Suppressed Upwelling Events in the Seychelles–Chagos Thermocline Ridge of the Southwestern Tropical Indian Ocean

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Abstract
A prominent subsurface upwelling over the Seychelles–Chagos Thermocline Ridge (SCTR) in the southwestern tropical Indian Ocean is suppressed when downwelling Rossby waves propagate from the eastern Indian Ocean during the positive phase of the Indian Ocean Dipole (IOD) or El Niño periods. Recent studies have suggested that the upwelling can be further suppressed during the co-occurrence years of positive IOD and El Niño, associated with the strong easterly wind anomalies in the eastern Indian Ocean. This study examines the temporal variations in the SCTR upwelling strength during 1968–2017, identifies strong upwelling suppression events, and analyzes their characteristics by focusing on the role of both remote and local wind forcing, not limited to the remote influence linked to the IOD and the El Niño Southern Oscillation. A total of nine events were identified for the 50-year period, with seven of them occurring 3–6 months after the co-occurring peaks of positive IOD and El Niño in the eastern Indian Ocean. However, only IOD exhibited a positive phase before the 2011–2012 suppression event associated with the downwelling-favorable anticlockwise local wind anomalies in the SCTR. Moreover, the 1978–1979 suppression event was primarily caused by the strong anticlockwise wind anomalies in the SCTR, without any significant influence of the positive IOD or El Niño. The results suggest that the role of local winds, as well as the remote forcing, are significant in the upwelling suppression over the SCTR, which would contribute to understanding the potential predictability of the upwelling variations.

Keywords El Niño–Southern Oscillation · Indian Ocean Dipole · Seychelles–Chagos Thermocline Ridge · Southwestern tropical Indian Ocean · Upwelling

1 Introduction
The Seychelles–Chagos Thermocline Ridge (SCTR) is a prominent upwelling region in the southwestern tropical Indian Ocean (Woodberry et al. 1989; McCreary et al. 1993; Schott et al. 2009) (Fig. 1a). It has also been referred to as the Seychelles Dome or Thermocline Ridge of the Indian Ocean (Hermes and Reason 2008; Yokoi et al. 2008; Jayakumar et al. 2011). Because of the nutrient-rich water brought up by the upwelling, the SCTR is known to exhibit high biological productivity (Xie et al. 2002; Dilmahamod et al. 2016). However, the upwelling is not easily depicted from sea surface temperature (SST) since it exhibits a subsurface cooling, not a typical surface cooling (Fig. 1b). The SCTR upwelling is thus identified by a shallow thermocline depth, and the 20 °C isotherm depth (D20) is used to represent the strength of the subsurface upwelling (Xie et al. 2002) (Fig. 1a). The shallow D20 but still warm SST contributes to a strong air–sea interaction in the SCTR (Xie et al. 2002). It has been noted that the temperature variability over the SCTR is associated with tropical cyclone activity and anomalous rainfall in East Africa (Jury et al. 1999; Xie et al. 2002; Ummenhofer et al. 2009; Vialard et al. 2009).

The SCTR upwelling occurs to the south of the equator because the mean wind along the equator is not easterly, which is different from the Pacific and Atlantic Oceans, where the easterly trade wind dominates along the equator (Schott et al. 2009; Wang and McPhaden, 2017). A major driving force of the SCTR upwelling is negative (clockwise) wind stress curl by the southeasterly trades to the south of the SCTR and the equatorial westerlies to the north of the...
SCTR, which is relatively strong in boreal winter (Masumoto and Meyers 1998; Xie et al. 2002; Hermes and Reason 2008; Yokoi et al. 2008; Vialard et al. 2009) (Fig. 1a). The southeasterly trades to the south of the SCTR drive the South Equatorial Current, whereas the equatorial westerlies to the north of the SCTR drive the South Equatorial Counter Current (L’Hégaret et al. 2018).

The SCTR upwelling exhibits significant interannual variability, and the upwelling is known to be particularly suppressed related to the positive phase of the Indian Ocean Dipole (IOD) or the El Niño Southern Oscillation (ENSO) (Chambers et al. 1999; Klein et al. 1999; Xie et al. 2002; Vinayachandran and Saji 2008; Rao and Behera 2005; Yu et al. 2005). The easterly wind anomalies in the eastern Indian Ocean during the positive IOD or ENSO induce downwelling waves in the ocean, and the westward propagation of the downwelling Rossby waves suppresses the SCTR upwelling (Masumoto and Meyers 1998; Rao and Behera 2005; Yu et al. 2005). Moreover, during the co-occurrence years of the positive IOD and ENSO, the westward propagation of downwelling Rossby waves tends to sustain for a longer period due to persistent easterly wind anomalies in the eastern Indian Ocean, which causes stronger suppression of the upwelling in the SCTR (Chakravorty et al. 2014).

Most of the previous studies have focused on the role of remote impact on the upwelling variability in the SCTR associated with the IOD and ENSO. However, a recent study by Tozuka et al. (2010) conducted a composite analysis of the upwelling suppression events for the period of 1980–2007 based on the Ocean General Circulation Model experiments and showed that the local wind forcing could not be neglected in discussing the interannual variability of the SCTR upwelling. Considering a few years of time intervals between the upwelling suppression events, the role of local wind forcing needs to be further examined for a longer period to include a greater number of cases. In addition, the previous results based on the numerical model experiments require further confirmation using a reanalysis dataset.

The present study thus aims to understand the characteristics of the upwelling suppression events from a 50-year reanalysis dataset. This long-term data analysis, compared to the previous studies, would help examine the influence of IOD and ENSO on the suppression events with more cases. Further, the role of local wind forcing is discussed in comparison with the remote effect of the westward propagation of downwelling anomalies from the eastern Indian Ocean, based on the ocean reanalysis dataset. The rest of the paper is organized as follows. A description of the data and methods used in this study is given in the next section. Section 3 provides an examination of the upwelling suppression events. A summary and discussions follow in Sect. 4.

### 2 Data and Methods

The SST was obtained from the Hadley Centre Global Sea Ice and Sea Surface Temperature (HadISST) with a spatial resolution of 1° (https://climatedataguide.ucar.edu/climatedata/sst-data-hadisst-v11). The ocean subsurface temperature was provided by the Simple Ocean Data Assimilation (SODA) version 2.2.4 for the period 1968–1979 and by the SODA version 3.4.2 for the period 1980–2017 (https://www2.atmos.umd.edu/~ocean/). The SODA is based on the Parallel Ocean Program with a spatial resolution of 0.5°, and the number of vertical levels is 40 for version 2.2.4 and 50 for version 3.4.2. The SODA is a reanalysis product assimilated using the various in situ hydrographic data and satellite data based on the ensemble Kalman filter (Carton et al. 2018). The 10 m winds were obtained from the National Center for Environmental Prediction-National Centers for Atmospheric Research (NCEP-NCAR) reanalysis data available on a T62 Gaussian grid.

All the data were obtained as monthly means, and then monthly anomalies were calculated as deviations from their monthly climatology during 1968–2017. The monthly anomalies were low-pass filtered at 12 months to examine their interannual variability related to the upwelling in the
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The “anomalies” thus refer to the low-pass filtered time series of the deviations from the monthly climatology.

The Dipole Mode Index (DMI), which is defined as the difference of the SST anomaly between the western tropical Indian Ocean (10° S–10° N; 50° E–70° E) and the southeastern tropical Indian Ocean (10° S–0° N; 90° E–110° E), was obtained from National Oceanic and Atmospheric Administration (NOAA). The Ocean Niño Index (ONI), which is defined as the SST anomalies averaged over the Niño-3.4 region (5° S–5° N; 170° W–120° W), was also obtained from NOAA. The DMI and ONI are used as the indices of the IOD and ENSO, respectively. The strong positive phase of IOD and ENSO were defined based on the normalized time series of the DMI and ONI.

The area-averaged D20 over the SCTR region (10° S–5° S, 48° E–68° E) presents the strength of SCTR upwelling (Fig. 2a). The Seychelles Upwelling Index (SUI) was defined as the normalized anomalies of the area-averaged D20 for the period of 1968–2017. A positive SUI (shallower D20) indicates a strengthening of the upwelling, and a negative SUI (deeper D20) a suppression of the upwelling. Note that the SUI calculated from SODA shows a good agreement (correlation coefficient: 0.88) with that from the quality-controlled EN4 observational dataset from Met Office Hadley Centre (Good et al. 2013; https://www.metoffice.gov.uk/hadobs/en4/).

A total of nine cases were identified as suppressed upwelling events based on the SUI anomalies greater than one standard deviation (σ). Figure 2b shows the seven events (72–73, 82–83, 91–92, 94–95, 97–98, 06–07, and 15–16) that occurred after both the IOD and ENSO displayed a strong positive phase (> 1σ). However, only the IOD showed a positive phase before the 11–12 suppression event, and neither the IOD nor ENSO showed a significant positive phase before the 78–79 suppression event. Based on the comparison between the SUI, DMI, and ONI, a composite analysis was conducted for the seven events, and the results were compared with the other two events, 78–79 and 11–12.

3 Results

3.1 The Composite of the Seven Suppression Events

Figure 3 illustrates the composites of D20 anomalies and wind stress anomalies for the seven suppressed upwelling events (72–73, 82–83, 91–92, 94–95, 97–98, 06–07, and 15–16) that occurred after both the IOD and ENSO displayed strong positive phase. Note that the spatial pattern of this composite is generally opposite to the positive composite (SUI > 1), and the difference between the composite of the seven events and the positive composite is greater than
the standard deviations of the D20 anomalies (not shown). Peak (0) indicates when the SUI exhibits a negative maximum during the event, i.e., the peak of the suppression. The timing of the negative maximum varies, ranging from December to April. Before the peak of the seven suppression events, easterly wind anomalies are observed in the eastern Indian Ocean, which is related to the downwelling D20 anomalies. Westward propagation of the downwelling anomalies, in turn, induces the suppression of upwelling in the SCTR after a few months. The downwelling anomalies in the central Indian Ocean start developing about 6 months before the suppression peak and reach the eastern boundary of the SCTR about 4 months before the suppression peak. The downwelling anomalies in the SCTR last until 3 months after the suppression peak and start to disappear about 4 months after the peak. The composite analysis indicates that the strong easterly wind stress anomalies in the eastern Indian Ocean induce the downwelling anomalies, which propagate westward and suppress the upwelling in the SCTR.

The time–longitude plot of the D20 anomalies for the composites of the seven suppression events presents the characteristics of westward propagation from the eastern Indian Ocean (Fig. 4a). The strong positive D20 anomalies in the SCTR (>20 m) originate from the longitude of 70° E–90° E about 6 months before the suppression peak. The time–longitude plot of the Ekman pumping velocity anomalies for the composites of the seven suppression events presents strong negative (downwelling) anomalies for the composites of the seven suppression events. The contour intervals are 10 m and $1 \times 10^{-6}$ m/s, respectively. The plots are produced by averaging over the latitude band of the SCTR (5° S–10° S). Red lines denote the longitude band of the SCTR (48° E and 68° E).
with persistent southeasterly wind anomalies in the eastern Indian Ocean more than 6 months before the suppression peak (Fig. 4b). However, the negative Ekman velocity anomalies are not as strong over the SCTR, indicating a limited influence of the local wind stress curl on the suppression events.

Figure 5 shows the composites of D20 anomalies, wind stress anomalies, and Ekman pumping velocity anomalies based on the DMI and ONI. During the 50 years (1968–2017), 37 months are identified as the co-occurrence months of the positive IOD and El Niño (normalized DMI ≥ 1 and normalized ONI ≥ 1). Note that 50 months are identified as the positive IOD, and 47 months are identified as the positive ENSO. The composites are the averages for only peak months during each period of the corresponding phase (one peak month per each period of the positive phase). The spatial patterns of the D20 anomalies during the peak months of the co-occurrence are similar to those of the composites in Fig. 3. Moreover, the D20 anomalies are stronger during the co-occurrence period (Fig. 5a, b) than those during the DMI-only peak months (Fig. 5c), which supports the further suppression of the SCTR upwelling during the co-occurrence of positive IOD and ENSO. The negative Ekman velocity anomalies in the eastern Indian Ocean are also stronger during the co-occurrence period compared to those during the DMI-only peak months. However, the D20 anomalies during the ENSO-only peak months are not particularly weak compared to those during the co-occurrence period despite weaker negative wind anomalies in the eastern Indian Ocean.

3.2 78–79 Event and 11–12 Event

The spatial patterns of the D20 anomalies and wind stress anomalies during the 78–79 suppression event are shown in Fig. 6. Neither the IOD nor the ENSO exhibited strong positive phases before the peak of 78–79 event. The easterly wind anomalies are not clearly seen in the eastern Indian
Ocean before the suppression peak. The downwelling anomalies started developing to the southeast of the SCTR about 6 months before the suppression peak (November in 1978), extended to the SCTR almost in the same month, and lasted until a few months after the peak of suppression. Figure 7 shows the time–longitude plots of D20 anomalies and Ekman pumping velocity anomalies for the 78–79 event averaged over the latitude of SCTR (5°S–10°S). The positive D20 anomalies (> 30 m) are observed at about 60°E inside the SCTR, which suggests that the 78–79 event is different from the seven events in terms of the remote influence from the eastern Indian Ocean. The location of strong negative Ekman velocity anomalies is also different from that of the composites of seven events in Fig. 4a. The strong D20 anomalies (> 30 m) are observed at about 60°E inside the SCTR, which suggests that the 78–79 event is different from the seven events in terms of the remote influence from the eastern Indian Ocean.

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Figure 8 shows the spatial patterns of D20 anomalies and wind stress anomalies during the 11–12 event (peak month: September in 2011). This event occurred after only IOD displayed a significant positive phase. In the eastern Indian Ocean, the southerly or southeasterly wind stress anomalies are not observed similar to those during the 78–79 event. Relatively weak downwelling anomalies are found in the eastern Indian Ocean about 6 months before the suppression peak of the SCTR upwelling. The
downwelling anomalies seem to propagate westward to the SCTR, but the propagation tends to be confined within a relatively narrow latitudinal band. This has been reported as one of the typical characteristics that can be found during the positive IOD (Rao and Behera 2005; Yu et al. 2005). The time–longitude plots of D20 anomalies and Ekman pumping velocity anomalies for the 11–12 event are shown in Fig. 9. During this event, the propagation characteristics of the downwelling anomalies are not as clear as those observed during the seven events in Fig. 4a. The negative Ekman velocity anomalies are observed both to the east and inside the SCTR. For the 11–12 event, the remote influence associated with the positive IOD may have played a role in the suppression of the SCTR upwelling, but a local wind forcing seems to be an additional factor to suppress the upwelling further.

### 4 Summary and Discussions

This study investigated the temporal changes in the upwelling strength of the SCTR, which would have an important implication in the biological productivity and air–sea interaction over the tropical Indian Ocean. During the 50 years of the analysis period, there were nine events of strong suppression of the upwelling. Seven of them occurred under the influence of both positive IOD and ENSO. This result is consistent with the previous studies that showed the easterly wind anomalies during the positive IOD and ENSO generate downwelling anomalies which propagate westward and suppress the upwelling in the SCTR (Xie et al. 2002; Vinayachandran and Saji 2008; Rao and Behera 2005; Yu et al. 2005). However, the suppression event during 1978–1979 occurred without any significant influence of the IOD and ENSO. The event during 2011–2012 occurred after a significant positive phase of IOD only. The 11–12 event was relatively weak compared to the seven events that occurred after the positive phase of both IOD and ENSO,
but the suppression during the 78–79 event was relatively strong (Fig. 2).

The suppression events without a remote influence associated with the IOD and ENSO suggest an important role of local winds on the upwelling variability in the SCTR. Figure 10 shows a comparison of the Ekman pumping velocity anomalies at the peaks of each suppression event and the time series of area-averaged Ekman pumping velocity anomalies in the SCTR. During a typical upwelling period in the SCTR (SU1 > 1), the Ekman velocity anomalies exhibit positive values. However, the Ekman velocity anomalies averaged over the SCTR exhibit negative values during the suppression peaks of all the nine events, which indicates a role of local winds even during the seven events that occurred after the positive phase of both IOD and ENSO. Over the SCTR, the negative Ekman velocity anomalies are particularly stronger during the peaks of 78–79 and 11–12 events. It is also notable that the negative Ekman velocity anomalies exhibit their peaks before the upwelling suppression peak during the seven events and the 11–12 events, whereas the peak of the negative Ekman velocity anomalies occurred at the time of the upwelling suppression peak during the 78–79 event.

Previous studies have emphasized the role of IOD and ENSO on the interannual upwelling variability over the SCTR. Considering the potential predictability of the IOD and ENSO, the strong suppression of the SCTR upwelling may also be predictable based on the DMI and ONI. For example, the SCTR upwelling could be suppressed after a few months when the positive IOD or ENSO develops. The suppression could be stronger when both of them occur at the same time. However, this study suggests that the role

![Figure 10](image-url)
of local winds, as well as the remote forcing, is significant in the upwelling suppression over the SCTR based on the long-term reanalysis dataset for 50 years.

A quantitative comparison between the contribution of the remote forcing and the local winds remains a future study. It will clarify the relative importance of the forcing on the upwelling suppression. In addition, the suppression-favorable anticlockwise wind anomalies need to be further examined, considering a possible effect of the IOD and ENSO on the local winds, not only on the winds in the eastern Indian Ocean. Despite the above limitations, this study would contribute to a better understanding of the interannual variability of the upwelling in the SCTR and a possible improvement of the upwelling prediction by emphasizing the role of local winds together with the effects of the IOD and ENSO.

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