010). In this study, the generation mechanisms of decadal SST variability are explored using an ocean general circulation model (OGCM), and its impact on atmosphere is evaluated using an atmospheric general circulation model (AGCM). OGCM experiments reveal that Pacific forcing through the Indonesian Throughflow explains $<20\%$ of the total SST variability, and the contribution of local wind stress is also small. These wind-forced anomalies mainly occur near the Western Australian coast. The majority of SST variability is attributed to surface heat fluxes. The reduced upward turbulent heat flux (Q_T ; latent plus sensible heat flux), owing to decreased wind speed and anomalous warm, moist air advection, is essential for the growth of warm SST anomalies (SSTAs). The warming causes reduction of low cloud cover that increases surface shortwave radiation (SWR) and further promotes the warming. However, the resultant high SST, along with the increased wind speed in the offshore area, enhances the upward Q_T and begins to cool the ocean. Warm SSTAs co-occur with cyclonic low-level wind anomalies in the SEIO and enhanced rainfall over Indonesia and northwest Australia. AGCM experiments suggest that although the tropical Pacific SST has strong effects on the SEIO region through atmospheric teleconnection, the cyclonic winds and increased rainfall are mainly caused by the SEIO warming through local air–sea interactions.

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1. Introduction

The southeast Indian Ocean (SEIO) region located to the west of the Australian continent exhibits prominent oceanic and atmospheric variations that are linked to important climate phenomena such as El Niño–Southern Oscillation (ENSO) (e.g., Feng et al. 2004; Cai et al. 2005; Wijffels and Meyers 2004). This region is receiving increasing attention owing to the recent

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invigoration of a local interannual climate mode, namely, the Ningaloo Niño/Niña, which manifests as sea surface temperature (SST) warming/cooling near the Western Australian coast and coincides with large-scale anomalies of low-level winds and rainfall (e.g., [Feng et al. 2013](#); [Doi et al. 2013](#); [Kataoka et al. 2014](#); [Tozuka et al. 2014](#); [Marshall et al. 2015](#)). Particularly, the unprecedented Ningaloo Niño event in 2010/11 austral summer emerged with SST anomalies (SSTAs) of >3.0 K and caused devastating consequences to local marine ecosystems including massive coral bleaching and dramatic changes in biodiversity patterns ([Depczynski et al. 2013](#); [Pearce and Feng 2013](#); [Wernberg et al. 2013](#)). This event was succeeded by two weaker but also influential warming events in the following two austral summers, exerting persistent stress on local environment ([Feng et al. 2015](#); [Zhang et al. 2017](#)). Decadal SST variability in the SEIO is invoked to explain the reemergence of Ningaloo Niño/Niña. Specifically, the rapid decadal warming of the SEIO under La Niña-like condition of the Pacific climate along with anthropogenic forcing was in favor of the development of successive Ningaloo Niño events during 2010–13 (e.g., [Feng et al. 2015](#); [Zinke et al. 2014](#)). Therefore, investigating the causes for decadal variability of the SEIO is important for understanding and predicting Ningaloo Niño/Niña events and their environmental impacts.

Owing to the extensive study of the Ningaloo Niño/Niña phenomenon, many processes affecting SST variability of the SEIO have been put forward. Generally, ENSO is believed to play a major role in generating SSTAs in the SEIO (e.g., [Feng et al. 2013](#); [Benthuyens et al. 2014](#); [Kataoka et al. 2014](#)); recent studies, however, point out that some Ningaloo Niño/Niña events can develop without ENSO effect ([Kataoka et al. 2018](#); [Zhang et al. 2018](#)). The tropical Pacific affects the SEIO through both oceanic and atmospheric teleconnections. The enhanced Pacific trade winds during La Niña excite oceanic Rossby and coastally trapped waves that penetrate through the Indonesian Archipelagos and reach the SEIO ([Clarke and Liu 1994](#); [Meyers 1996](#); [Cai et al. 2005](#); [Feng et al. 2004, 2010](#); [Wijffels and Meyers 2004](#)), which cause SST warming by suppressing coastal upwelling and enhancing warm-water advection of the Indonesian Throughflow (ITF) and the Leeuwin Current (a poleward-flowing shallow warm-water current along the Western Australian coast). Meanwhile, cyclonic wind anomalies are generated in the SEIO during La Niña conditions, possibly as the response to the equatorial Pacific SST cooling ([Feng et al. 2013](#); [Tozuka et al. 2014](#)). The wind cyclone involves northerly wind anomalies near the Western Australian coast that drive onshore Ekman transport and thus coastal downwelling, enhance the Leeuwin Current, and weaken the

prevailing southerly winds. All of the three processes are in favor of SST warming ([Feng et al. 2013](#); [Kataoka et al. 2014, 2017](#); [Zhang et al. 2018](#)). In addition to the Pacific forcing, local air–sea interaction is also important for the amplification of SSTAs and atmospheric anomalies. Several positive feedback regimes have been proposed, including the coupling between reduced wind speed, suppressed evaporation, and warm SST [wind–evaporation–SST (WES) feedback; [Marshall et al. 2015](#); [Feng et al. 2015](#)], the “coastal Bjerknes feedback” between coastal northerly winds, thermocline deepening, and warm SST ([Kataoka et al. 2014](#); [Kido et al. 2016](#)), and the cloud–radiation–SST feedback between reduced cloud, increased surface solar radiation, and warm SST ([Tozuka and Oettli 2018](#); [Zhang et al. 2018](#)). Moreover, [Kataoka et al. \(2017\)](#) pointed out that the surface mixed layer depth is shallower during Ningaloo Niño, which may also promote or maintain SST warming. [Zhang et al. \(2018\)](#) suggested that advection of surface air temperature and water vapor by anomalous winds also contributes to the SST variability.

Despite the rapid integration of our knowledge, it is still difficult to envision whether the processes reviewed above are also at work on longer time scales. Decadal SST variability with typical periods longer than 7 years may or may not share the spatial structure and mechanisms of Ningaloo Niño/Niña. The relative importance of different processes probably varies with time scale. Recent researches have revealed prominent decadal variations of the SEIO in sea level, circulation, and ocean heat content, and remote wind forcing effects from the tropical Pacific act as the primary driver (e.g., [Feng et al. 2004, 2010, 2016](#); [Han et al. 2014a](#); [Lee and McPhaden 2008](#); [Li and Han 2015](#); [Llovel and Lee 2015](#); [Han et al. 2017a](#); [Li et al. 2017, 2018](#)). Yet comparing with these variables, SST is more sensitive to the local atmospheric changes and may have different variability mechanisms.

More importantly, SSTAs in the SEIO are able to cause large-scale changes in winds and precipitation. During Ningaloo Niño years, low sea level pressure (SLP) and cyclonic winds emerge in the SEIO, and Western Australia receives more precipitation ([Doi et al. 2015](#); [Kataoka et al. 2014](#); [Marshall et al. 2015](#); [Tozuka et al. 2014](#)). It is of interest to examine whether the ocean–atmosphere covariance exists on decadal time scale and to what extent the atmospheric anomalies are caused by remote forcing from the Pacific and by local SST forcing in the SEIO. In particular, an intriguing inter-basin coupling mechanism between the SEIO and the central tropical Pacific was recently proposed, in which the impact of the SEIO SSTAs on ENSO is underlined ([Zhang and Han 2018](#)). This mechanism has not yet been examined for decadal variability.

The present study aims to provide the first comprehensive investigation of the causes for decadal SST variability in the SEIO and its impacts on climate (e.g., atmospheric circulation and precipitation). To achieve the goal, we first use an ocean general circulation model (OGCM) to explore the mechanisms of SST variability by evaluating the effects of different forcing and feedback processes, analogous to Zhang et al. (2018) but for decadal time scale. Then we examine the impacts of SST variability on the atmosphere using an atmospheric general circulation model (AGCM), similar to Tozuka et al. (2014) but for decadal variability. The rest of the paper is organized as follows. Section 2 provides a brief description of datasets and models. Section 3 describes decadal SSTAs observed in the SEIO and explores the causes. Section 4 reports the results of AGCM experiments and evaluates the impacts of decadal SSTAs on the atmosphere. Section 5 provides a summary and discussion.

2. Data and models

a. Datasets

We utilize the $1^\circ \times 1^\circ$, monthly Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) data (Rayner et al. 2003) of 1955–2017 to evaluate the observed SST variability over the SEIO and calculate climate indices. Niño-3 and Niño-4 are used to represent the SST condition of the eastern (5°S – 5°N , 150° – 90°W) and central (5°S – 5°N , 160°E – 150°W) equatorial Pacific, respectively. The interdecadal Pacific oscillation (IPO) index is computed using the “tripole” method (Henley et al. 2015), which is the SSTA difference between the equatorial Pacific (10°S – 10°N , 170°E – 90°W) and the northwest plus southwest Pacific regions (25° – 45°N , 140°E – 145°W plus 50° – 15°S , 150°E – 160°W). Atmospheric data of the 1° European Centre for Medium-Range Weather Forecasts (ECMWF) twentieth-century reanalysis (ERA-20C) from 1955 through 2010 (Poli et al. 2016) are also used in our analysis. The ERA-20C data are extended with ECMWF interim reanalysis (ERA-Interim) (Dee et al. 2011) from 2011 through 2017. For each oceanic or atmospheric variable, the climatologic seasonal cycle and linear trend are removed to obtain the anomaly, and then a 5-yr low-pass Hanning-window digital filter is applied to highlight decadal variability. Here a relatively narrow-filter window is chosen to avoid further lowering the effective degree of freedom. Using a 7- or 9-yr window does not affect our conclusions.

b. OGCM

To understand the mechanisms of SST variability, we utilize Hybrid Coordinate Ocean Model (HYCOM)

version 2.2.18 (Bleck 2002; Halliwell 2004). The model domain is the Indo-Pacific Oceans within 55°S – 50°N , 19°E – 68°W (Li et al. 2017, 2018). The zonal resolution is $\frac{1}{3}^\circ$ between 70° and 170°E and gradually changes to 1° west of 40°E and east of 160°W , while the meridional resolution is $\frac{1}{3}^\circ$ at low latitudes (25°S – 25°N) and gradually changes to 1° at midlatitudes (poleward of $\sim 44^\circ$). There are 35 hybrid vertical layers, and the top-layer thickness is 5 m. On the northern and southern open-ocean boundaries, 5° sponge layers are adopted, in which model temperature and salinity are relaxed to *World Ocean Atlas 2013* (WOA13) monthly climatology (Locarnini et al. 2013; Zweng et al. 2013).

Atmospheric fields of ERA-20C are used as the surface forcing, including surface wind stress, 10-m wind speed, surface net shortwave and longwave radiations (SWR and LWR), precipitation, and 2-m air temperature and humidity. Monthly climatologic river discharge data of Dai et al. (2009) are used as lateral freshwater flux forcing. To suppress the long-term drifting of model salinity, a weak sea surface salinity nudging toward monthly climatology is applied. In our setting, wind stress and wind speed are separately exerted onto the model ocean. They affect SST through different processes. Wind stress drives ocean dynamical processes such as advection, upwelling, and turbulent mixing, whereas wind speed affects SST through surface turbulent heat flux (latent plus sensible heat fluxes). Rather than prescribed, turbulent heat fluxes and evaporation are calculated with the model SST in an online manner using the Coupled Ocean–Atmosphere Response Experiment, version 3.0 (COARE 3.0), algorithm (Fairall et al. 2003; Kara et al. 2005). A 30-yr spin-up run under monthly climatologic forcing was first performed to allow the model to reach a quasi-equilibrium stage, and then the model was integrated forward from 1940 to 2010 using daily ERA-20C fields. This experiment is named as “Main Run” (MR). The MR is supposed to contain complete processes for SST variability and compared with observational data to evaluate the model performance. Three additional experiments (Table 1) were performed to evaluate effects of different processes (see section 3 for description).

c. AGCM

To evaluate the impact of SST variability on regional climate, we also performed AGCM experiments. The Max Planck Institute (MPI) ECHAM, version 4.6 (Roeckner et al. 1996), is utilized. The model has a T42 horizontal resolution (corresponding to 2.8125°) over the global domain. It has 19 vertical levels, and the variables at 1024 hPa are taken as the surface fields. ECHAM is forced with monthly HadISST data and

TABLE 1. HYCOM and ECHAM experiments performed in the study.

	Model	Period	Ensemble	Forcing
MR	HYCOM	1940–2010	1	Daily forcing
PAC	HYCOM	1940–2010	1	Daily forcing in the Pacific, climatologic forcing in the Indian Ocean
TAU	HYCOM	1940–2010	1	Daily wind stress, other forcing fields are climatology
WND	HYCOM	1940–2010	1	Daily wind stress and wind speed, other forcing fields are climatology
CTR	ECHAM	1974–2015	10	Monthly SST forcing
TPR	ECHAM	1974–2015	10	Monthly SST forcing in the tropical Pacific, climatologic SST forcing in other areas
SEIR	ECHAM	1974–2015	10	Monthly SST forcing in the SEIO region, climatologic SST forcing in other areas

integrated from 1974 to 2015. We discard the output of the first year and use the output data of 1975–2015 for analysis. In addition to the Control Run (CTR), which uses realistic monthly SST forcing over the global atmosphere, we performed other two experiments to evaluate effects of regional SSTAs (Table 1), which are described in section 4. Each of the three experiments has ten ensemble members using slightly different initial conditions but the same SST forcing. Ensemble mean of the ten members represents the response to SST forcing, while their spread quantifies the uncertainty arising from atmospheric internal instability.

3. Decadal SST variability in the SEIO

a. Characteristics

Superimposed upon the prominent interannual fluctuations, decadal variations of the observed SST are evident in the SEIO (Fig. 1a; 35°–15°S, 100°–116°E). Persistent warm SSTAs occurred during the periods of 1955–65, 1982–90, 1995–2002, and 2009–14, and cold SSTAs were seen during 1966–81, 1991–94, and 2003–08. Quantified by the 5-yr low-passed SSTAs, decadal SST variability has a typical amplitude of ~0.2–0.3 K, weaker than the ~0.5-K interannual variability. The simulated SEIO SSTAs by HYCOM MR are shown in Fig. 1b. HYCOM is able to reproduce the decadal SSTAs after the late-1970s. Prior to that period, the simulation is inconsistent with observation. One plausible reason for this mismatch is the unavailability of satellite data for constraining the ERA-20C reanalysis system. This argument is supported by the fact that surface winds from different reanalysis products agree since the 1970s and diverge before that (Han et al. 2017b). Because of large errors in forcing fields, HYCOM is not capable of realistically simulating the observed SST variability of that period. The observed SST data are also

questionable, given that in situ observations were rather sparse in the SEIO region prior to the 1970s. We also analyzed Extended Reconstructed SST (ERSST), version, 4 data (Huang et al. 2015), which show roughly consistent decadal anomalies with HadISST data (figure not shown). Hereafter, we will focus on the 1975–2010 period during which the correlations between HYCOM and observation are $r = 0.86$ and $r = 0.88$ for the filtered and unfiltered anomalies, respectively.

The statistical relationship between the SEIO SST and the Pacific climate is also examined (Fig. 1c). The observed SEIO SSTs show correlations of -0.34 , -0.20 , and -0.63 with the IPO, Niño-3, and Niño-4 indexes, respectively. Only the $r = -0.63$ one with Niño-4 exceeds the 95% significance. In this study, the significance test for correlation coefficients uses effective degree of freedom computed based on autocorrelation (Bretherton et al. 1999). The IPO contains signals of both tropical and subtropical variance, whereas the SEIO is mainly affected by the tropical Pacific (e.g., Clarke and Liu 1994; Meyers 1996; Cai et al. 2005; Feng et al. 2004), which may explain the low correlation with IPO. Evident differences between Niño-3 and Niño-4 arise mainly from strong eastern Pacific-type El Niño events such as the 1982/83 and 1997/98 ones (e.g., Yu and Kim 2013). Such events have strikingly large signatures in Niño-3 but not in Niño-4 or SEIO SST. Also because of these strong El Niño events, Niño-3 exhibits mainly interannual rather than decadal variability (figure not shown). As a result, the decadal SST of the SEIO achieves a better relationship with Niño-4 than with Niño-3. We also computed the lead-lag correlations, which do not dramatically differ from the instantaneous correlations. Therefore, the decadal SST warming (cooling) of the SEIO is closely associated with a cooling (warming) condition of the central Pacific, similar to the relationship between Ningaloo Niño/Niña and ENSO. HYCOM tends to produce a stronger

FIG. 1. Monthly SSTAs of the southeast Indian Ocean (SEIO; 35°–15°S, 100°–116°E) derived from (a) HadISST and (b) HYCOM MR; 5-yr low-passed anomalies are shown as thick lines. (c) 5-yr low-passed indexes of IPO, Niño-3, and Niño-4 computed with HadISST data. Correlation coefficients are computed for the period of July 1978–June 2008.

link between the SEIO and the tropical Pacific, showing $r = -0.60$, -0.24 , and -0.77 with IPO, Niño-3, and Niño-4, respectively. Our analysis of HYCOM and ECHAM experiments presented below will show that the remote forcing from the Pacific plays a minor role in causing the decadal variability, and therefore this bias may not greatly affect our conclusions.

Spatial patterns of the SST variability are also of interest. In Figs. 2a–f we display observed SST trends of the SEIO for the three SST rising periods (1979–84, 1993–99, and 2004–10) and three SST falling periods (1984–93, 1999–2004, and 2011–15). Strong SST trends

are seen in the Niño-4 region, mimicking the central Pacific-type ENSO structures. The tropical Indian Ocean shows persistent warming over the entire 1979–2015 period (e.g., Du and Xie 2008; Han et al. 2014b; Dong and Zhou 2014; Dong et al. 2014), regardless of the conditions of the tropical Pacific Ocean and the SEIO. Climate models suggest that the persistent warming of the tropical Indian Ocean is primarily induced by anthropogenic greenhouse gas forcing, which overcomes the effect of the tropical Pacific natural variability in dictating the overall SST trend of the tropical Indian Ocean (Dong and McPhaden 2017).

FIG. 2. SST trend maps for the periods of (a) 1979–84, (b) 1984–93, (c) 1993–99, (d) 1999–2004, (e) 2004–10, and (f) 2011–15, derived from HadISST data. Stippling indicates >95% statistical significance based on a Mann–Kendall test. Black and white rectangles denote the SEIO and Niño-4 regions, respectively. (g)–(k) SST trends (color shading) of HYCOM MR and surface wind trends (vectors) of ERA-20C data for the periods of (g) 1979–84, (h) 1984–93, (i) 1993–99, (j) 1999–2004, and (k) 2004–10.

The SST trend patterns along with the tight relationship between the SEIO and the central Pacific are well reproduced by HYCOM MR (Figs. 2g–k), albeit with stronger trend magnitudes. Surface wind trends are estimated with ERA-20C data. Warming (cooling) of the central Pacific causes westerly (easterly) wind trends in the western Pacific basin that are in favor of the cooling (warming) of the SEIO by attenuating (enhancing) the ITF and evoking upwelling (downwelling) oceanic waves (e.g., Feng et al. 2004, 2013; Wijffels and Meyers 2004; Lee et al. 2015; Li et al. 2017).

b. Mechanisms

Figures 1 and 2 suggest the fidelity of HYCOM MR in simulating decadal SST variability during the 1975–2010 period. Validations for sea level, circulation, ocean heat content, and Ningaloo Niño/Niña events of HYCOM

MR were presented in Li et al. (2017, 2018) and Zhang et al. (2018). We hereafter use HYCOM experiments (Table 1) to explore physical processes causing the SST variability. Unlike MR, which contains complete processes of SST variability, other experiments contain only one or several processes, and therefore SSTAs in them are much weaker than that of MR (Fig. 3a). The remote forcing effect of the Pacific via oceanic connection (the ITF) is evaluated by the “Pacific” (PAC) run, in which the forcing fields in the Indian Ocean are fixed to monthly climatology (containing only the seasonal cycle), whereas the Pacific Ocean is still forced by daily atmospheric forcing as in MR. The boundary between daily and climatological forcing is roughly at 136°E for the model domain south of 10°S. For the specific definition of the forcing in PAC, readers are referred to Fig. S1 of Li et al. (2018). The PAC run produces some of the decadal SSTAs in the SEIO that are roughly in

FIG. 3. Decadal SSTAs simulated by HYCOM experiments. (a) 5-yr low-passed SST of the SEIO derived from MR, PAC, TAU, and WND. The correlation coefficients of PAC, TAU, and WND with MR are 0.56, 0.68, and 0.86, respectively. (b)–(f) Regression of SSTAs of (b) MR, (c) PAC, (d) TAU, (e) WND, and (f) MR – WND (the difference between MR and WND) regressed onto the 5-yr low-passed SEIO SST of MR. Stippling indicates $>90\%$ significance. In (c), (d), regression of ERA-20C surface winds onto the 5-yr low-passed SEIO SST of MR are shown as black vectors.

phase with those of MR. Their correlation is $r = 0.56$, significant at 90% confidence level. However, its magnitude (0.03 K in standard deviation) is much weaker than that of MR (0.15 K). It means that even if we ignore the phase difference, the Pacific forcing effect through the ITF can explain at most 20% (0.03/0.15 K). Regression maps of SSTAs onto the normalized decadal SEIO SSTA show that the weak SSTAs in PAC are confined near the Western Australian coast (Fig. 3c). Some weak warming is discernible in the offshore area poleward of 20°S. In both MR and PAC, warming signatures in the off-equatorial western Pacific areas and the Indonesian Seas are seen.

In addition to Pacific winds, local wind forcing can also drive SST variability. Cyclonic winds appear in the SEIO region (Figs. 3c,d), and the associated northerly winds near the Western Australian coast are in favor of SST warming by inducing onshore Ekman transport

and coastal downwelling and by enhancing the warm-water advection from lower latitudes (Feng et al. 2013; Kataoka et al. 2014). The wind stress run (TAU) is to measure the effect of wind stress-driven oceanic processes, including both local winds and Pacific winds. Note that the local wind variability here also contains teleconnection signatures from the Pacific through the atmospheric bridge. In TAU we use daily wind stress over the entire model domain, and all the other forcing fields are fixed to monthly climatology. TAU produces slightly stronger SSTAs (0.04 K in standard deviation) than PAC, and its correlation with MR is $r = 0.68$. The Wind Run (WND) uses daily wind stress and wind speed, and other forcing fields are fixed to monthly climatology. By taking into account wind speed change, SSTAs in WND are strengthened to a standard deviation of 0.06 K, and its correlation with MR is $r = 0.86$, explaining $\sim 40\%$ of that in MR (0.06/0.15 K). The

FIG. 4. (a) Normalized monthly anomalies of surface net heat flux (Q_{net}) and SST tendency ($\partial\text{SST}/\partial t$) of the SEIO region. (b) Monthly anomalies of Q_{net} , turbulent heat flux (Q_T), longwave radiation (LWR), and shortwave radiation (SWR) of the SEIO. (c) Regression of SST onto the normalized decadal $\partial\text{SST}/\partial t$ of the SEIO. Stippling indicates $>90\%$ significance. Solid and dashed rectangles denote the eastern (35° – 15° S, 108° – 116° E) and western (35° – 15° S, 100° – 108° E) parts of the SEIO region. SWR and LWR are from ERA-20C data, while SST, $\partial\text{SST}/\partial t$, Q_{net} , and Q_T are derived from HYCOM MR.

reduced wind speed by northerly wind anomalies acts to enhance the SST warming in coastal areas (Fig. 3e). The difference between WND and PAC is mainly caused by local wind forcing effect (both wind speed and wind stress), and its contribution is stronger than Pacific wind forcing (0.04 K in standard deviation as computed with WND – PAC; $r = 0.62$ with MR). The results in Fig. 3 are not sensitive to the choice of filter. We also tried 7- and 9-yr low-pass filters, and the conclusions are not dramatically changed (Fig. S1 in the online supplemental material).

The large difference between MR and WND in SSTA is mainly due to surface heat flux changes excluded from WND (Fig. 3f). Note that in WND wind speed also affects SST through surface heat flux, which increases the SST variability from 0.04 K in TAU to 0.06 K in WND. But this portion is merely the linear effect of wind speed. A comparison between MR and WND suggests that wind speed explains less than half of the surface turbulent heat flux (Q_T) variability in MR, and MR – WND is more close to MR in Q_T (Fig. S2). Other factors such as air temperature (T_a), air humidity (q_a), and the nonlinear effect of different factors are also important for Q_T and SST changes.

The total effect of surface heat fluxes on SSTA is 0.12 K as approximately quantified by the standard deviation of MR – TAU, which accounts for $\sim 78\%$ of the

total variance. The surface net heat flux Q_{net} shows a good relationship with SST tendency ($\partial\text{SST}/\partial t$) in MR (Fig. 4a). Most of the SST warming (cooling) tendencies coincide with positive (negative) Q_{net} anomalies. Among the three components of Q_{net} (Fig. 4b), the surface turbulent heat flux Q_T plays the dominant role in Q_{net} variability. Q_T and Q_{net} have close amplitudes (14.48 W m^{-2} versus 16.9 W m^{-2} in standard deviation) and a correlation of $r = 0.97$. SWR also contributes to Q_{net} , with a standard deviation of 6.95 W m^{-2} and a correlation of $r = 0.50$ with Q_{net} . The variability of LWR is small (standard deviation is 3.57 W m^{-2} ; $r = -0.18$). Figure S3 shows the 5-yr low-passed heat fluxes, which exhibit similar relationships. To better understand the processes causing the growth of SSTAs, we regress oceanic and atmospheric fields onto $\partial\text{SST}/\partial t$ of the SEIO. Figure 4c shows the regression map of SST onto 5-yr low-passed $\partial\text{SST}/\partial t$ of MR. The initial growth of warm SSTAs occurs near the coast of northwest Australia accompanied with cold SSTAs in the ocean interior. It is interesting to see that there are cooling signatures along the Sumatra–Java coast in the tropical Indian Ocean mimicking a positive Indian Ocean dipole (IOD) condition (Saji et al. 1999). The relationship and possible interaction between the IOD and the SEIO are, however, beyond the scope of this research. Figure 5 compares the SSTAs of the eastern

FIG. 5. 5-yr low-passed SSTA of the (a) eastern and (b) western parts of the SEIO region. Results from MR, PAC, TAU, and WND are plotted in different colors.

and western parts of the SEIO region. The SSTA in the eastern box (coastal region) tends to lead that of the west by 1–2 yr. HYCOM experiments suggest that wind-forced processes play an essential role in the eastern box (58%), in which the ITF and wind speed-controlled Q_T are two major contributors (Fig. 5a). The situation is radically different in the western box, where wind-forced processes explain only a small portion of SSTA (Fig. 5b).

By regressing atmospheric fields onto $\partial\text{SST}/\partial t$ and SST of the SEIO (Fig. 6), we gain insights into the processes at work during the growing stage and peak stage of warm SSTAs, respectively. In the growing stage, the entire region eastern Indian Ocean and western Pacific is controlled by high SLP, and anticyclonic winds are seen in the SEIO region (Fig. 6a). The northeasterly wind anomalies along the coast may be generated and maintained by two effects. The first is local coupling with the coastal SST warming as shown in the eastern part of SEIO (Fig. 4c). The warming, which may be initialized by the ITF, enhances local deep convection and induces the northeasterly winds blowing from the Maritime Continent to the SEIO. The second is remote forcing by the central Pacific cooling, which induces prevailing easterly winds in the western Pacific and along the northwestern coast of Australia (Figs. 2g,i,k). These northeasterly anomalies promote coastal warming probably by driving coastal downwelling and poleward advection of warm water (Feng et al. 2013; Kataoka et al. 2014; Kido et al. 2016). These wind anomalies induce sea level rise through shoreward Ekman transport (Fig. S4). As a result, southwestward surface current anomaly emerges in geostrophic balance with the anomalous sea level gradient, bringing warm water from lower latitudes. In this stage, Q_T is the dominant process driving SST warming, whereas

SWR and LWR have limited effect (Figs. 6b–d). When the warm SSTAs are fully developed in the SEIO, low SLP and cyclonic winds are dominant over this region (Fig. 6e), which can promote the SST warming. Surface SWR shows strong positive anomalies in the SEIO, also favoring further SST warming (Fig. 6f). These effects are, however, offset by the strong cooling effect of Q_T (Fig. 6g). Therefore, the warm SST peak is the net result of the competition between the strong damping effect by Q_T and the maintaining effects by SWR plus wind-driven ocean dynamics.

The different roles played by SWR and Q_T are of interest and require further investigation. Cloud cover is the major factor that affects surface SWR into the ocean (Fig. 7). Tozuka and Oettli (2018) demonstrated that in the SEIO both low cloud and high cloud may affect the surface SWR and interact with local SST. Regressions onto $\partial\text{SST}/\partial t$ suggest that neither low cloud nor high cloud shows large anomalies during the growing stage of SSTAs (Figs. 7a–c). The total cloud cover is slightly increased in the northwest Australia region (Fig. 7a) in the form of high cloud (Fig. 7c). This is primarily the response of atmospheric deep convection to coastal SST warming in that region, as suggested by the enhanced rainfall (Fig. 7d). Increased high cloud and rainfall occur only at low latitudes probably because that region has sufficiently high mean SST for deep convection (e.g., Waliser et al. 1994; Johnson and Xie 2010; Tozuka and Oettli 2018). Note that these signatures are still insignificant at this stage with anomaly centers located northeast of the SEIO. Upon reaching the peak stage, the total cloud shows a dipole structure in the SEIO with increased cloud in the tropical region equatorward of 20°S and reduced cloud in the subtropics (Fig. 7e). The increased tropical cloud is again mainly contributed by high cloud (Fig. 7g), reflecting the

FIG. 6. (a) SLP (color shading) and surface winds (vectors), (b) SWR, (c) Q_T , and (d) LWR regressed onto 5-yr low-passed SEIO $\partial\text{SST}/\partial t$ of MR. Stippling indicates >90% significance. (e)–(h) As in (a)–(d), but regressed onto 5-yr low-passed SEIO SST of MR.

enhanced deep convection (Fig. 7h) by SST warming. It provides a damping effect (negative feedback) for the warm SSTAs by reducing surface SWR in this area (see the negative SWR anomalies in Fig. 6f). Beyond that area, most of the SEIO has reduced total cloud cover, which is contributed by both low cloud and high cloud (Figs. 7e–g). The reduction in low cloud cover is the typical response of subtropical atmosphere to SST warming, since warm SST destabilizes the lower troposphere and suppresses subsidence motion (e.g., Klein and Hartmann 1993; Li and Philander 1996). The reason for reduced high cloud cover is not evident. It

is possibly associated with the prevailing subtropical drying condition over the south Indian Ocean (Fig. 7h). Anomalies of drying and high cloud reduction (both indicating suppressed deep convection) are centered in the southwest Indian Ocean and extend eastward to the SEIO. It may be the response to SST cooling of the southwest Indian Ocean (Fig. 3b) or enhanced tropical rainfall via atmospheric teleconnection, which will be further discussed in section 4.

Primary processes modulating Q_T changes are discussed in Fig. 8. During the growing stage of SST warming, the northeasterly wind anomalies in the

FIG. 7. (a) Total cloud cover, (b) low cloud cover, (c) high cloud cover, and (d) precipitation rate regressed onto 5-yr low-passed SEIO $\partial\text{SST}/\partial t$ of MR. Stippling indicates >90% significance. (e)–(h) As in (a)–(d), but regressed onto 5-yr low-passed SEIO SST of MR. Climatological cloud cover fields are also plotted as black contours.

nearshore area reduce the total wind speed (Fig. 8a) and bring warm, moist air from the Maritime Continent (Figs. 8b,c), and their integral effect is to suppress the upward Q_T toward the atmosphere. In this study we did not perform extra experiments to evaluate the relative importance of wind speed, T_a , and q_a . It is likely the generation of T_a and q_a anomalies also rely on the advection of winds. Therefore, the nonlinear effect of winds, T_a , and q_a is likely important in driving the changes of surface Q_T and SST. In the western box of the SEIO, wind speed is also reduced between 30° and 20°S,

but SSTAs are weak or negative (Fig. 4c). The cold SSTa is likely more essential than the weakened wind speed for the suppressed Q_T , because the structure of positive Q_T anomalies resembles that of negative SSTAs in this region (comparing Figs. 4c and 6c). At the peak stage, the prevailing high SST in the SEIO (Fig. 3b) drives increased upward Q_T regardless of the warm and moist anomalies in surface air (Figs. 7e,f). The cyclonic winds in the SEIO cause increased wind speed in the offshore area (Fig. 7d), which contributes to the strong upward Q_T there (Fig. 6g).

FIG. 8. (a) Surface winds (vectors) and wind speed (color shading), (b) 2-m air temperature (T_a), and (c) 2-m specific humidity (q_a) regressed onto 5-yr low-passed SEIO $\partial\text{SST}/\partial t$ of MR. Stippling indicates $>90\%$ significance. (d)–(f) As in (a)–(c), but regressed onto 5-yr low-passed SEIO SST of MR. Climatological T_a and q_a fields are also plotted as black contours.

The results shown in Figs. 3–8 suggest that different forcing and feedback processes are at work in driving and damping decadal SSTAs in the SEIO. The SST warming occurs first near the Western Australian coast. The growth of warming anomalies is mainly through the suppressed upward Q_T due to reduced wind speed, advection of warm and moist air, and offshore cold SSTAs. Wind stress-driven ocean dynamics also contribute to the growth of the SST warming, as part of the positive Bjerknes feedback loop, but this effect plays a minor role. When large-scale SST warming is established in the SEIO, the cloud–radiation–SST feedback is triggered. The reduced cloud cover and increased SWR act to further promote or at least maintain the SST warming. However, the high SST and increased offshore wind speed (by the cyclonic winds) switches the Q_T feedback to negative (enhanced upward Q_T), providing a damping effect for warm SST and compensating positive feedbacks by SWR and wind stress-driven ocean dynamics. Therefore, in our results

surface Q_T can provide both positive and negative feedbacks to SSTAs, depending upon the specific atmospheric and oceanic conditions.

4. Impacts on regional climate

In this section we evaluate impacts of decadal SST variability on regional climate using ECHAM experiments. This also helps us to answer the question as to whether the atmospheric anomalies in SLP, winds, and rainfall are caused by remote SST forcing from the Pacific or local SST forcing in the SEIO. The simulated atmospheric circulation by ECHAM CTR is compared with observation (ERA-20C/ERA-Interim data) in Fig. 9. The SEIO region is characterized by southerly winds and low precipitation rate in climatology that are faithfully represented in ECHAM CTR. Figure 9c examines SLP variability over the 25° – 15° S, 80° – 110° E region as the center of the low SLP and cyclonic winds in the SEIO. The SLP variability is consistent between

FIG. 9. Climatological precipitation rate (color shading) and surface winds (vectors) for 1975–2015 derived from (a) ERA-20C/ERA-Interim data and (b) ensemble mean of ECHAM CTR. The red box in (b) marks the area with realistic wind forcing in TPR (20°S–20°N, 120°E–80°W), while the blue box marks that in SEIR (36°–12°S, 90°–120°E). (c) SLP anomaly of the 25°–15°S, 80°–110°E region [marked by a black rectangle in (a)] derived from CTR and ERA-20C/ERA-Interim data.

CTR and observation, and the ensemble mean monthly SLP yields a correlation of $r = 0.48$ with observation. For the 5-yr low-passed SLP, the correlation is 0.76. The fluctuation of the observed SLP generally lies within the spread of CTR members, suggesting the robustness of decadal variability relative to atmospheric internal instability. SLP, surface winds, and precipitation of the CTR ensemble mean are regressed onto the observed decadal SEIO SSTA (Figs. 10a,b). The regression patterns are overall consistent with those derived from ERA-20C data. Important variability features such as low SLP, cyclonic winds, and enhanced rainfall over the tropical SEIO region are faithfully reproduced.

ECHAM experiments (Table 1) are used to evaluate the effects of the tropical Pacific SSTAs and the SEIO SSTAs. The Tropical Pacific Run (TPR) retains realistic SST variability only in the tropical Pacific (20°S–20°N, 120°E–80°W; Fig. 9b), and in other areas SST forcing is fixed to monthly climatology. The results of TPR show large-scale low SLP anomalies over the

eastern Indian Ocean (Fig. 10c), as response of the Indian Ocean atmosphere to the western Pacific warming under La Niña-like condition. This confirms the impact of ENSO's atmospheric teleconnection on the Indian Ocean (e.g., Alexander et al. 2002; Klein et al. 1999; Venzke et al. 2000). However, the low SLP patch is too large in spatial scale to generate cyclonic winds in the SEIO, and the enhanced rainfall over the tropical SEIO seen in CTR and observation is absent in TPR (Fig. 10d). These climate anomalies are not likely atmospheric teleconnection signatures of the tropical Pacific SSTAs. Alternatively, the Southeast Indian Ocean Run (SEIR; realistic SST forcing in the 36°–12°S, 90°–120°E and monthly climatic SST in other areas) is able to produce the cyclonic winds and rainfall enhancement (Figs. 10e,f), suggesting the importance of local SST forcing in the SEIO in generating atmospheric variations. Diabatic heating of warm SSTAs in the SEIO enhances local rainfall by driving atmospheric deep convection and generates a low SLP center and cyclonic winds to its west through atmospheric Rossby wave

FIG. 10. (a) Surface winds (vectors) and SLP (color shading) and (b) precipitation rate of CTR regressed onto 5-yr low-passed SEIO SST derived from HadISST data. (c),(d) As in (a),(b), but for regression of TPR fields. (e),(f) As in (a),(b), but for the regression of SEIR fields. Stippling in (b),(d),(f) indicates $>90\%$ significance.

response. These results underscore the importance of local air–sea interaction, compared to remote forcing from the Pacific, in generating decadal climate variability (including both oceanic and atmospheric variations) in the SEIO.

We also note that the wind and rainfall anomalies in the SEIR are evidently weaker than those in observation and CTR. It is possible that atmospheric nonlinearity can amplify the anomalies when both tropical Pacific and local SSTAs are included. SST variability of other regions may also contribute to these atmospheric anomalies, such as those in the tropical Indian Ocean and the southwest Indian Ocean. We have seen drying anomalies in the southwest Indian Ocean in observation and CTR but not in TPR and SEIR. As mentioned in section 3, it is probably induced by underlying SST cooling co-occurring with the SEIO SST warming, in resemblance of the subtropical dipole mode proposed by Behera and Yamagata (2001). This drying may have impact on the SEIO by modulating the large-scale

atmospheric circulation. The impact of tropical Indian Ocean is also likely, given the large oceanic and atmospheric anomalies in that region coinciding with the SEIO variability (see Figs. 3, 4, 6, and 10). Zhang et al. (2018) showed that positive IOD condition is favorable for the generation of Ningaloo Niño. This effect may be also significant on decadal time scale, as indicated by the persistent positive IOD-like condition (cooling near the Sumatra–Java coast) during SEIO warming (Figs. 3b and 4c).

The TPR and SEIR have ten ensemble members, allowing us to evaluate the robustness of the SST forcing effects relative to atmospheric internal instability. In Fig. 11 we examine this issue for three key atmospheric variables: SLP in the SEIO, meridional wind along the Western Australian coast, and precipitation in the tropical SEIO. One can see that for both the TPR or the SEIR, standard deviation of ensemble members is rather large in comparison with the ensemble mean anomaly. For the coastal wind and precipitation,

FIG. 11. The 5-yr low-passed time series of (a) the SEIO SLP (25° – 15° S, 80° – 110° E), (b) alongshore meridional wind (30° – 5° S, 110° – 115° E), and (c) tropical precipitation (15° – 5° S, 90° – 130° E) derived from TPR. Thick line and shading denote the mean and standard deviation of ensemble members, respectively. The 5-yr low-pass SEIO SST of HadISST data is plotted as a dashed line for comparison. (d)–(f) As in (a)–(c), but for SEIR.

variations forced by tropical Pacific SST (Figs. 10b,c) are in fact no weaker than locally generated variations (Figs. 11e,f), but these Pacific-originated variations are not well correlated with the SEIO SST (0.24 and -0.09 for coastal wind and precipitation, respectively) and do not contribute to the positive feedback processes. The locally generated atmospheric anomalies show high correlations with the SEIO SST (-0.88 and 0.51 for coastal wind and precipitation, respectively) but are also subjected to large internal instabilities. The signal-to-noise ratio is 0.84 for coastal wind and 0.51 for precipitation, reminding caution for interpretation and attribution of climate variability in this region.

In addition to local impact, the remote impact of the SEIO SSTAs is of interest. Zhang and Han (2018) put forward a possibility of the SEIO SSTAs affecting tropical Pacific climate on interannual time scale. In the SEIR ensemble mean, we see some signatures in the western equatorial Pacific and the tropical Indian Ocean (Fig. 10e). The SST warming in the SEIO causes easterly winds in the far western equatorial Pacific and westerly winds toward the island of Java. Figure 12 shows zonal wind variability in these two regions from the TPR and SEIR experiments. Comparing with the variability in TPR, wind variability in the SEIR is weak and shows large uncertainty of atmospheric internal instability as indicated by the standard deviation of

ensemble members. The strong internal variability also acts to obscure the distinction between the Pacific and the SEIO effects. Figures S5 and S6 suggest that, for most cases, the ensemble difference between TPR and SEIR cannot exceed the one standard deviation range of members. It is likely that on decadal time scales the SEIO SST is not able to exert robust impact on the tropical Pacific and tropical Indian Ocean through atmospheric teleconnection.

5. Summary and discussion

Prominent decadal SST variability is observed over the SEIO, which may have significant effect on regional climate and marine ecosystems. In the present study, we provide a comprehensive investigation of decadal SST variability in the SEIO and its impacts on regional climate. This is pursued by two steps. First, we explore the generation mechanisms of decadal SST variability from an oceanic point of view. OGCM experiments are performed and analyzed to isolate effects of different processes that control SST variability. Then we focus on the atmospheric response to SST forcing. AGCM experiments are used to evaluate the impact of the SEIO SSTA on regional low-level wind, SLP, and precipitation variations. Hopefully the efforts from two perspectives can improve our understanding of decadal

FIG. 12. 5-yr low-passed zonal winds of (a) the tropical Indian Ocean (15° – 5° S, 70° – 110° E) and (b) the western equatorial Pacific (2° S– 2° N, 130° – 140° E) derived from TPR. Thick line and shading denote the mean and standard deviation of ensemble members, respectively. (c),(d) As in (a),(b), but for SEIR, and 5-yr low-pass SEIO SST of HadISST data is plotted for comparison.

climate variability in the SEIO. Findings of our analysis are summarized below.

- 1) The SEIO exhibits evident decadal variations in SST with amplitudes of ~ 0.2 – 0.3 K, showing persistent warm SSTAs in the periods of 1955–65, 1982–90, 1995–2002, and 2009–14 and cold SSTAs during 1966–81, 1991–94, and 2003–08. The SEIO bears a close relationship with the central tropical Pacific. It shows a correlation of $r = -0.63$ with Niño-4 on decadal time scale.
- 2) HYCOM can realistically simulate the observed decadal SST variability since the late 1970s, and HYCOM experiments reveal that ocean dynamical processes driven by wind stress variability make minor contributions to SST variability. The Pacific wind forcing through the ITF explains $<20\%$ of the total decadal SSTA. Local wind forcing in the SEIO, through both wind stress and wind speed effects, explains only $\sim 27\%$, as quantified by the difference between WND and PAC. The wind-forced SST variability mainly occurs in the coastal region east of 108° E.
- 3) The majority of the SST variability is attributed to surface heat fluxes. The reduced upward Q_T owing to reduced wind speed, cold offshore SSTa, and suppressed advection of cold, dry air from the mid-latitudes (or equivalent to anomalous advection of warm, moist air from the Maritime Continent) is particularly important for the growth of warm SST anomalies. When large-scale warm SSTAs are generated, the positive cloud–radiation–SST feedback is triggered, which acts to further promote or at least

maintain the SST warming; meanwhile, the warming effects of SWR and wind stress-driven ocean dynamics are compensated by enhanced upward Q_T due to high SST and increased wind speed. Therefore, the feedback by surface Q_T is most important in driving and damping decadal SSTAs.

- 4) Decadal warming of the SEIO coincides with low SLP and cyclonic low-level winds in the SEIO and enhanced rainfall in the tropical SEIO including the northwest Australia and Indonesian Seas. AGCM experiments are performed to explore how these anomalies are generated. The results demonstrate that although cooling of the tropical Pacific can induce large-scale low SLP over the eastern Indian Ocean, the cyclonic winds and enhanced precipitation are primarily caused by SST warming in the SEIO. These results highlight the importance of local air–sea interaction rather than remote forcing from the Pacific in generating decadal climate variability of the SEIO.
- 5) AGCM simulations also reveal strong atmospheric internal instability as indicated by the large spread of ensemble members, suggesting that impacts of the SSTa forcing on regional climate are subject to large uncertainties.

In this study, we explore the coupling between ocean and atmosphere using stand-alone OGCM and AGCM experiments, which can more faithfully simulate the observed variability as compared to coupled models. The generation of oceanic or atmospheric anomalies can be unambiguously attributed to prescribed forcing processes. But we are also aware that this cannot fully depict

FIG. 13. Comparison of decadal SST variability and Ningaloo Niño/Niña. Regression maps of SSTA onto (a) the 5-yr low-passed SEIO SST and (b) the Ningaloo Niño index (NNI; average SST anomaly of 28° – 22° S, 108° – 116° E). Stippling indicates $>90\%$ significance. Note that different color scales are used for (a),(b).

air–sea coupling processes. In reality, variability in this region is never generated in a one-way manner. The complicated nonlinear interactions between different oceanic and atmospheric processes are key for the growth and decay of variability. For instance, the initial warming near the Western Australian coast may be generated by Pacific wind-forced downwelling waves. It then drives cyclonic winds in the SEIO and grows through local air–sea coupling processes. Coupled models are able to fully resolve these complicated interactions but subject to evident biases in mean state and variability (e.g., Kido et al. 2016). In addition to the impact from the tropical Pacific, our results also suggest a possible association with the tropical Indian Ocean, particularly the IOD mode. Decadal modulation of the IOD (Ashok et al. 2004; Tozuka et al. 2007) can modulate the SEIO, with positive IOD-like condition favoring the SEIO warming.

In our HYCOM configuration, Q_T is found to be the dominant driver of decadal SSTA. But Q_T is the result of complicated air–sea interaction rather than atmospheric forcing only. Here we did not perform experiments to assess the effects of different factors (air temperature, humidity, and nonlinear effect of different factors) on Q_T and therefore are not able to provide quantitative insights. Different from AGCM experiments, there is only one member for each HYCOM experiment, and therefore the sensitivity to initial condition is not evaluated. Ocean internal processes such as mesoscale eddies in the SEIO (Feng et al. 2005; Jia et al. 2011) may rectify onto decadal oceanic variability. Previous studies (e.g., Trenary and Han 2013; Li and Han 2015; Sérazin et al. 2016; Li et al. 2018) have demonstrated that this effect is large for sea level variability in the subtropical south Indian Ocean. However,

large-scale, low-frequency SST variability in the SEIO is mainly caused by deterministic forced processes (winds and heat fluxes) as shown in this study and many others (e.g., Feng et al. 2013; Kataoka et al. 2014; Marshall et al. 2015; Zhang et al. 2018). In such case, the influence of initial condition on the results is supposed to be small. Nevertheless, it is of interest to evaluate this sensitivity by performing ensemble simulations with eddy-resolving configurations. Investigation is in progress to address the above two issues.

Given that the Ningaloo Niño/Niña has been a well-known climate phenomenon, it is instructive to clarify the differences between decadal SST variability and Ningaloo Niño/Niña. First, amplitude of the Ningaloo Niño/Niña is at least 2 times stronger than that of decadal SSTA (Fig. 13). Second, although spatial structures of the two are broadly the same, the maximal decadal SSTAs are located along the 108° E meridian away from the coast, while those of Ningaloo Niño/Niña are concentrated near the Western Australian coast. Third, the Ningaloo Niño/Niña is phase locked to seasonal cycle with the largest variability occurring in austral summer, whereas decadal SST variability does not show evident seasonality. For the mechanisms, we have not yet found dramatic differences between the two variability phenomena. Processes modulating Ningaloo Niño/Niña, such as Pacific wind forcing, Q_T , and SWR, are also essential for decadal variability. Our analysis only covers 3–4 cycles of decadal variability, and after the low-pass filtering the effective degree of freedom has been rather low. The significance test for the regression coefficients here does not use effective degree of freedom, unlike the case of correlation coefficients. Otherwise, only a very small number of coefficients will be significant. Analyses of longer data records are required

to consolidate our results. Also worthy of further investigation is the interaction between Ningaloo Niño/Niña and decadal SEIO variability. Strong Ningaloo Niño/Niña events may be essential for the phase shifts of decadal SSTAs. Anomalies of Ningaloo Niño/Niña may persist or reemerge for several years through local air–sea feedbacks, rectifying onto decadal variability. On the other hand, decadal anomalies modulate the background upon which Ningaloo Niño/Niña events develop and thereby affect their amplitude and evolution.

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