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The Tropical Indian Ocean decadal sea level response to the Pacific Decadal Oscillation forcing

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Abstract

The tropical Indian Ocean sea level displayed decadal variations in response to Pacific Decadal Oscillation (PDO). Contrasting patterns of decadal oscillation in sea level is found during the opposite phases of PDO especially in the thermocline ridge region of the Indian Ocean (TRIO; 50°E–80°E; 15°S–5°S). Epochal mean sea level rise is observed over the TRIO region during the cold phase of PDO (1958–1977), whereas epochal mean sea level fall is observed during the warm phase of PDO (1978–2002). Analysis reveals that the decadal variability in the sea level pattern in the TRIO region is in accordance with the PDO phase shifts and is primarily caused by changes in the surface forcing over the Indian Ocean as a response to PDO. The changes in the large scale Walker circulation over the tropical Indian Ocean region during the different phases of PDO support our hypothesis. The winds and wind stress curl variations associated with these large scale circulation changes are primarily inducing the observed regional decadal sea level variability over TRIO. The decadal forcing through Indonesian Through Flow (ITF; oceanic channel) however did not show any significant impact on the TRIO sea level variability. Further Ocean General Circulation Model (OGCM) sensitivity experiments are carried out to understand the mechanisms and the possible contribution of the Pacific Ocean through oceanic pathways, in the decadal variability of the TRIO sea level. It is noted that wave propagation from the Pacific Ocean to the Indian Ocean through ITF region has contributed to the sea level variations in the eastern Indian Ocean. But on the decadal time scale, tropical Indian Ocean/TRIO sea level is unaffected by decadal variability in the ITF. Moreover, the ITF contribution to the decadal sea level variability in the Indian Ocean is found to be significant only in the region south of 20°S.

Keywords Sea level · Indian Ocean · Pacific Decadal Oscillation · Decadal variability

1 Introduction

Thermal expansion and continental ice melting are known to be the primary factors for the global mean sea level rise. Global mean sea level has been rising at a rate of 3.2 mm/year during the last several decades (e.g., Bindoff et al. 2007; Church et al. 2013). The Indian Ocean sea level rise in the last two decades of altimeter observations is significantly at a higher rate than before and is close to the global trends (Unnikrishnan et al. 2015). Low lying islands and highly populated coastal zones of Indian Ocean rim countries further increase the societal impacts associated with the Indian Ocean sea level rise. The decadal variability in the sea level over the Indian Ocean and its potential impacts have been reported by previous studies (e.g., Han et al. 2014; Lee and McPhaden 2008; Nidheesh et al. 2013). Understanding this variability is therefore essential especially in the context of improving the skill of decadal sea level prediction (e.g., Li and Han 2015; Han et al. 2014).

The studies in the past have primarily attributed this variability to the changes in the wind forcing over the Indo-Pacific region (e.g., Han et al. 2017; Lee and McPhaden 2008; Timmermann et al. 2010; Nidheesh et al. 2013). Using satellite altimeter data, Lee and McPhaden (2008) showed that the decadal reversals in the sea level trends over the

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Indo-Pacific region during 1993–2000 and 2000–2006 are associated with the decadal variations of the Indo-Pacific trade winds. Li and Han (2015) on the other hand speculated the possibility of heat fluxes and surface wind stress contributing to the decadal sea level variability. Llovel and Lee (2015) showed the contribution of halosteric component to the sea level changes over the southeast Indian Ocean region. Shankar and Shetye (1999) reported the decadal variability of sea level along the Indian coasts based on tide gauge observations. They identified a link between the variability of monsoon and sea level, arising from salinity changes in the coastal water. The decadal sea level variability in the north Indian Ocean has been addressed by many recent studies (e.g., Srinivasu et al. 2017; Swapna et al. 2017). According to Srinivasu et al. (2017), the decadal sea level reversal in the north Indian Ocean during the last two decades is a combined effect of meridional heat transport and surface turbulent heat flux, both being driven by decadal variations in the surface winds. Swapna et al. (2017) and Gera et al. (2016) linked the rise in the north Indian Ocean sea level to the weakening of summer monsoon circulation. According to them the reduced southward ocean heat transport resulted in the increased heat storage and thermosteric sea level rise in the north Indian Ocean.

The southwestern tropical Indian Ocean shows large sea level variability in the decadal and interannual time scales (Fig. 1a). This region, known as the Thermocline Ridge region of the Indian Ocean (TRIO; Jayakumar et al. 2011) is a region of strong air–sea coupling characterized by high sea surface temperature (SST) variations, shallow thermocline and mixed layer (e.g., Xie et al. 2002; Chowdary et al. 2009; Vialard et al. 2009). Han et al. (2010) attributed the sea level trends over this region to the wind stress curl resulting from the combined effect of enhanced Hadley and Walker cell over this region. Nidheesh et al. (2013) suggested that the wind stress curl in the southern Indian Ocean drives decadal variability in the southwestern Indian Ocean through planetary waves. Based on Ocean General Circulation Model (OGCM)

experiments, Trenary and Han (2013) attributed decadal sea level variations in TRIO to be wind stress curl driven and through westward propagating Rossby waves generated by winds in the central and eastern Indian Ocean basin. Li and Han (2015) also found that the decadal sea level variations in the southern tropical Indian Ocean are induced by decadal surface wind stress fluctuations.

Pacific Decadal Oscillation (PDO) is a dominant decadal climate mode in the north Pacific, with large impacts on global climate. The positive (negative) phase of PDO is characterized by negative (positive) SST anomalies in the western and central north Pacific surrounded by positive (negative) SST anomalies along the north American coast. Zhou et al. (2017) analysed the subsurface temperature trends in the southern Indian Ocean during 1960–2000 and showed that the trends are closely related to the phases of PDO. There are number of studies in the past examining the relation between PDO and the Pacific Ocean sea level variability (e.g., Bromirski et al. 2011; Moon et al. 2013; Zhang and Church 2012). However, the influence of PDO on the Indian Ocean sea level changes is not explored in the past. In this paper, we examine the Indian Ocean sea level response to PDO and understand the driving mechanisms.

The PDO can influence the Indian Ocean sea level through oceanic pathways as well as via atmospheric bridge. Ummenhofer et al. (2017) related the multidecadal variations in the Indian Ocean subsurface heat content with PDO and showed that transmission of the multidecadal signal from Pacific to the Indian Ocean occurs through the Indonesian Through Flow (ITF). These signals propagate westward and modulate the subsurface heat content variations in the Indian Ocean. In this context, whether ITF modulation will affect TRIO region in decadal time scales is a potential research problem. Deepa et al. (2018) reported that ITF induced sea level variability in the decadal time scales in the Indian Ocean is restricted only to the region south of 20°S. Trenary and Han (2008) suggested that the ITF influence on the Indian Ocean occurs only below 200 m depth. This further supports the need of understanding the mechanisms

Fig. 1 **a** Standard deviation of sea level anomalies (cm) in the decadal time scales (11 year running mean, shaded) and interannual time scales (contours). The box defines the region of maximum standard deviation (50°E–80°E, 15°S–5°S). **b** Climatological D20 (m, shaded) and sea level (cm, contours). Sea level and temperature data are from ORAS4 for the period 1958–2002
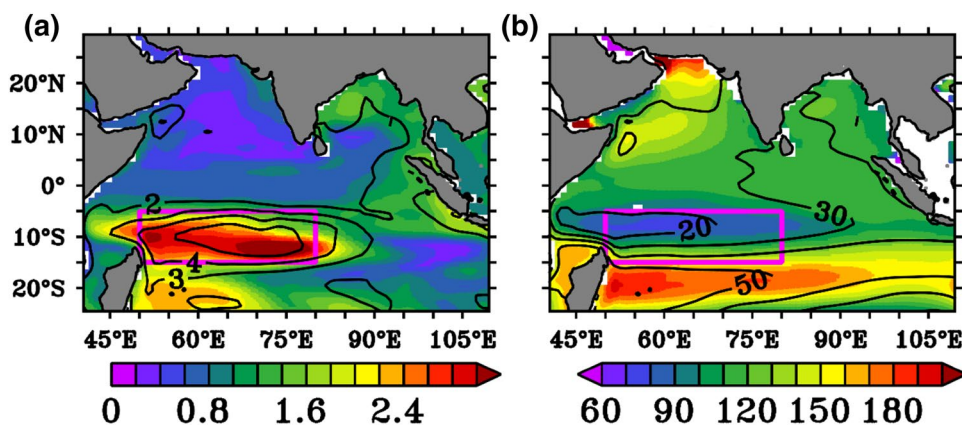
responsible for the decadal variability in the Indian Ocean sea level.

The mean ITF is maintained and driven by the large scale ocean pressure gradient between the Pacific and the Indian Ocean basins (Wyrtki 1987). Previous studies show that the ITF variability is largely modulated by local and large scale Pacific winds (e.g., Feng et al. 2018; Liu et al. 2015; Wainwright et al. 2008). It is important to note that the OGCM experiments are necessary to understand the physical mechanisms that link PDO and sea level variations over the TRIO region. In particular, the sensitivity of TRIO sea level variations related to PDO can be identified only through carefully designed OGCM sensitivity experiments. However, such sensitivity studies are not possible with only observations. None of the previous studies have used OGCM based sensitivity experiments to examine the impact of decadal variations of PDO on the tropical Indian Ocean sea level. It is important to note that OGCM experiments are essential to understand the mechanisms behind the observed variability and to quantify the relative contributions from different components. In the present study, the influence of PDO on TRIO sea level through atmospheric bridge and oceanic pathways is carefully examined using OGCM experiments. The variability in the surface forcing over the Pacific Ocean is modulated by carefully designed OGCM experiments so that the transmission of signals from the Pacific Ocean to the Indian Ocean through ITF is suppressed and quantified.

The rest of the paper is organized as follows. In Sect. 2, model details, experimental setup, data sets used and methodology adopted in the study are provided. Section 3 describes the PDO induced decadal sea level variability pattern in the Indian Ocean. Section 4 discusses the role of local versus remote forcing from the Pacific Ocean in driving the TRIO sea level pattern. Section 5 summarises the major findings of the present study.

2 Model, data and methods

2.1 Model and experiments

The OGCM Modular Ocean Model version 5 (MOM5), which is a tool for understanding the ocean climate system (Griffies et al. 2004) is used in this study for process studies. The model is configured for the global domain and has a uniform horizontal resolution of 1°. Vertical coordinate system is pressure based and there are 50 vertical levels with 10 m resolution in the top 250 m. MOM5 is forced by the Coordinated Ocean-ice Reference Experiments (CORE) surface forcing datasets for global ocean-ice modelling (Large and Yeager 2008) except the winds. The winds are from European Centre for Medium Range Weather Forecasts (ECMWF) Reanalysis (ERA) ERA40 (Uppala et al.

2005) and ERA-Interim (Simmons et al. 2007). The ERA 40 winds for the period 1958–1978 and ERA-Interim winds from 1979 onwards are combined (merged) to generate wind forcing for the entire study period. The model is spun up from a state of rest with CORE normal year forcing (NYF) for 50 years to attain the quasi steady state. The model has been further integrated by interannually varying CORE surface fluxes (heat and freshwater) and ERA winds during 1958–2009. The model outputs with interanually varying forcing fields are referred to as CTL. To study the contribution of signal transmission through oceanic pathways from Pacific, the surface forcing fields over Pacific is modulated in a separate experiment, in which climatological forcing fields are applied over the Pacific Ocean (NYF-PO). Therefore, NYF-PO experiment differs from CTL by the surface forcing over the Pacific, where the interannual forcing fields are replaced by the climatological forcing fields. The difference between CTL and NYF-PO (CTL minus NYF-PO) over the Indian Ocean region quantifies the variabilities that are of Pacific origin and transmitted to the Indian Ocean through ITF. In order to make sure that there are no artificial effects created due to abrupt changes of interannual to climatological forcing at the boundary between Indo-Pacific basin, smoothing is done over grid region of 10° on both sides of the boundary lines. It is important to note that most of the grid points on the boundary line lie on land areas. The role of tropical Pacific on the observed decadal variability is further examined by removing the El Niño and Southern Oscillation (ENSO) effects from the wind fields. With the purpose of suppressing ENSO impacts, an additional model sensitivity experiment (NO-ENSO) is also carried out by changing the zonal and meridional wind forcing fields. The monthly winds are first regressed with Niño 3.4 index and these ENSO regressed winds are subtracted from the monthly mean winds, so as to suppress the ENSO impacts from the wind forcing. The model is then forced with these ENSO suppressed winds. Therefore, NO-ENSO experiment suppresses ENSO impacts on the evolution of various fields in the model.

2.2 Data sets used and methodology

The Ocean Reanalysis System 4 (ORAS4) (Balmaseda et al. 2013) sea level is utilized to carry out the analysis. ORAS4 employs the variational ocean data assimilation system NEMOVAR using Nucleus for European Modelling of the Ocean (NEMO) version 3.0 ocean model. Temperature and salinity profiles as well as along-track altimeter sea level anomalies (SLA) are assimilated in NEMOVAR. ORAS4 is forced by daily surface fluxes of heat, momentum and fresh water from ECMWF atmospheric reanalysis. Balmaseda et al. (2015) reported higher mean sea level correlation and lower RMS error in ORAS4 with reference to Global Sea

Level Observing System (GLOSS) than many other widely used reanalysis products. ORAS4 has been found to be the best reanalysis product over the Indian Ocean compared to other available products (Karmakar et al. 2017). Satellite altimeter data from combined observations of TOPEX/ Poseidon, Jason-1 and Jason-2/OSTM with Glacial Isostatic Adjustment and Inverse Barometer effect corrections incorporated are compiled and gridded by Commonwealth Scientific and Industrial Research Organization (CSIRO) (http:// www.cmar.csiro.au/sealevel/sl_data_cmar.html). This sea level data from 1993 onwards is also used in this study to compare the ORAS4 sea level. ORAS4 monthly temperature data is used to compute thermocline depth (D20). The surface as well as pressure level winds used in this study are from ERA.

The PDO index is defined as in Mantua et al. (1997) as the principal component of the leading empirical orthogonal function of north Pacific (20°N–70°N) SST anomaly (after removing the long term trends). Monthly SST observations from Extended Reconstructed SST (ERSST) version 4b (Huang et al. 2015) is used to compute PDO index. The PDO index values from Joint Institute for the Study of the Atmosphere and Ocean (JISAO) (http://research.jisao.washington.edu/pdo/PDO.latest.txt) are used for intercomparison. Niño 3.4 index from ERSST is used for regression analysis. Partial correlation analysis is carried out as in Ashok et al. (2007) to suppress the influence of ENSO on SLA pattern over the Indian Ocean. The two longest contrasting phases (or epochs) of PDO (1958–1977 and 1978–2002) in the well observed period (1950 onwards) are analysed in this study. In addition to the above, the recent short cold phase of PDO (2003–2012) is compared with the positive phase period of 1993–2002 using the altimeter SLA data.

3 PDO induced decadal sea level variability in the tropical Indian Ocean

The standard deviation of sea level over the Indian Ocean in the decadal time scales as well as in the interannual time scales during 1958–2002 is shown in Fig. 1a. Maximum variability with amplitudes reaching up to 4 cm is seen

over the TRIO region. This region of decadal variability in the Indian Ocean is consistent with the previous studies as well (e.g., Nidheesh et al. 2013). The mean thermocline is very shallow in this region (Fig. 1b) and hence it is a region of strong air–sea coupling (e.g., Xie et al. 2002; Chowdary et al. 2009; Jayakumar et al. 2011). Therefore, it is important to understand the decadal sea level variability over this region and the associated driving mechanisms. It is important to note that the magnitude of sea level variations in the TRIO region is similar on both decadal and interannual time scales. The SLA averaged over the region of peak decadal variability from 50°E to 80°E and 15°S to 5°S (region enclosed by box in Fig. 1) is normalised and is shown along with decadal SLA and PDO index in Fig. 2. It is noted that there are dominant sea level highs during 1972 and 1997, which are caused by interannual events such as Indian Ocean Dipole (IOD) and El Niño (e.g., Deepa et al. 2018; Sayantani and Gnanaseelan 2015). It is important to note that the SLA computation for ORAS4 is based on the sea level climatology for the period 1958–2017 and that for altimeter is based on 1993–2017 climatology. It is evident that ORAS4 SLA and altimeter SLA are consistent throughout, with only slight deviations in the SLA peaks (Fig. 2). These differences in SLA peaks might have resulted from internal biases in ORAS4, arising mostly due to the different periods considered for the climatology and the assimilation strategy adopted in ORAS4. It is interesting to see that the TRIO SLA in the decadal time scales and PDO index show an anti-phase evolution throughout with concurrent phase change in late 1970s and early 2000s, indicating a possible link between TRIO sea level and PDO. The contrasting differences in the SLA during the different phases of PDO are observed in the recent period as well, which is discussed in detail in Sect. 5. It is important to understand whether the pattern of sea level variability in decadal time scales has any association with PDO and whether PDO is the main driver for the observed decadal variability in the southern tropical Indian Ocean and TRIO in particular.

The PDO pattern, which is the leading mode in the EOF of detrended SST anomalies over northern Pacific, from both ERSST and model (CTL) is shown in Fig. 3a. This mode accounts for respectively 22% and 20% of the

Fig. 2 Sea level anomalies (cm) averaged over the box in Fig. 1. Decadal sea level anomalies (black) and JISAO PDO index (red) are shown in overlaid thick lines. All the time series are normalised by the respective standard deviation
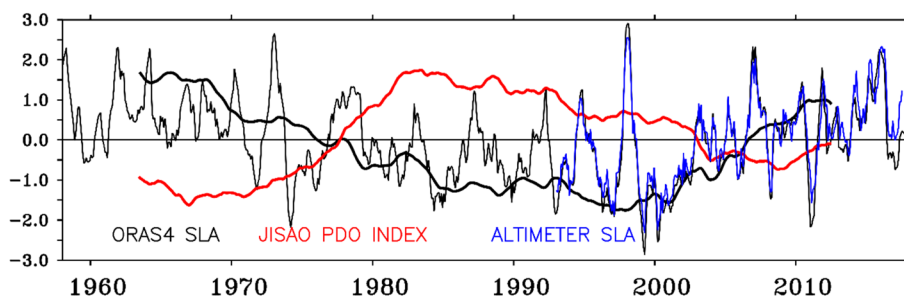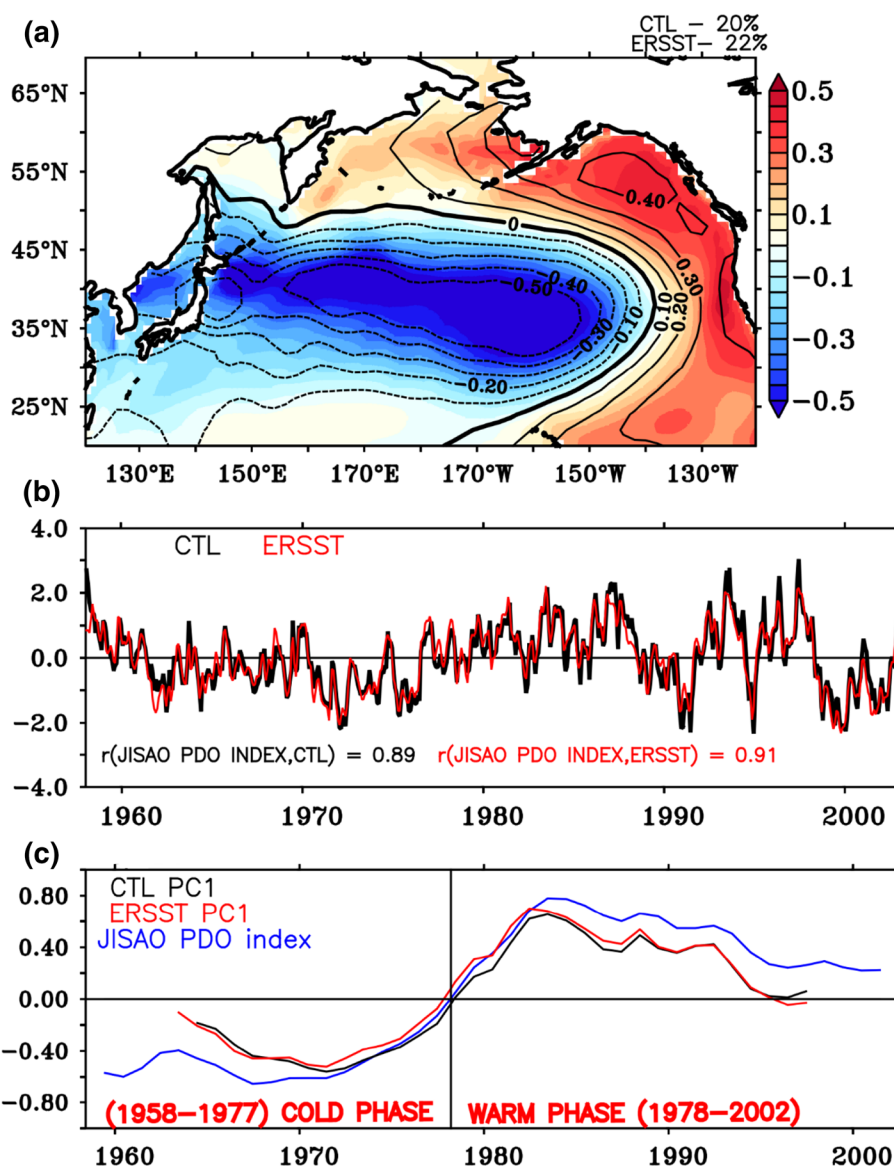
Fig. 3 a EOF-1 of detrended SSTA over North Pacific from ERSST (shaded) and CTL (contours), **b** PC-1 from ERSST (red) and CTL (black), **c** PDO index from ERSST, model and JISAO; 11 year running mean of ERSST PC-1 (red), CTL PC-1 (black) and JISAO PDO index (blue)

north Pacific SST variance in ERSST and the model. The PDO pattern is well represented by the model (CTL). The PDO index defined as the PC-1 of this leading mode as in Mantua et al. (1997) is also similar across ERSST and model (Fig. 3b). The PDO index from reanalysis (and model) is highly correlated (correlation value 0.91 and 0.89 respectively for ERSST and model) with JISAO PDO index, an observation based index. This highlights the reliability of the model in simulating PDO. A 11 year running mean is applied on the PDO index in order to suppress the variabilities those are smaller than decadal time scales (Fig. 3c). The resulting time series is used to identify the cold (1958–1977) and warm (1978–2002) episodes (or epochs) of PDO. The PDO phase change from the negative to positive as shown in the observation and reanalysis is captured by the model as well. Both the

positive and negative phases are longer than 20 years. The longer time scale ocean–atmosphere climate variability can apparently cause significant impacts on global climate. In order to study the impact of PDO on the Indian Ocean sea level in detail, the Indian Ocean SLA composites during the cold and warm phases of PDO from reanalysis and CTL are displayed in Fig. 4. It is important to note that the Indian Ocean sea level patterns have undergone extreme phase change as PDO shifts from cold to warm phase. Close examination reveals that negative (positive) phase of PDO forces positive (negative) SLA in the south tropical Indian Ocean. Both the above positive and negative anomalies are widespread and covering the entire basin width. However, the maximum amplitudes in SLA are closely located in the TRIO region (Fig. 1b). In fact, the PDO can influence the TRIO either through atmospheric bridge or

Fig. 4 Sea level anomaly composite (cm) from ORAS4 (shaded) and CTL significant at 95% confidence level based on two-tailed Student's t-test (contours) during **a** cold phase, **b** warm phase
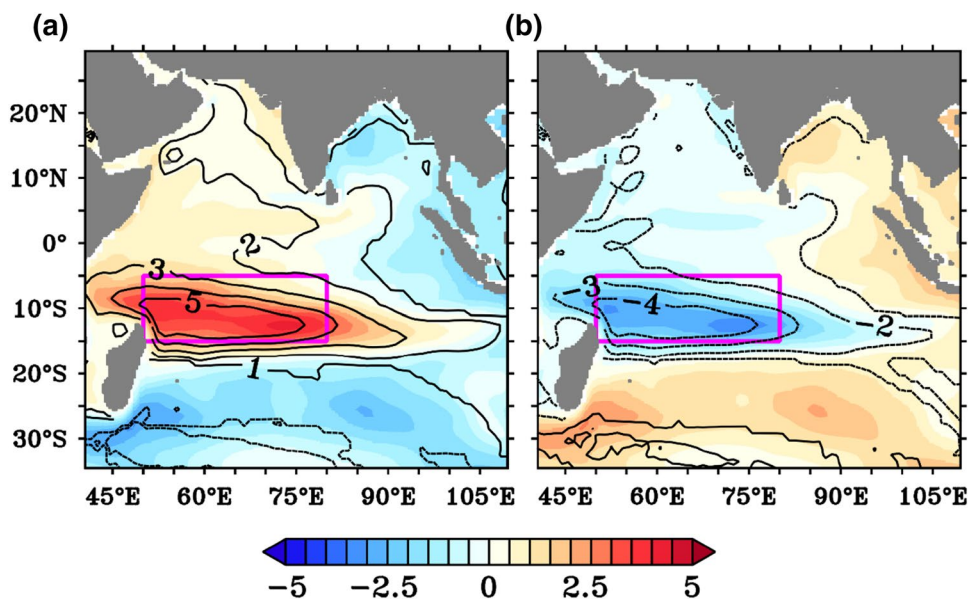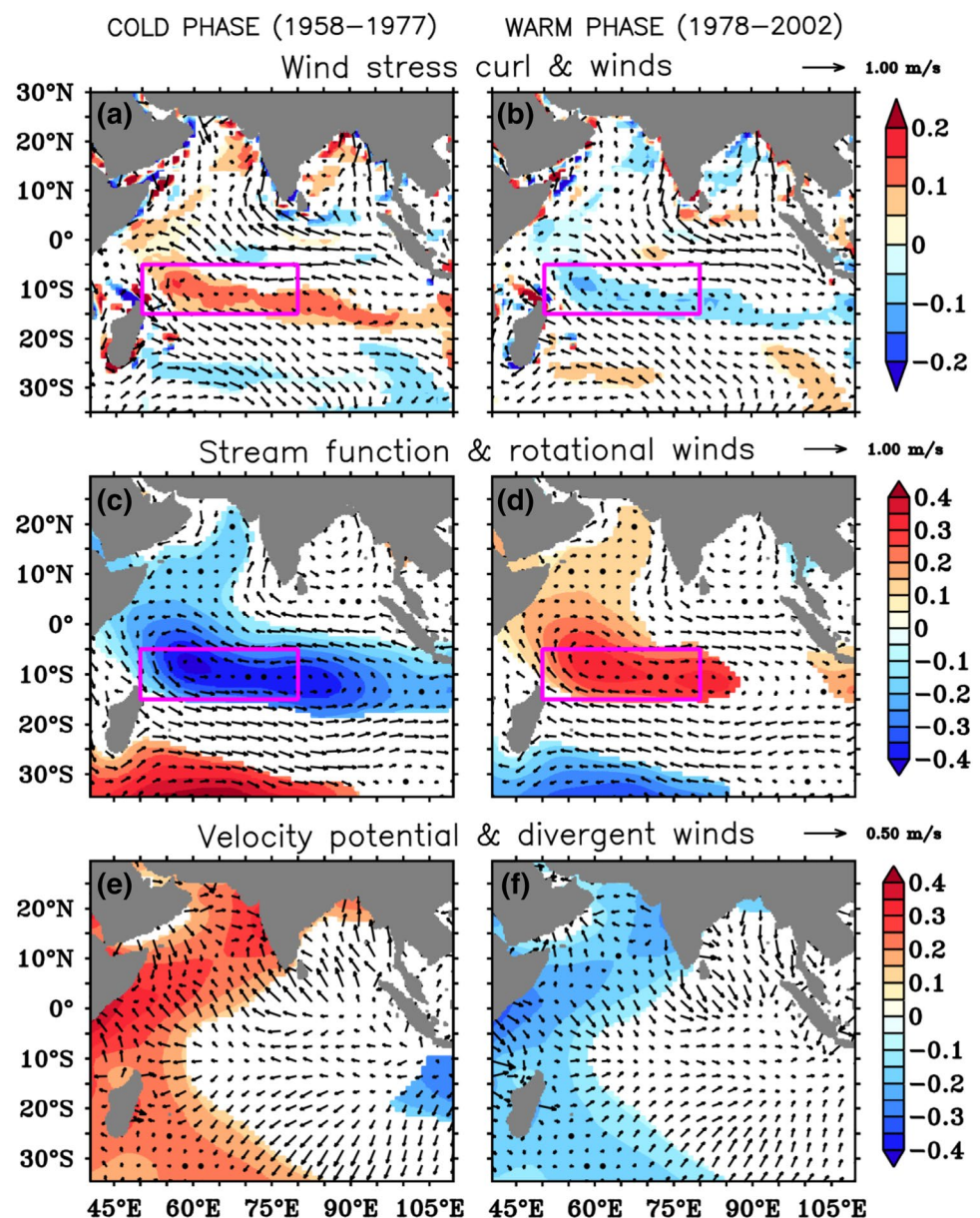

through oceanic pathways. This has been carefully examined in this paper by OGCM sensitivity experiments.

A major part of the sea level variability in the Indo-Pacific region is contributed by wind stress induced thermocline changes (e.g., Lee and McPhaden 2008; Gnanaseelan et al. 2008; Timmermann et al. 2010; Gnanaseelan and Vaid 2010). Composites of surface wind anomalies and wind stress curl anomalies in the two phases are shown in Fig. 5a, b. The composites are prepared as the annual mean of the respective time periods of PDO. It is interesting to note that the negative (positive) PDO phase is characterized by easterly (westerly) anomalies over the equatorial Indian Ocean. These anomalous winds induce anticyclonic (cyclonic) circulation over the TRIO region during the cold (warm) phase of PDO. The anomalous winds as well as the meridional wind shear in the horizontal winds are responsible for positive (negative) wind stress curl over 12°S–8°S region in the cold (warm) phase (Fig. 5a, b). The positive (negative) wind stress curl during cold (warm) phase of PDO is strong and widespread covering the longitudinal band from 50°E to 100°E. The positive (negative) curl during the cold (warm) phase induces downward (upward) Ekman pumping and deepening (shallowing) of the thermocline, which propagates westward as downwelling (upwelling) Rossby waves and amplify (weaken) the sea level in the TRIO region. The integration of Rossby waves generated along the latitudinal belt from 12°S to 8°S supports the westward accumulation of sea level anomalies. It is important to note that the coupled air–sea interaction over this region supports the sustenance of these signals in the decadal time scale in addition to the decadal forcing from PDO. We further decomposed the winds into the rotational and divergent components to explore the forcing mechanism. Anomalous easterlies in the

equatorial region are seen in the cold phase of PDO, whereas westerlies are observed during the warm phase (Fig. 5c, d). Anticyclonic circulation is evident in the rotational wind component during the cold phase around 12°S–8°S region in the TRIO region (Fig. 5c). Instead, a cyclonic circulation is observed in the same region during the warm phase (Fig. 5d) thereby imposing a symmetric response to the opposite phases of PDO. Convergence of easterlies to the TRIO during cold phase and divergence of westerlies out of TRIO during warm phase of PDO is also noted (Fig. 5e, f). The wind structure observed during the two phases largely explains the sea level variations in the TRIO, through generation of Rossby waves.

The evolution of downwelling and upwelling Rossby waves in the TRIO region in the negative and positive phases of PDO is displayed as time longitude plot of SLA and wind stress curl anomalies (Fig. 6). It is very clear that the sustained wind stress curl is forcing the SLA over the southern tropical Indian Ocean region. During the negative phase of PDO, persistent generation and propagation of downwelling Rossby waves are seen, especially during 1958–1973 and persistent upwelling Rossby waves are seen from 1979 onwards. The transition periods, 1974–1976 experiences upwelling waves and the period 1977–1978 is experiencing downwelling wave. It is also noticed that there are weaker upwelling Rossby waves in the negative phase and weaker downwelling Rossby waves in the positive phase, without making much impact on the decadal timescale. The wind stress curl also exhibits decadal variability during the period (Fig. 7), with positive decadal wind stress curl forcing downwelling waves in the negative phase and negative decadal wind stress curl forcing upwelling waves after 1978. The decadal variability in wind stress curl is therefore

4 Role of atmospheric and oceanic pathways in TRIO decadal sea level variability

To understand the origin of equatorial easterlies seen in the cold phase of PDO and westerlies in the warm phase as well as the contrasting rotational and divergent winds, we have examined the Walker cell structure (Fig. 8). The differences in the anomalous Walker circulation (cold phase–warm phase) over the Indo-Pacific Ocean averaged along 5°S–5°N (Fig. 8a) and 12°S–8°S (Fig. 8b) exhibit modulation in the

responsible for the decadal sea level variations in the TRIO. It is also important to note the similar phase shift in both the sea level and wind stress curl.

large scale circulation patterns. The typical Walker cell with convection in the eastern equatorial Indian Ocean region is observed during cold phase, whereas subsidence is noted over this region during the warm phase (figure not shown). The divergence pattern during the two phases (Fig. 5e, f) also support this. Moreover, the existence of a meridional cell with ascending branch over the equatorial Indian Ocean region and descending branch over the southern Indian Ocean region is also noted (figure not shown). It appears that PDO influences the surface winds through atmospheric teleconnection via modulating the Walker cell/large-scale atmosphere circulation adjustments. The large scale atmospheric circulation changes associated with PDO has strong influence on the Indian Ocean Walker cell as well. These changes forced the surface wind anomalies and the

Fig. 6 Time evolution of ORAS4 SLA (cm, shaded) and curl of wind stress anomalies (N/m³, contours) averaged over 15°S–5°S region during 1958–2002
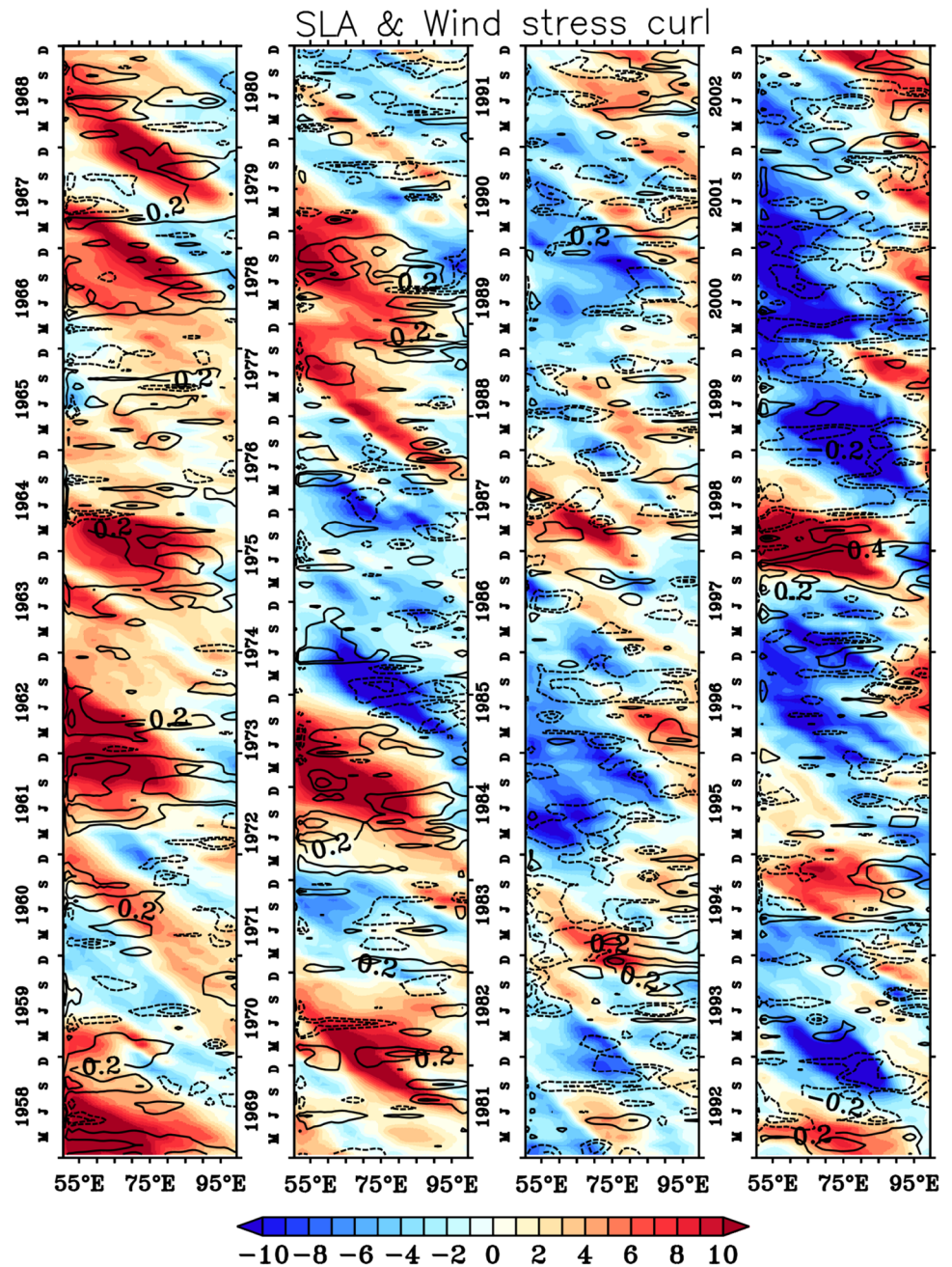
associated wind stress curl, which in turn are responsible for the observed contrasting sea level pattern during the negative and positive phases of PDO. Further the anomalous Walker circulation in both the phases induced contrasting circulation patterns in the upper troposphere as well (at the heights of 100 hPa; figure not shown).

To understand the role of oceanic pathways that link PDO and sea level variations over TRIO region, we have carried out sensitivity experiments with an ocean model by controlling the forcing fields. In NYF-PO experiment, the interannual forcing over Pacific is replaced by climatological forcing fields and in the NO-ENSO experiment, ENSO

related wind components are suppressed from monthly mean winds. The SLA composites as well as 0–100 m heat content anomaly composites during the cold and warm phases of PDO from CTL and NYF-PO experiments are shown in Fig. 9. Even though the forcing fields over Pacific are controlled, NYF-PO run is able to reproduce the dominant pattern of decadal sea level variability in the TRIO during the two phases and is consistent with that shown by CTL (Fig. 9a). The sea level pattern to the south of 20°S in NYF-PO is differing from the CTL (Figs. 4, 9a) indicating that the Pacific forcing could change the sea level over that region through oceanic pathways, which is consistent with earlier
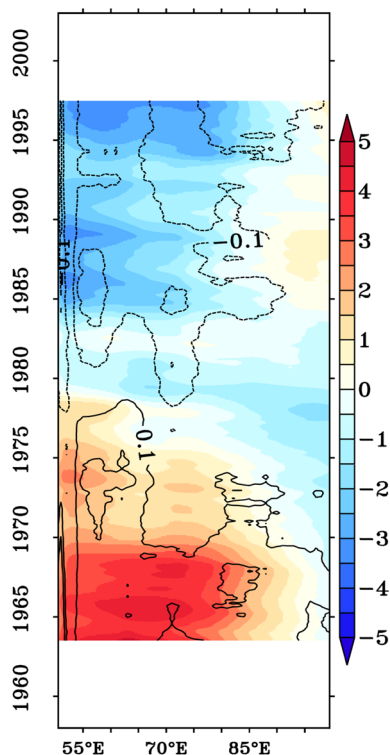
Fig. 7 Time evolution of ORAS4 decadal (11 year running mean applied) SLA (cm, shaded) and curl of wind stress anomalies (N/m³, contours) averaged over 15°S–5°S region during 1958–2002

studies such as Deepa et al. (2018). It is important to note that the upper ocean heat content shows diverging picture (Fig. 9). For example, the upper ocean heat content anomalies in NYF-PO is found to be weaker than that of CTL but
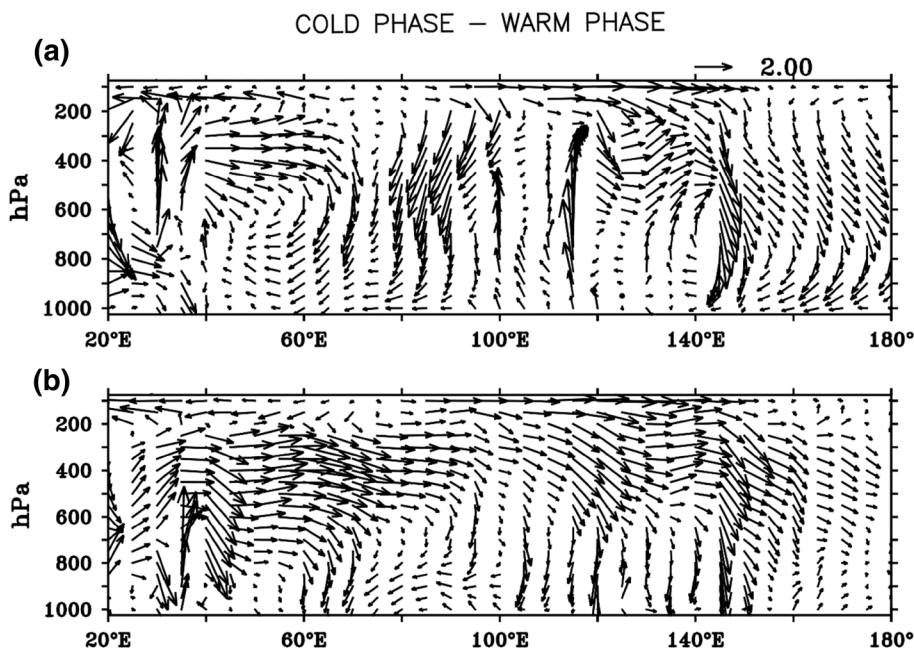
the SLA in both NYF-PO and CTL are similar. This strongly suggests the role of local wind stress and wind stress curl forcing on the decadal SLA variability in the TRIO region.

The time longitude plot of SLA in the 12°S–8°S region from ORAS4, CTL, NYF-PO and NO-ENSO experiments is shown in Fig. 10. The sea level evolution in CTL is similar to ORAS4 (Fig. 10a, b). It is interesting to see the relative consistency among CTL, NYF-PO and NO-ENSO experiments over this region, indicating that the oceanic channel is not really playing any significant role in controlling the sea level pattern in the TRIO on decadal time scale. In the interannual time scales however, there is radiation of Rossby waves from the Pacific Ocean to the Indian Ocean along 12°S–8°S region (Fig. 10a, b) and is consistent with Vaid et al. (2007). The persistent downwelling Rossby waves prior to the transition period of cold phase of PDO are noted, and thereafter persistent upwelling waves are seen (Fig. 10a, b). However, in the decadal time scales, such radiation of Rossby waves from the Pacific Ocean is not seen in both ORAS4 and model. The decadal signals are found originating west of 100°E. This strongly suggests that these signals are due to local atmospheric changes as a response to PDO forcing. The remote forcing from the Pacific Ocean through oceanic pathways is not significant.

The correlation (and partial correlation after excluding the ENSO effects) of PDO index with wind anomalies over the tropical Indian Ocean is shown in Fig. 11a, b. It is important to note that the partial correlation shows higher values over the tropical Indian Ocean than that of correlation, indicating that after removing ENSO impacts, the PDO induced winds strengthened. It is important to note that the SLA patterns over TRIO region got amplified when the ENSO

Fig. 8 Difference in the anomalous Walker cell composite during cold phase and warm phase of PDO averaged for **a** 5°S–5°N, **b** 12°S–8°S. The ERA winds, zonal (m/s) and vertical (-1×10^{-2} Pa/s) are used to produce the Walker circulation

Fig. 9 Composite of SLA (cm, contours) and 0–100 m heat content anomalies ($\times 10^9$ J/m^2, shaded) from **a** CTL and **b** NYF-PO experiments during cold phase (first column) and warm phase (second column). SLA composites are significant at 95% confidence level and heat content anomalies are significant at 90% confidence level based on two-tailed Student's t-test
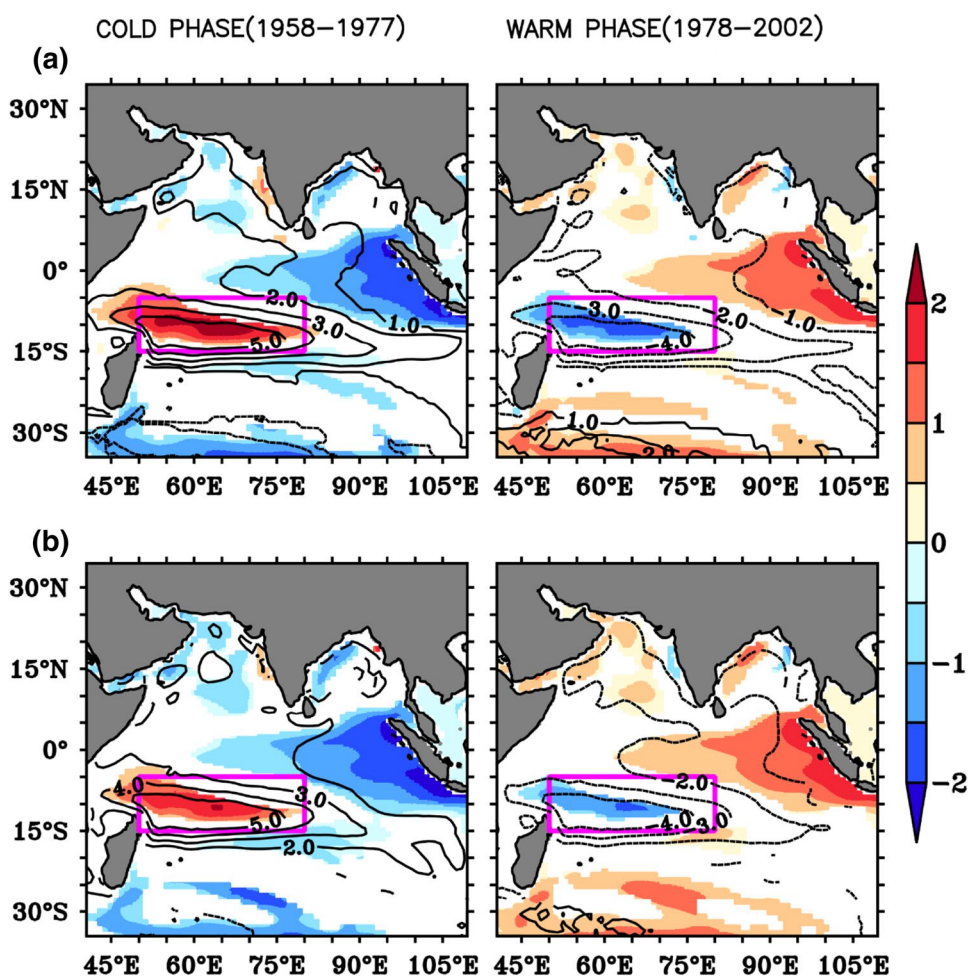
effects from winds are suppressed (Fig. 11c, d). This suggests that the decadal SLA patterns in the Indian Ocean are most likely induced by PDO. However, the PDO variability may be influenced by tropical Pacific ENSO forcing as reported by previous studies (e.g., Schneider and Cornuelle 2005; Newman et al. 2003).

5 Summary and discussion

The Indian Ocean decadal sea level variability is a research area of high societal importance. The TRIO region in particular shows large amplitude in the decadal sea level variability and hence the present study aims to understand the underlying mechanisms, using ORAS4 and ocean model sensitivity experiments. The impact of the large scale decadal climate modes, such as PDO, on the Indian Ocean sea level variability is explored. During the cold phase of PDO or cold epoch (1958–1977), a sea level high is reported in the TRIO and during the warm phase or warm epoch (1978–2002), a sea level low is seen. The SLA pattern changes in association with PDO phase shifts are present in the recent cold

PDO phase of 2003 to 2012 as well. The altimeter SLA for the period 1993–2012 (Fig. 12) is analysed in detail to confirm the results obtained from the re-analysis and OCGM. The period 1993–2002 comes as part of the long warm phase of PDO (1978–2002) and 2003–2012 is the short cold phase following the previous long warm phase. It is clear from Fig. 12 that the patterns are similar for the altimeter period as well. The large scale Walker circulation changes associated with the different phases of PDO induce contrasting wind stress curl anomalies over the southern tropical Indian Ocean. The wind stress curl driven by surface winds generate Rossby waves, which propagate westward and control the thermocline in the TRIO region. Persistent downwelling Rossby waves in the TRIO region during the cold phase of PDO and upwelling Rossby waves during the warm phase contribute to the TRIO decadal SLA variations. There is clear decadal variability in the TRIO wind stress curl, which is persistent and widespread throughout the tropical Indian Ocean longitudes.

None of the previous studies have used OGCM sensitivity experiments to examine the impact of decadal variations such as PDO on the tropical Indian Ocean sea level.

Fig. 10 Time evolution of SLA (cm) averaged for 12°S–8°S from **a** ORAS4, **b** CTL, **c** NYF-PO and **d** NO-ENSO runs. 11 year running mean of the respective fields is shown as contours
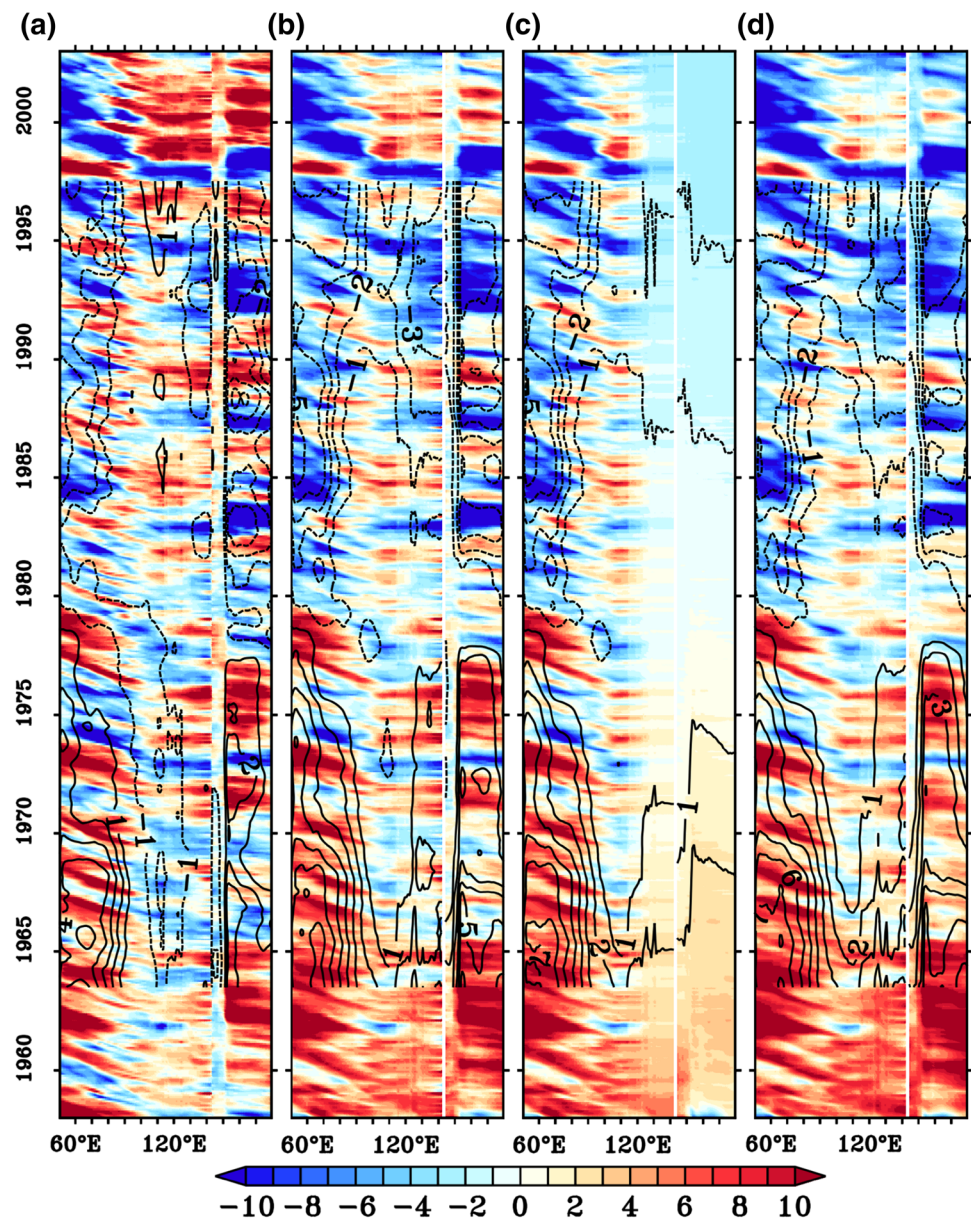
In the present study, the influence of PDO on the TRIO sea level through atmospheric bridge and oceanic pathways is carefully examined using OGCM experiments. The remote forcing from Pacific via oceanic pathways (ITF) is not significantly contributing to the decadal sea level variability over the TRIO region. However, the region south of 20°S experiences decadal energy transfers from Pacific through ITF. Further analysis reveals that decadal variability in the sea level pattern in the TRIO region is in accordance with the PDO phase shifts and is primarily caused by the winds and wind stress curl variations associated with the large scale circulation changes. We have therefore established the possible relationship between the PDO phases and the decadal sea level variability over TRIO. The present study highlights the need for taking in to account of the decadal sea level variations in the Indian Ocean especially in the context of the observed significant regional sea level trends over the Indian Ocean.

Fig. 11 a Correlation of PDO index with zonal wind anomalies (shaded) and with zonal and meridional wind anomalies (vectors). **b** Partial correlation of PDO index with zonal wind anomalies, controlling the effects of ENSO (shaded) and with zonal and meridional wind anomalies, controlling the ENSO effects (vectors). SLA composites (cm) from CTL (shaded) and NO-ENSO run (contours) during **c** cold phase, **d** warm phase. SLA composites are significant at 95% confidence level based on two-tailed Student's t-test
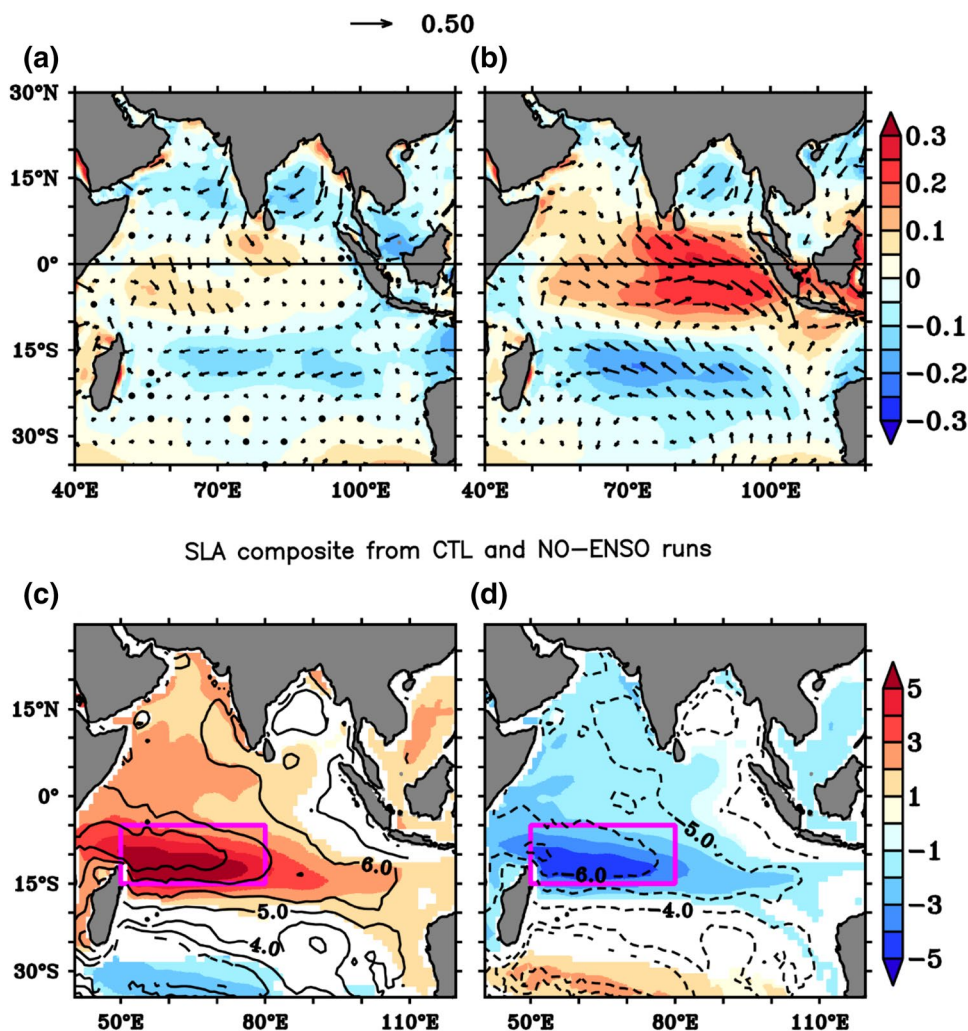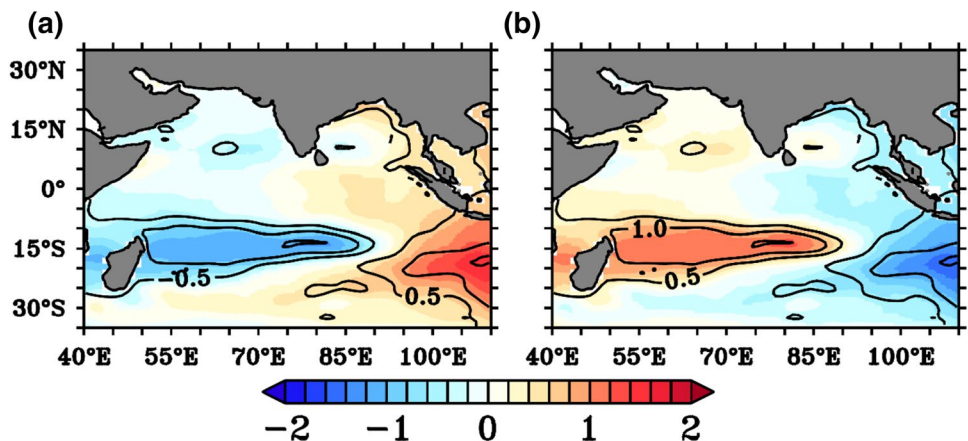

SLA composite from CTL and NO−ENSO runs

Fig. 12 SLA composite (cm) of detrended altimeter data (shaded) and ORAS4 data (contours) for the period 1993–2012. **a** Composite during the warm phase (1993–2002) and **b** composite during the cold phase (2003–2012)

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