RESEARCH LETTER
10.1029/2022GL099721

Key Points:
- A biological subsurface eastern equatorial Indian Ocean was observed during decaying period in a positive Indian Ocean Dipole (pIOD)
- The shoaling thermocline driven by reinforced easterly wind lift the nutricline into the euphotic layer in the pIOD phase
- Increased turbulent nutrient flux is caused by vertical shear due to the opposite-flowing surface current and Equatorial Undercurrent

Supporting Information:
Supporting Information may be found in the online version of this article.

Citation:
Li, H., Zhang, J., Wang, X., Zhu, Y., Liu, L., Wang, B., et al. (2022). Robust subsurface biological response during the decaying stage of an extreme Indian Ocean Dipole in 2019. Geophysical Research Letters, e2022GL099721. https://doi.org/10.1029/2022GL099721

Received 5 JUL 2022
Accepted 9 AUG 2022

Abstract  The Indian Ocean Dipole (IOD) is the predominant interannual climate mode and is critical in regulating the biogeochemical cycles of the equatorial Indian Ocean (EIO). However, the dynamics of nutrient supply and the magnitude of biological responses are less understood. Here, by comparing the biophysical in situ observations across the eastern EIO during the decaying period of a positive IOD in 2019 and neutral condition in 2017, we identify that the shoaling thermocline, initiated by reinforced easterly wind, lifts the nutricline into the euphotic layer under positive IOD condition. Coincidently, the strong turbulent mixing induced by the shear instability between the opposite-flowing surface current and subsurface Equatorial Undercurrent increases the upward turbulent nutrient flux into the euphotic zone. Their combined effect triggers a vigorous subsurface biological response over the eastern EIO, with approximately twofold higher integrated chlorophyll-a contents in the water column during a positive IOD than under neutral condition.

Plain Language Summary  The availability of nutrients in the sunlit layer is the main factor limiting microalgae growth, which is the primary producer of marine ecosystems. Particularly in the tropical oceans, strong stratification due to warm and less salty surface water inhibits nutrient replenishment from the deep layer, leading to weak seasonality in primary production. With the occurrence of climate variability, for example, the Indian Ocean Dipole (IOD), an ocean-atmosphere coupled climatic phenomenon, can significantly impact the biogeochemistry and ecosystem in the equatorial Indian Ocean (EIO). However, high-resolution in situ observations, which can reveal the mechanism of nutrient supply and biological response under IOD conditions, are still lacking. We show that the combined effect of thermocline shoaling and subsurface turbulent mixing triggered an apparently elevated nutrient supply and a subsequent robust subsurface biological response in the eastern EIO during the decaying period of an extreme positive IOD in 2019. This study may aid in understanding the evolution of marine ecosystems during the greenhouse warming era.

1. Introduction

The equatorial Indian Ocean (EIO) has unique climate features compared to other equatorial regions, including an intermittent Equatorial Undercurrent (EUC) and nonpermanent upwelling ascribed to the monsoonal transition (Chen et al., 2015; Schott & McCreary, 2001). As part of the Indo-Pacific warm pool, the eastern EIO is covered by warm (>28°C) and less salty water, and thus enhanced stratification may inhibit nutrient supply from the subsurface, resulting in the surface chlorophyll-a (Chl a) concentration being constrained to a narrow seasonality (<0.2 mg m⁻³) (Hood et al., 2017; Rixen et al., 2019; Strutton et al., 2015). However, the Indian Ocean Dipole (IOD), the main mode of interannual climate variability resulting from coupled ocean-atmosphere interactions (Saji et al., 1999; Webster et al., 1999), can strongly elevate the biological productivity in the eastern EIO in its positive phase (Hood et al., 2017; Murtugudde et al., 1999; Shi & Wang, 2021; Wiggert et al., 2009).
To date, most of the studies on biogeochemical responses are based on satellite observations or models, seldom in situ measurement to reveal mechanism of nutrient injection to the euphotic layer during IOD events.

The nutricline depth, a proxy of nutrient supply to the euphotic layer, usually co-occurs with the thermocline, and is negatively correlated with primary production in the open ocean (Baer et al., 2019; Cermeño et al., 2008). The fluctuation in the nutricline depth is largely determined by the degree of water column stratification and the magnitude of convective overturn associated with wind stress (Cermeño et al., 2008). In the tropical oceans, because of the widespread barrier layer and thus strong stratification, thermocline shoaling and upwelling are the main processes of nutrient supply to the sunlit layer (Chavez et al., 2011). Specific to the eastern EIO, the thermocline depth mainly varies with wind stress and is subjected to monsoonal transition and interannual climate change (Chen et al., 2015; Rao et al., 2010). When the easterly wind prevails, the westward surface current is developed, driving the thermocline to tilt zonally through Ekman divergence in the eastern EIO, for example, deepening in the west, but shoaling in the east (Chen et al., 2015). Especially under pIOD conditions, the shoaling of the thermocline could last 4–5 months due to Bjerknes feedback in the eastern EIO (Cai et al., 2013; Chen et al., 2015; Nagura & McPhaden, 2010; Rao et al., 2010; Zhang et al., 2014), which may trigger a significant subsurface Chl$\alpha$ maximum (SCM) in the eastern EIO. In addition, active oceanic Rossby waves could impact the thermocline/nutricline and thus the subsurface biological activity in the tropical Indian Ocean (Liao et al., 2020; Webber et al., 2014). Moreover, researchers have recently suggested that turbulent mixing is important for the vertical transport of nutrients in stratified oceans (Thorpe, 2004; Van Haren et al., 2021). When the westward surface current is forced by strong easterly wind anomalies associated with pIOD events, an eastward zonal pressure gradient, which drives a subsurface eastward-flowing EUC in the EIO, is established (Chen et al., 2015, 2016; Iskandar et al., 2009; Zhang et al., 2014). The core of the EUC is usually located near approximately 60–160 m at the thermocline beneath the westward surface current (Chen et al., 2016; Hase et al., 2008). Similar to the eastern equatorial Pacific, intense turbulent mixing could be induced by strong vertical shear between the opposite-flowing currents in the eastern EIO, which is attributed to the absence of the Coriolis force in equatorial regions (Gregg, 1987; Liu et al., 2016). As a result, vertical turbulent mixing, together with shoaling of the thermocline, are potential critical processes of nutrient input to the sunlit layer under pIOD conditions.

These processes, however, have been poorly explored in the literature. A shipboard time-series observation in the central EIO showed deepening and decreasing of the subsurface chlorophyll maximum with relaxation of the pIOD in 2006 (Kumar et al., 2012). However, in situ observations with which to reveal the biogeochemical response to the shoaling thermocline and turbulent mixing across the eastern EIO during pIOD events, are still lacking. Based on simultaneous high resolution physical-biogeochemical measurements, this study attempts to reveal the processes of nutrient entrainment and evaluate the biological response during the decaying period of the extreme pIOD event in 2019, which was the strongest of this kind since the 1960s (Du et al., 2020; Lu & Ren, 2020).

2. Materials and Methods

2.1. Sampling and Measurements

Two cruises were conducted along the equatorial section in the eastern EIO during 22–29 December 2017 and 20–29 December 2019 (Figures 1a–1b). With a rosette of Niskin bottles attached to a conductivity-temperature-depth (CTD) profiler (SBE 911 plus, Sea-Bird Inc.), discrete seawater samples were collected at 15 and 24 stations during 2019 and 2017, respectively. Thermocline depth was defined as the depth of maximum buoyancy frequency. Lowered acoustic Doppler current profiler (LADCP, RD300K, Teledyne Inc.) was deployed with the CTD profiler to measure the ocean velocity.

Water samples for nutrient and Chl$\alpha$ determinations were taken at 2, 50, 75, 100, 150, 200, and 300 m. Detailed analytical methods are introduced in Supporting Information S1 (Grasshoff et al., 1999; Welschmeyer, 1994). Simultaneously, high-resolution nitrate concentrations were obtained from a continuous profiling sensor (Deep-SUNA, Satlantic Inc.) during the 2019 cruise. The nitrate profiles from the sensor were calibrated by the data from discrete seawater measurements (Figure S1 in Supporting Information S1). Nutricline depth was defined as the depth where the nitrate concentration equals to 5 $\mu$mol L$^{-1}$.
2.2. Estimation of Turbulence and Nitrate Vertical Flux

The Thorpe method is used to calculate the dissipation rate of turbulent kinetic energy ($\varepsilon$) and the diapycnal diffusivity ($K_{\rho}$), which identifies turbulent overturns with potential density inversion (Thorpe, 1977). The Thorpe scale ($L_T$) is the root mean square of the Thorpe displacement ($d$) and is expressed as:

$$L_T = \sqrt{\sum d^2}$$

(1)

where the Thorpe displacement ($d$) is calculated by comparing raw and reordered density profiles. Then, the dissipation rate of turbulent kinetic energy ($\varepsilon$) is expressed as:

$$\varepsilon = 0.64 \times L_T^2 \cdot N^3$$

(2)

where $N$ is the buoyancy frequency and is calculated as $N^2 = -g/\rho_0 \partial \rho/\partial z$; $g$ is the gravitational acceleration; $\rho$ and $\rho_0$ are the density of water and its reference, respectively.

To calculate the diapycnal nitrate vertical flux, the diapycnal diffusivity ($K_{\rho}$) is computed using the dissipation rate of turbulent kinetic energy ($\varepsilon$) and the squared buoyancy frequency ($N^2$) as follows (Osborn, 1980):

$$K_{\rho} = 0.2 \times \frac{\varepsilon}{N^2}$$

(3)

According to Fick’s law, the nitrate vertical turbulent flux is estimated using the diapycnal diffusivity ($K_{\rho}$) and the nitrate vertical concentration gradient as follows:

$$F_{\text{diff}} = -K_{\rho} \cdot \frac{\partial c}{\partial z}$$

(4)

where $F_{\text{diff}}$ is the nitrate vertical flux, and positive values correspond to an upward flux. $\partial c/\partial z$ is the nitrate vertical concentration gradient.

The squared vertical shear gradient ($S^2$) is calculated as:

$$S^2 = \left( \frac{\partial u}{\partial z} \right)^2 + \left( \frac{\partial v}{\partial z} \right)^2$$

(5)
where \( u \) and \( v \) are the zonal and meridional current velocities obtained by the LADCP.

2.3. Auxiliary Parameters

The NOAA 1/4° daily Optimum Interpolation Sea Surface Temperature (OISST; Huang et al., 2021) was used to calculate the dipole mode index (DMI), which is a sea surface temperature (SST) difference between the west (10°S–10°N, 50°E–70°E) and east (10°S–0°, 90°E–110°E) EIO (Saji et al., 1999). The daily surface current from altimetric geostrophic current and modeled Ekman current reprocessing, the 6 hr blended mean wind fields with 1/4° spatial resolution, the multisatellite merged sea level anomaly (SLA) L4 product, and the multisatellite merged Chl\( a \) L4 data are obtained from the Copernicus Marine Environment Monitoring Service. The monthly 20°C isotherm (D20) are obtained from the European Center for Medium-Range Weather Forecasting ocean reanalysis system, version 5 (Zuo et al., 2019). Anomalies of the above environmental parameters were normalized to their own climatology over the full period. The euphotic zone depth \( (Z_{eu}) \) was calculated from moderate-resolution imaging spectroradiometer (MODIS) data (Lin et al., 2016).

3. Results and Discussion

3.1. Biological Subsurface Under the Extreme pIOD Event

In the 2019 boreal autumn, an extreme pIOD event developed, and the DMI increased substantially beginning in August, peaking at 2.15°C in October and then decreasing to 0.31°C by the end of December 2019 (Figure 1c). In contrast, as a reference year, the DMI remained almost in the neutral range (±0.4°C) in 2017. The basin-wide distribution of the mean seasonal SST anomaly (SSTA) also showed an obvious dipole mode during 2019 but a nondipole mode in 2017 (Figures 1a–1b). To explore the oceanographic physical-biological responses to the pIOD event in the eastern EIO, we examined the vertical profiles of temperature, salinity, Chl\( a \) and nitrate along the equatorial transections under the decaying period of an extreme pIOD event in December 2019 and the neutral conditions in December 2017.

During December 2019, the surface layer was warm (>29°C), low salinity (33–34), nitrate-depleted and oligotrophic with low concentration of Chl\( a \) (<0.1 mg m\(^{-3}\)) (Figures 2a–2d). With the water depth increased to 30–75 m, the most striking characteristic was the coupling among the SCM (∼0.9 mg m\(^{-3}\)), the thermocline and nutricline, and the highest-salinity core (∼35.5) of the water column. Given the deep \( Z_{eu} \), ranging from 107 to 130 m (Figure 2), the subsurface phytoplankton were relieved from nutrient limitation. Notably, the depth of the coupled subsurface physical–biological phenomenon tilted zonally along the equatorial section, gradually shoaling from ∼75 in the west to ∼30 m in the east (Figures 2a–2d). The time-series variations of the D20 at the equator (80.5°E, 83°E, 86°E and 90°E) also showed a similar phenomenon (Figure 1d). Moreover, it was noteworthy that, with the occurrence of the high salinity core, the nitrate exhibited highly dynamic variabilities compared to other areas, indicating highly turbulent mixing in these areas.

In comparison, the oceanographic features of the eastern EIO in December 2017 were markedly different from those in 2019 (Figures 2e–2h). The warm (>28°C), less salty (33.6–34.5), and oligotrophic surface water covered only the upper 25 m in the east and middle of the section (85–93°E) but was absent on the west side. Under the surface layer, the highest salinities (35.5–35.6) and relatively warm water (28°C), with nitrate-depleted characteristics, occupied the subsurface layer (25–120 m). Corresondingly, the top of the thermocline was as deep as 100–120 m, which was roughly consistent with the depth of the nutricline and the base of the euphotic zone (104–116 m). Apparently, the phytoplankton in the euphotic layer suffered from severe nutrient limitation, leading to a low level of SCM (∼0.4 mg m\(^{-3}\)) in 2017 and approximately half of that in 2019 (Figure 2d). Remarkably, contrary to the 2019 situation, the depth of the SCM (30–75 m) was decoupled with the deepened nutricline and thermocline in 2017, likely indicating different nutrient sources (e.g., nitrogen fixation) to the sunlit layer in addition to replenishment from the deep water (Raes et al., 2015).

To more clearly present the prominent subsurface biological response to the extreme pIOD event, the nitrate inventory and integrated Chl\( a \) content in the euphotic layer were compared between 2019 and 2017 (Figure S2 in Supporting Information S1). Specifically, the averaged nitrate inventory of the euphotic zone was 742 ± 73 (95% confidence interval) mmol m\(^{-2}\) in 2019, while it was only 152 ± 56 mmol m\(^{-2}\) in 2017, indicating considerable differences in the nutrient replenishment between the pIOD and neutral conditions in the eastern EIO.
Figure 2. Sectional distribution of the hydrography and biogeochemistry in the upper 300 m along the Equator. Distributions of (a) potential temperature (°C), (b) salinity, (c) nitrate (μmol L⁻¹), and (d) Chl-a (mg m⁻³) in 2019. Same for (e–h), but in 2017. In each panel, the horizontal white and black dashed lines denote the thermocline and the euphotic depth (Zₑu), and the vertical black dot lines represent sampling profiles. In panel (a, b, e, and f), the dashed red lines represent the potential density (σθ). In panel (c and g), the thick black lines indicate the nutricline.
integrated Chl a content of the euphotic layer was an average of 43 ± 4 mg m$^{-2}$ in 2019, which was approximately 2 times greater than that (22 ± 4 mg m$^{-2}$) in 2017, and with significant difference between them ($t$-test, $t = 6.543, p < 0.01$). The results again supported that the robust subsurface biological activity in the eastern EIO may be associated with the extreme pIOD event in 2019.

3.2. Shoaled Nutricline Linked With the pIOD

The adjustment of the nutricline during the extreme pIOD conditions could have been the main reason for the distinct biological response in 2019. According to the time-series satellite physical data (Figure S3 in Supporting Information S1), the westerly zonal surface current, driven by the easterly zonal wind bursts, persisted from August to the middle of December and then turned inversely with the zonal wind in the eastern EIO in 2019. This result is verified by the wind and surface current at 10 m based on the Research Moored Array for African-Asian-Australian Monsoon Analysis and Prediction (RAMA) in the eastern EIO (Figure S4 in Supporting Information S1). Moreover, both the SLA and SSTA showed negative anomalies (Figure S3 in Supporting Information S1), indicating an evident upwelling feature in the eastern EIO during the pIOD event in 2019. This deduction is supported by the variation of D20, which was shoaled by ~30 m during the pIOD (Figure 1d and Figure S3 in Supporting Information S1). Notably, the surface current and SST changed inversely with the surface wind transition since the end of December 2019, but the shoaling anomaly of the thermocline lasted until January 2020. Horii et al. (2008) also found that the thermocline anomaly persisted longer than the surface physical signals in the eastern EIO during the pIOD in 2006. Furthermore, downwelling Rossby wave was forced simultaneously with the pIOD occurrence (Figure S3 in Supporting Information S1), and had propagated to the western EIO during the sampling period. Hence, the zonally tilted thermocline along the transection in the eastern EIO may be also associated with the Rossby wave.

By comparison, the physical forcings completely reversed during the neutral conditions in December 2017. The prevailing westerly zonal wind was accompanied by an eastward zonal surface current in the eastern EIO during the second half of 2017 (Figure S3 in Supporting Information S1). This phenomenon suggested that the warm and salty Wyrtki Jet occupied the eastern EIO from September to December 2017 (Duan et al., 2016; Schott et al., 2009). Accordingly, the thermocline depth was deepened due to the accumulation of surface water transported by the Wyrtki Jet (Duan et al., 2016; Rao et al., 2010; Schott et al., 2009). This inference was indicated by the deepened D20 in December 2017 (Figure 1d) and in phase with the variability of the positive SLA and SSTA in the eastern EIO (Figure S3 in Supporting Information S1). Similarly, upwelling Rossby wave was triggered in April 2017, then propagated to the west during the end of December 2017 (Figure S3 in Supporting Information S1). The relative shallow thermocline in the west of the transection (82–86°E) may be forced by the Rossby wave.

In summary, the variability of the thermocline in the eastern EIO, driven by the surface zonal wind stress transition, caused shoaling of the nutricline during the pIOD event than that under neutral condition. The shallower the nutricline is, the more beneficial it is to replenish nutrients from deep waters to the sunlit layer. In response, the time-series surface Chl a values suggested an apparent phytoplankton bloom to the east of the eastern EIO during the pIOD event in 2019 but no occurrence under neutral conditions in 2017 (Figure S3 in Supporting Information S1). More importantly, the biological response presented an apparent SCM (Figure 2d) over the entire eastern EIO rather than being confined to the upwelling zone along the eastern coast in 2019.

3.3. Elevated Diapycnal Nutrient Flux via Shear Instability During the pIOD

The nitrate inventory of the euphotic layer, however, did not wholly increase with the shoaling of the nutricline in 2019 (Figure S2 in Supporting Information S1). Instead, the high nitrate inventory tended to occur with the small squared buoyancy frequency ($N^2$) at the subsurface (Figure 3a), although where was well stratified in general, and occupied by the high-salinity core (Figure 2b). Moreover, the large vertical shear variance also occurred at these regions (Figure 3b), which was consistent with the strong instability in the vertical profiles of nitrate concentration. Hence, diapycnal turbulent mixing may be another important mechanism that regulates the replenishment of nutrients from deep water to the euphotic layer during pIOD event.
Figure 3. Sectional distribution of the water column stability and diapycnal nitrate flux in the upper 300 m along the Equator. Distribution of (a) buoyancy frequency squared ($s^{-2}$), (b) vertical shear squared ($s^{-2}$), (c) diapycnal diffusivity ($m^2 s^{-1}$), and (d) nitrate diffusive flux (mmol m$^{-2}$ d$^{-1}$) in 2019. Same for (e–h), but in 2017. In each panel, the horizontal white and black dashed lines denote the thermocline and the euphotic depth ($Z_{eu}$), respectively. In panel (d), the downward nitrate fluxes were not shown between 200 and 275 m. In panel (h), the zero nitrate flux was not shown in the upper layer.
As shown in Figure S5 of Supporting Information S1, the turbulent kinetic energy dissipation rate ($\varepsilon > 10^{-7}$ W kg$^{-1}$) was high in general at the subsurface thermocline, particularly peaking ($10^{-6}$–$10^{-5}$ W kg$^{-1}$) with the high-salinity core occurrence in 2019. The only exception was the high $\varepsilon$ ($\sim 10^{-6}$ W kg$^{-1}$) in deep waters (150–200 m) on the eastern side of the transect (90°-93°E). The pattern of vertical diapycnal diffusivity ($K_{\rho}$, Figure 3c) was in phase with the dissipation rate variability, which was elevated from a base value ($10^{-5}$ m$^{2}$ s$^{-1}$) to a high level ($10^{-2}$–$10^{-1}$ m$^{2}$ s$^{-1}$) at the thermocline and the deep waters in the east. As a result, when the largest $K_{\rho}$ and vertical nitrate gradient (Figure S5 in Supporting Information S1) were consistent with the high-salinity core, the turbulent nitrate flux ($F_{\text{diff}}$) around the base of the euphotic layer reached a maximum value of 120 mmol m$^{-2}$ d$^{-1}$, approximately two orders of magnitude higher than those without saline water (Figure 3d). This result indicates that the turbulent mixing process played an important role in sustaining the nutrient supply to the SCM during the pIOD event in 2019. Moreover, the $F_{\text{diff}}$ also increased to a high level (0.5–10 mmol m$^{-2}$ d$^{-1}$) around the nitratcline, suggesting that the diapycnal nutrient flux could penetrate the mostly stratified water to the oligotrophic well-mixed surface water during the pIOD event. By comparison, the deep thermocline and strong stratification may inhibit turbulent nutrient supply to the euphotic zone in 2017. As shown in Figure 3e, the depth of maximum buoyancy frequency was deeper than $Z_{eu}$ in 2017, and its magnitude was one order greater than that in 2019. Furthermore, the vertical shear in the thermocline (Figure 3f) was weaker in 2017 than in 2019. Although the $\varepsilon$ (Figure S5 in Supporting Information S1) was greater in 2017 than in 2019 on average, the decreased $K_{\rho}$ in the thermocline (Figure 3g) hampered the $F_{\text{diff}}$ (Figure 3b) across the intensely stratified thermocline to the euphotic zone between 86 and 93°E. With uplifting of the thermocline/nutricline, the $F_{\text{diff}}$ increased around the $Z_{eu}$ in the west (82–86°E). In summary, the $F_{\text{diff}}$ was generally restricted deeper than 90–120 m in 2017, while the $F_{\text{diff}}$ could reach to $\sim$30 m in 2019.

The enhanced vertical turbulence and nitrate flux within the thermocline may be forced by the strong shear instability, attributed to the dynamic ocean circulation in 2019. Because the pIOD event was in its decaying stage at the end of December 2019, the surface current changed eastward (Figure S3 in Supporting Information S1). However, according to the zonal currents observed by the RAMA moorings at 0°, 80.5°E and 0°, 90°E and the Nucleus for European Modeling of the Ocean (Madec, 2008), a strong eastward EUC developed beneath the westward surface current in the eastern EIO from August to December during the pIOD event in 2019 (Figures S4 and S6 in Supporting Information S1). Considering that the EUC can bring the saltier Arabian Sea water to the eastern EIO (Chen et al., 2016; Hase et al., 2008), the appearance of the high-salinity cores at the thermocline could be the residual signal of the EUC. The water-mass signature of the high-salinity cores (Figure 2b) was well consistent with the range of $\sigma_{\theta}$ (22.8–24.0 kg m$^{-3}$) for the Arabian Sea High-Salinity Water (Jain et al., 2017; Sanchez-Franks et al., 2019). In other words, the large diapycnal nitrate fluxes and strong vertical shear were generated via the westward-flowing oligotrophic surface water and the opposing eastward-flowing EUC under pIOD conditions. Notably, beneath the thermocline, the $F_{\text{diff}}$ was downward between 200 and 275 m, further indicating the unique role of the nutrients carried by the EUC in the biogeochemistry in the eastern EIO during the pIOD event.

4. Conclusions

Based on in situ bio-physical data collected during decaying stage of pIOD and neutral conditions, we assessed the nutrient replenishment and subsequent biological response to climate change in the eastern EIO. During the pIOD phase, forced by the unique coupled atmospheric and oceanic processes, the combined effect of thermocline shoaling and subsurface turbulent mixing triggered an apparently elevated nutrient supply and robust subsurface biological response (Figure 4a). In neutral phase, the physical forcing was in the climatological conditions, and the deepened thermocline and strong stratification created an oligotrophic sunlit layer with low phytoplankton biomass in the eastern EIO (Figure 4b). With the projected threefold increase in extreme pIOD events over the twenty-first century (Cai et al., 2014) and by further altering of the mean climate conditions of the EIO to more closely resemble a pIOD state under greenhouse warming (Cai et al., 2013), we should expect active biogeochemical cycles fueling a more productive eastern EIO in the future.
Figure 4. Conceptual scheme of the subsurface biogeochemical responses to physical dynamics during positive Indian Ocean Dipole (IOD) and neutral conditions in the eastern equatorial Indian Ocean (EIO). (a) During the positive IOD, the reinforced easterly wind forces the eastward surface current, which lifts the thermocline/nutricline into the euphotic zone owing to Ekman divergence in the eastern EIO and enhances the turbulent nutrient flux via strong vertical shear associated with the opposite-flowing surface current and subsurface Equatorial Undercurrent (EUC). The subsurface biological response is subsequently enhanced due to their combined effect. (b) Under neutral condition, the eastward surface current driven by prevailing westerly winds deepens the thermocline/nutricline and thus causes strong stratification, which creates an oligotrophic euphotic zone with low phytoplankton biomass in the eastern EIO.

Data Availability Statement
The authors acknowledge the NOAA for providing the daily Optimum Interpolation (OI) Sea Surface Temperature (SST) V2 at https://www.esrl.noaa.gov/psd/data/gridded/data.noaa.oisst.v2.html, the Copernicus Marine Environment Monitoring Service (CMEMS) for providing the daily surface current, the 6-hourly blended mean wind fields, the sea level anomaly (SLA) L4 product, and the multisatellite merged Chla L4 data at https://resources.marine.copernicus.eu, the Copernicus Climate Change Service (C3S) for providing the monthly 20°C isotherm (D20) at https://cds.climate.copernicus.eu, the MODIS for euphotic zone depth (Z_eu) at https://oceancolor.gsfc.nasa.gov. In situ observational data sets for this research are available at this site (https://zenodo.org/record/6469059%23.Yl4keuhByUk).

Acknowledgments
This study was supported by the Global Change and Air-Sea Interaction II Program (Grant Nos. GASI-04-HYST-01, GASI-01-EIND-STwin), Natural Science Foundation of China (Grant Nos. 42176039, 42176148, 42076134, 41876028), the Special Fund for Basic Scientific Research of the Second Institute of Oceanography, State Oceanic Administration (Grant No. ZZ2001), the China Ocean Mineral Resources R & D Association (Grant No. DY135-E2-4-03), and the Tsinghua Scholars Programs of Shandong Province (No. tsqnc201909165, 20190963). The authors would like to thank the crew of R/V Xianyanghong-06 for their assistance in the deployment and recovery of the instruments.

References
Baer, S. E., Rauschenberg, S., Garcia, C. A., Garcia, N. S., Martiny, A. C., Twining, B. S., & Lomas, M. W. (2019). Carbon and nitrogen productivity during spring in the oligotrophic Indian Ocean along the GO-SHIP IO9N transect. Deep-Sea Research Part II Topical Studies in Oceanography, 161(80), 81–91. https://doi.org/10.1016/j.dsr2.2018.11.008
Cai, W., Santos, A., Wang, G., Weller, E., Wu, L., Ashok, K., et al. (2014). Increased frequency of extreme Indian Ocean dipole events due to greenhouse warming. Nature, 510(7494), 254–258. https://doi.org/10.1038/nature13327
Cai, W., Zheng, X. T., Weller, E., Collins, M., Cowan, T., Lengaigne, M., et al. (2013). Projected response of the Indian Ocean dipole to greenhouse warming. Nature Geoscience, 6(12), 999–1007. https://doi.org/10.1038/ngeo2009
Cermeño, P., Dutkiewicz, S., Harris, R. P., Follows, M., Schofield, O., & Falkowski, P. G. (2008). The role of nutricline depth in regulating the ocean carbon cycle. Proceedings of the National Academy of Sciences of the United States of America, 105(51), 20344–20349. https://doi.org/10.1073/pnas.0811302106
Chavez, F. P., Messé, M., & Pennington, J. T. (2011). Marine primary production in relation to climate variability and change. Annual Review of Marine Science, 3(1), 227–260. https://doi.org/10.1146/annurev.marine.010908.163917
Chen, G., Han, W., Li, Y., Wang, D., & McPhaden, M. J. (2015). Seasonal-to-interannual time-scale dynamics of the equatorial undercurrent in the Indian Ocean. Journal of Physical Oceanography, 45(6), 1532–1553. https://doi.org/10.1175/JPO-D-14-0225.1
Chen, G., Han, W., Shu, Y., Li, Y., Wang, D., & Xie, Q. (2016). The role of equatorial undercurrent in sustaining the eastern Indian Ocean upwelling. Geophysical Research Letters, 43(12), 6444–6451. https://doi.org/10.1002/2016GL069433
Du, Y., Zhang, Y., Zhang, L. Y., Touzuka, T., Ng, B., & Cai, W. (2020). Thermocline warming induced extreme Indian Ocean dipole in 2019. Geophysical Research Letters, 47(18), 1–10. https://doi.org/10.1029/2020GL090079
Duan, Y., Liu, L., Han, G., Liu, H., Yu, W., Yang, G., et al. (2016). Anomalous behaviors of Wyrtki jets in the equatorial Indian Ocean during 2013. Scientific Reports, 6(1), 1–7. https://doi.org/10.1038/srep29688
Grasshoff, K., Kremling, K., & Manfred, E. (1999). Methods of seawater analysis (3rd ed.). Wiley-VCH.
Gregg, M. C. (1987). Diapycnal mixing in the thermocline: A review. Deep Sea Research - Part B: Oceanographic Literature Review, 92(C5), 5249–5286. https://doi.org/10.1029/93JB005249
Hase, H., Masumoto, Y., Kuroda, Y., & Mizuno, K. (2008). Semiannual variability in temperature and salinity observed by triangle trans-ocean buoy network (TRITON) buoys in the eastern tropical Indian Ocean. *Journal of Geophysical Research, 113*, C01016. https://doi.org/10.1029/2006JC004026

Hood, R. R., Beckley, L. E., & Wiggert, J. D. (2017). Biogeochemical and ecological impacts of boundary currents in the Indian Ocean. *Progress in Oceanography, 150*(9), 299–325. https://doi.org/10.1016/j.pocean.2017.04.011

Horii, T., Hase, H., Ueki, I., & Masumoto, Y. (2008). Oceanic precondition and evolution of the 2006 Indian Ocean dipole. *Geophysical Research Letters, 35*(3), 1–6. https://doi.org/10.1029/2007GL032464

Huang, B., Liu, C., Banzon, V., Freeman, E., Graham, G., Hankins, B., et al. (2021). Improvements of the daily optimum interpolation sea surface temperature (DOISST) version 2.1. *Journal of Climate, 34*(8), 2923–2939. https://doi.org/10.1175/JCLI-D-20-0166.1

Iskandar, I., Masumoto, Y., & Mizuno, K. (2009). Subsurface equatorial zonal current in the eastern Indian Ocean. *Journal of Geophysical Research, 114*, C06005. https://doi.org/10.1029/2008JC005188

Jain, V., Shankar, D., Vinayachandran, P. N., Kankanakor, A., Chatterjee, A., Amol, P., et al. (2017). Evidence for the existence of Persian Gulf water and red Sea water in the Bay of Bengal. *Climate Dynamics, 48*(9–10), 3207–3226. https://doi.org/10.1007/s00382-016-3259-4

Kumar, S. P., David, T. D., Byju, P., Narvekar, J., Yoneyama, K., Nakatani, N., et al. (2012). Bio-physical coupling and ocean dynamics in the central equatorial Indian Ocean during 2006 Indian Ocean dipole. *Geophysical Research Letters, 39*(14), 1–6. https://doi.org/10.1029/2012GL052609

Liao, X., Du, Y., Wang, T., He, Q., Zhan, H., Hu, S., & Wu, G. (2020). Extreme phytoplankton blooms in the southern tropical Indian Ocean in 2011. *Journal of Geophysical Research: Oceans, 125*, 1–20. https://doi.org/10.1029/2019JC015649

Lin, J., Lee, Z., Onurmuş, M., & Kabru, M. (2016). Attenuation coefficient of useable solar radiation of the global oceans. *Journal of Geophysical Research: Oceans, 121*, 3228–3236. https://doi.org/10.1002/2015JC011690

Liu, C., Kohl, A., Liu, Z., Wang, F., & Stammer, D. (2016). Deep-reaching thermocline mixing in the equatorial pacific temperature. *Nature Communications, 7*(1), 1–11. https://doi.org/10.1038/ncomms11576

Liu, B., & Ren, H. L. (2020). What caused the extreme Indian Ocean dipole event in 2019? *Geophysical Research Letters, 47*(11), 1–8. https://doi.org/10.1029/2020GL087768

Macedo, G. (2008). NEMO ocean engine. *Note de Pôle de modélisation*. Institut Pierre-Simon Laplace (IPSL). Retrieved from http://eprints.soton.ac.uk/64324/

Murtugudde, R. G., Signorini, S. R., Christian, J. R., Busalacchi, A. J., McClain, C. R., & Picaut, J. (1999). Ocean color variability of the tropical Indo-Pacific basin observed by SeaWiFS during 1997-1998. *Journal of Geophysical Research, 104*, 18351–18366. https://doi.org/10.1029/1999jc900135

Nagura, M., & McPhaden, M. J. (2010). Dynamics of zonal current variations associated with the Indian Ocean dipole. *Journal of Geophysical Research, 115*, C11026. https://doi.org/10.1029/2010JC006423

Osborn, T. (1980). Estimates of the local rate of vertical diffusion from dissipation measurements. *Journal of Physical Oceanography, 10*(1), 83–89. https://doi.org/10.1175/1520-0485(1980)010<0083:EOTLRO>2.0.CO;2

Raes, E. J., Thompson, P. A., McKinnes, A. S., Nguyen, H. M., Hardman-Mountford, N., & Waite, A. M. (2015). Sources of new nitrogen in the Indian Ocean. *Global Biogeochemical Cycles, 29*(8), 1283–1297. https://doi.org/10.1002/2015GB005194

Rao, R. R., Girish Kumar, M. S., Ravichandran, M., Rao, A. R., Gopalakrishna, V. V., & Thadathil, P. (2010). Interannual variability of Kelvin wave propagation in the wave guides of the equatorial Indian Ocean, the coastal Bay of Bengal and the southeastern Arabian Sea during 1993-2006. *Deep-Sea Research Part I Oceanographic Research Papers, 57*(1), 1–13. https://doi.org/10.1016/j.dsr.2009.10.008

Riten, T., Gaye, B., Emieis, K. C., & Ramaswamy, V. (2019). The ballast effect of lithogenic matter and its influences on the carbon fluxes in the Indian Ocean. *Biogeosciences, 16*(2), 485–503. https://doi.org/10.5194/bg-16-485-2019

Saji, N. H., Goswami, P. N., Vinayachandran, P. N., & Yamagata, T. (1999). A dipole mode in the tropical Indian Ocean. *Journal of Geophysical Research, 104*, 8361–8380. https://doi.org/10.1029/1999JC900340

Sanchez-Franks, A., Webber, B. G. M., King, B. A., Vinayachandran, P. N., Matthews, A. J., Sheehan, P. M. F., et al. (2019). The railroad switch for ocean and sea-ice: A description of the system and assessment. *Ocean Science, 15*(3), 779–808. https://doi.org/10.5194/os-15-779-2019