

## The Relation of Wind-Driven Coastal and Offshore Upwelling in the Benguela Upwelling System

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**ABSTRACT:** Spatial and temporal variations of nutrient-rich upwelled water across the major eastern boundary upwelling systems are primarily controlled by the surface wind with different, and sometimes contrasting, impacts on coastal upwelling systems driven by alongshore wind and offshore upwelling systems driven by the local wind stress curl. Here, concurrently measured wind fields and satellite-derived chlorophyll-*a* concentration, along with a state-of-the-art ocean model simulation, spanning 2008–18 are used to investigate the connection between coastal and offshore physical drivers of the Benguela upwelling system (BUS). Our results indicate that the spatial structure of long-term mean upwelling derived from Ekman theory and the numerical model is fairly consistent across the entire BUS and is closely followed by the chlorophyll-*a* pattern. The variability of the upwelling from the Ekman theory is proportionally diminished with offshore distance, whereas different and sometimes opposite structures are revealed in the model-derived upwelling. Our result suggests the presence of submesoscale activity (i.e., filaments and eddies) across the entire BUS with a large modulating effect on the wind-stress-curl-driven upwelling off Lüderitz and Walvis Bay. In Kunene and Cape Frio upwelling cells, located in the northern sector of the BUS, the coastal upwelling and open-ocean upwelling frequently alternate each other, whereas they are modulated by the annual cycle and are mostly in phase off Walvis Bay. Such a phase relationship appears to be strongly seasonally dependent off Lüderitz and across the southern BUS. Thus, our findings suggest that this relationship is far more complex than is currently thought and seems to be sensitive to climate changes, with short- and far-reaching consequences for this vulnerable marine ecosystem.

**KEYWORDS:** Ekman pumping; Wind; Atmosphere-ocean interaction; Climate change; Ocean models; Climate variability

### 1. Introduction


Similar to other major upwelling systems, any changes in the intensity and timing of upwelling and associated nutrient flux have different and sometimes contrasting impacts on the growth of individual phytoplankton size classes across the Benguela upwelling system (BUS; Rykaczewski et al. 2008; Lamont et al. 2019). In fact, this determines the regional diversity and population of the commercially valuable pelagic fishes (Rykaczewski et al. 2008). Given the increased observational evidence concerning natural and anthropogenic climate change across the BUS (Doney et al. 2012; Bachèlery et al. 2016; Bonino et al. 2019), sensitivity of the local upwelling (e.g., timing, intensity, duration) to the regional surface atmospheric flow is of great ecological and socioeconomic importance.

The BUS is characterized by distinctive features in comparison with other main eastern boundary upwelling systems (Fig. 1). It is identified with a perennial upwelling across its

northern part and a strong seasonal cycle across the southern part that peaks between November and March (Bakun and Nelson 1991; Shannon and Nelson 1996; Duncombe Rae 2005; Lass and Mohrholz 2008; Junker et al. 2015, 2017). It is bounded both southward and northward by warm waters (Shannon and Nelson 1996; Hutchings et al. 2009; Bachèlery et al. 2016; Lamont et al. 2019). In the southern flank across the southern tip of Africa (Fig. 1), it meets warmer and saltier waters carried by the swift Agulhas Current with large variability in the form of submesoscale eddies (Lass and Mohrholz 2008; Hutchings et al. 2009; Rubio et al. 2009; Bachèlery et al. 2016). In the northern flank, it is bordered by warm and nutrient-enriched waters transported by the poleward Angola Current (Shannon 1985; Hutchings et al. 2009; Bachèlery et al. 2016; Koseki et al. 2019).

Ocean dynamics controls the strength of the upwelling that affects the availability of nutrients, the primary production, and consequently the abundance and the diversity of pelagic fishes (Barlow et al. 2005; Schmidt and Eggert 2016; Lamont et al. 2019). Ekman transport driven by alongshore equatorward winds over the southwestern African coast pushes surface coastal waters offshore (Shannon 1985; Lass and Mohrholz 2008). Near the coast, a narrow-band of upwelling forms that brings nutrient-enriched water into the euphotic zone (Shannon 1985; Fennel 1999; Siegfried et al. 2019). This type of upwelling is

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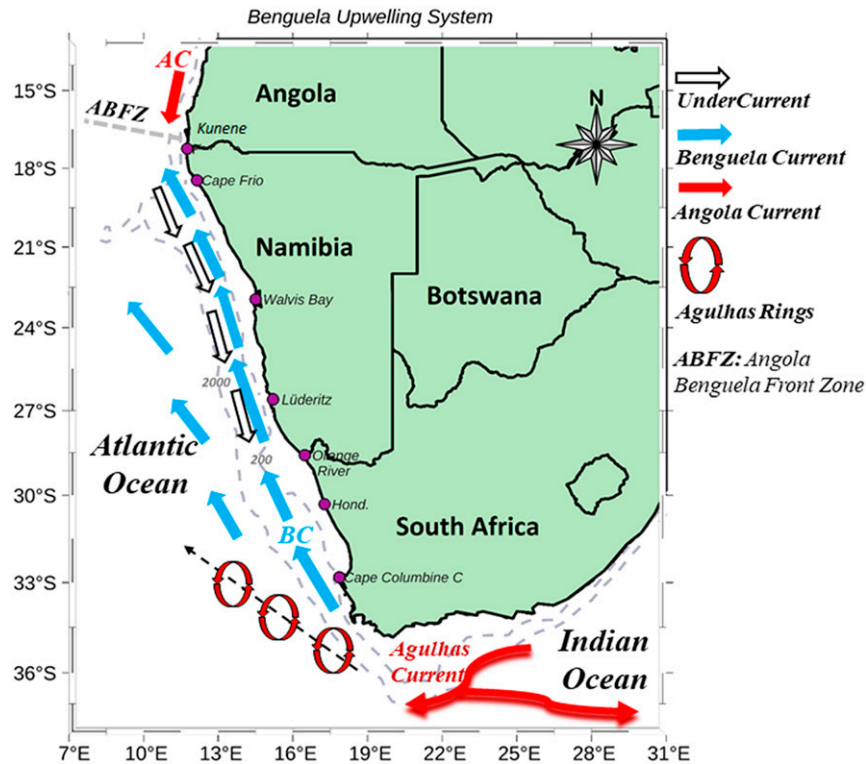


FIG. 1. Schematic of the mean surface and near-surface flow across the BUS. The blue northward arrows indicate the Benguela Current (BC). The red arrows on the top and bottom represent the Angola Current (AC) and Agulhas Current, respectively. Black arrows display southward undercurrents. Agulhas rings are depicted as red circles. The Angola-Benguela Frontal Zone (ABFZ) separates the BUS from the southward AC. Dashed gray contours are the 200- and 2000-m isobaths.

known as Ekman coastal upwelling. Its fast vertical velocity favors large-cell phytoplankton with high nutrient demands (Ryckaczewski et al. 2008; Lamont et al. 2019). Given the predator-prey body-size relationship, the zooplankton communities are, thus, driven toward large size species that primarily shape the aquatic food web (Ryckaczewski et al. 2008; Wasmund et al. 2014; Lamont et al. 2019). Farther offshore, the slow wind-stress-curl-driven (WSCD) upwelling supports a lower rate of nutrient flux to the euphotic zone (Shannon 1985; Fennel and Lass 2007; Mazzini and Barth 2013; Jacox et al. 2018). This favors small-size species with enhanced surface-to-volume ratio and high adaptability to nutrient-limited environments (Johnson 1976; Hense and Beckmann 2008; Ryckaczewski et al. 2008; Acevedo-Trejos et al. 2015). Consistently, the phytoplankton communities across the BUS feature a shift from the nearshore dominance of fast-growing species (e.g., diatoms) to more abundant small species (e.g., dinoflagellates) toward the open-ocean (Barlow et al. 2005, 2006; Wasmund et al. 2014; Louw et al. 2016). Recently, a seesaw behavior between the abundance of the small- and large-cell species across the BUS has been reported (Lamont et al. 2019). This might be related to the fact that the coastal upwelling and the WSCD upwelling are not necessarily in phase. In turn, they appear sometimes completely out of phase (Jacox et al. 2018). Such a phase relationship may drive

systematic transitions from larger species dominance to smaller species dominance or vice versa, which can alter the structure of the regional aquatic food web (Doney et al. 2012; Lamont et al. 2019). Hence, the relationship between the rapid coastal upwelling and the slow WSCD upwelling is of vital importance for the regional primary production.

Ekman coastal upwelling and that associated with the wind stress curl are not readily discernible and often spatially overlapping (Lamont et al. 2019). More importantly, they are not the exclusive drivers of the regional upwelling. Mesoscale and submesoscale ageostrophic processes (e.g., eddies, filaments, fronts, internal waves) play an important role in the time and space evolution of the BUS through horizontal and vertical advection, shaping the nutrient distribution and enhancing mixing (Lass and Mohrholz 2008; Veitch et al. 2010; Bettencourt et al. 2012; Muller et al. 2013; Veitch and Penven 2017; Lamont et al. 2019). For example, cyclonic submesoscale eddies intensify the upwelling and cause shoaling of the pycnocline/nutricline in the open-ocean waters. The opposite is true for anticyclonic eddies (Lamont et al. 2019). Further, a strong horizontal density contrast associated with coastal upwelling is formed at the base of the mixed layer depth that largely influences mixing and diabatic processes (de Szoeke and Richman 1984). Also, flow-topography interactions,

coastal geometry, and cross-shore geostrophic transport may alter the size and spatial pattern of the upwelling on different time scales (Mazzini and Barth 2013). Moreover, the development and propagation of coastal Kelvin waves arising from alongshore wind stress variability can entirely shut down the coastal upwelling whereas it has little impact on the offshore ocean dynamics (Fennel 1999; Fennel et al. 2012; Lübbecke et al. 2010, 2019).

In this study, we explore the relationship between the temporal evolution of the coastal and the WSCD upwelling. We also investigate the impacts of mesoscale and submesoscale ocean dynamics on these two types of upwelling and their relationship. To this end, we first examine the upwelling estimated by using the analytical method based on Ekman theory applied to observed surface wind data. We, additionally, analyze the output of a state-of-the-art forced ocean circulation model with sufficiently high resolution to simulate mesoscale dynamics and partially resolve submesoscale instabilities. This allows us to examine how the interplay among different upwelling drivers, which are excluded in the Ekman theory but captured by our numerical model (eddies, filaments, fronts, internal waves), alters the regional upwelling and where the Ekman theory can adequately represent the upwelling. Altogether, this approach is a trade-off between the complexity of a full analytical solution (Fennel 1999) and missing contribution to upwelling from processes not covered by the Ekman theory. With a particular focus on the phase difference, we assess the relationship between the time evolution of rapid coastal upwelling and slow WSCD upwelling in the Ekman theory together with the results of the high-resolution model.

Our paper is structured as follows. We first describe the observational data in section 2a. Section 2b presents the model and our experimental setup. The details of conceptual Ekman theory are described in 2c. The performance of the numerical model is evaluated by using different observational records in 3a. Long-term means of the simulated and analytical upwelling are presented in section 3b. The spatial pattern of variability over the entire BUS and the linkage between coastal and offshore (i.e., WSCD in the analytical theory and open-ocean in the model) upwelling are discussed in section 3c. The time series of coastal, analytical WSCD, and modeled open-ocean upwelling for several cross-shore transects are displayed and discussed in section 3d. We summarize our results in section 4.

## 2. Methods and materials

### a. Observational data

Among available surface wind products, we choose the observational wind at 10-m height above the sea surface  $U_{10}$  from ASCAT products, in which C-band scatterometer sensors were used to measure the ice-free ocean surface roughness (Ricciardulli and Wentz 2016). In fact, the surface roughness is a function of surface wind vectors. To retrieve the wind speed and direction from the ocean surface roughness, a numerical approach termed as Geophysical Model Function with a special focus on ASCAT records has been developed and used (Ricciardulli and Wentz 2016). The swath data are mapped on a  $0.25^\circ \times 0.25^\circ$  regular grid and are daily composites from data of three days. Our analysis covers the 11-yr period from

2008 to 2018. The climatological annual cycle of ASCAT wind vectors represents temporal and spatial patterns very consistent with other available observational wind products (Desbiolles et al. 2017). ASCAT mission is a running program with publicly accessible near-real-time data at the time of this study, which facilitates the real-time estimate of wind-driven upwelling. Such a fast evaluation of the upwelling state meets the needs of researchers in different disciplines of ocean science who might need a real-time estimate of upwelling when they are on the cruise. We assumed a neutral boundary layer and calculated the wind stress  $\tau$  from the daily ASCAT wind and using the commonly used bulk formula given by

$$\tau = \rho_a c_D \mathbf{U}_{10} \cdot U_{10}, \quad (1)$$

where  $\rho_a$  and  $c_D$  are surface air density and nondimensional drag coefficient, respectively, and are assumed to be constant at  $1.23 \text{ kg m}^{-3}$  and 0.0013, respectively (Gill 1982).

In addition, daily satellite-based chlorophyll-a (Chl-a) concentration and sea surface temperature (SST) datasets with 4km spatial resolution were taken from *Aqua* MODIS products (NASA Goddard Space Flight Center 2018). We also used the daily SST field from NOAA satellite OIv2 measurements (Reynolds et al. 2007). We mainly used the observed SSTs to validate the performance of the model. To this end, we compute the SST-based upwelling index over the BUS. This index is defined as the zonal SST contrast between coastal and open-ocean areas (Zhaoyun et al. 2012; Benazzouz et al. 2014) as follows:

$$\text{SST}_{\text{index}} = \text{SST}_{\text{coast}} - \overline{\text{SST}_{\text{open-ocean}}}, \quad (2)$$

where  $\text{SST}_{\text{open-ocean}}$  is computed as the average over 500–600 km offshore.

High Chl-a values ( $>30 \text{ mg m}^{-3}$ ) are very likely biased in postprocessing algorithms employing atmospheric corrections (e.g., cloud coverage, fog, and aerosol dust) or due to nearshore turbid water. As in similar studies (Thomas et al. 2012; Mohrholz et al. 2014), we considered the Chl-a values near the coast, with cross-shore distance less than 9 km, as well as suspiciously high values ( $>30 \text{ mg m}^{-3}$ ) as noise and flagged them as missing values.

To validate the performance of the numerical model, AVISO daily satellite altimetry date sets with  $0.25^\circ$  horizontal resolution for the 1993–2018 period (<https://cds.climate.copernicus.eu>) along with daily tide gauge measurements from the University of Hawaii Sea Level Center (<https://uhslc.soest.hawaii.edu>) are additionally used (Caldwell et al. 2015; Clementi et al. 2019). Among different satellite altimetry products, absolute dynamic topography (ADT), defined as the local deviation of sea surface height from the geoid, is chosen and analyzed. The deviation of observed ADT and simulated sea surface height from their areal average over the model domain are computed and termed as dynamic sea level (DSL; Greatbatch 1994; Bordbar et al. 2015). The main reason for choosing the DSL comes back to the fact that we considered Boussinesq approximation (i.e., conservation of volume rather than mass) in the model simulation. Thus, sea level displacement associated with changes in the density of the water column is not properly captured by the

model (Greatbatch 1994). With respect to the tide gauge records, time series of the research quality relative sea surface height from Simon's Town, Walvis Bay, Port Sonara, Takoradi, Ascension Island, and Macae stations for the available time period that matches the period of model simulation (i.e., 1992–2018) are analyzed.

We used the General Bathymetric Chart of the Oceans (GEBCO) data (<https://www.gebco.net>) in our numerical simulation (see Fig. S1 in the online supplemental material). The GEBCO seafloor bathymetry data were obtained from the aggregation of a number of internationally and nationally measured or estimated seafloor topography data by using Generic Mapping Tools routines. These seafloor topographic maps are mostly derived from direct measurements of single or multibeam echo sounders during regular regional surveys. GEBCO has a spatial resolution of 15 arc s (GEBCO Bathymetric Compilation Group 2019). These data show a shelf structure well comparable to echo soundings from cruises. Several seamounts and deep canyons on the Namibian shelf (Fig. S1) were identified as artifacts and were smoothed out using the ETOPO5 data as reference.

### b. Ocean-model simulation

We employed the Modular Ocean Circulation Model MOM-5, (Griffies 2012), forced by observational ocean–atmosphere momentum, heat and mass fluxes, and the river runoff. This model is a B-grid hydrostatic model and is run using the Boussinesq approximation in our integration. Prognostic variables are horizontal ocean velocity, sea level elevation, conservative temperature, and salinity. Vertical velocity is a diagnostic variable. Density and pressure in the water column are calculated from the TEOS-10 equation of state in hydrostatic approximation. For horizontal and vertical turbulent mixing, the Smagorinsky and the  $k$ -profile schemes were employed (Large et al. 1994). The latter includes the diagnosis of a mixed and a mixing layer depth. The vertical coordinates were  $z^*$  coordinate (Griffies 2012), which reduces numerical mixing from the undulating sea surface. The numerical integration was carried out with a 1200-s baroclinic time step; the barotropic split is 100 s.

To account for the interaction between the tropical and subtropical Atlantic Ocean and impacts of the South Atlantic subtropical gyre, Agulhas current, and southeast Atlantic central water mass, the model domain covers the Atlantic from 35°S to 12°N. Near the southwest African coast, which covers the BUS region, the resolution in the east–west direction is approximately 5 km and slowly stretched to 15 km westward. The latitudinal resolution is 5 km in the BUS and is extended to 8 km northward. The vertical resolution near the ocean surface is sufficiently high to capture internal wave dynamics on the shelf. It starts from about 3 m near the surface and reaches to about 10 m at 300-m depth but is further stretched below. Overall, the model consists of 118 vertical levels. This experimental setup is capable of adequately resolving the mesoscale and permits some submesoscale ocean dynamics (e.g., submesoscale eddies and filaments), equatorial waves, coastal trapped Kelvin waves, and associated mass and heat transports.

The time series of the main river discharges were obtained online (<https://hybam.obs-mip.fr/>). The ocean bathymetry is taken from GEBCO and is mapped onto the model grid by bilinear interpolation. The atmospheric momentum flux is estimated by using 6-hourly Cross-Calibrated Multi-Platform (CCMP2) data with  $0.25^\circ \times 0.25^\circ$  resolution (Atlas et al. 2011; Wentz et al. 2015). This dataset was obtained from a combination of several wind field products and covers a longer period than does ASCAT. Thus, forcing the model by CCMP2 data over a long period (i.e., 16 years) before the onset of ASCAT records enables the model to reach a statistically equilibrium state in 2008. CCMP2 wind products are not available as real-time data and are updated biannually. Across the model domain, it shows very close values to ASCAT that is the base of our analytical consideration over 2008–18. Atmosphere–ocean heat and freshwater fluxes (i.e., evaporation and precipitation) are estimated from ERA-Interim reanalysis with  $0.25^\circ \times 0.25^\circ$  horizontal resolution using the bulk formulas of Beljaars (1995). The prescribed quantities such as air pressure, temperature, and humidity were converted into the ocean model horizontal grid using bilinear interpolation. An exception is the wind field that was interpolated with the bicubic method to reduce artifacts in the wind stress curl pattern induced by interpolation. It should be here reminded that, unlike ASCAT data, CCMP2 includes wind field on lands. Such an interpolation likely leads to a smooth transition from the land to the nearshore wind field. All fluxes are calculated on the ocean model grid. The integration was carried out from 1992 to 2018. Here, we mainly analyzed the model results between 2008 and 2018 and took the vertical velocity at the base of mixed layer depth in the model output as the measure of upwelling intensity (McCreary 1981; Jacox et al. 2018).

The model employs open boundary conditions at the northern and southern boundaries. To incorporate the impacts of physical processes outside the domain, the potential temperature and salinity with 10-day time step and sea surface height with 12-h time step were taken from the ECCO (Estimating the Circulation and Climate of the Ocean) modelcube-92 (<https://www.ecco-group.org>). Since Agulhas rings are obviously underestimated in the model (see Fig. 2) but known to contribute significantly to the heat and mass transport over the very southern part of the BUS (Veitch et al. 2010; Veitch and Penven 2017), we conjecture possible impacts on the upwelling variability and take a very cautious interpretation of model results for this area.

We termed offshore upwelling in the model as open-ocean upwelling that is the equivalent of WSCD upwelling in the analytical theory. Similar to previous model studies (Veitch et al. 2010; Veitch and Penven 2017; Bonino et al. 2019), we use observed SST and the DSL to evaluate the performance of the numerical model.

### c. Coastal and offshore upwelling

According to the Ekman theory of wind-driven ocean circulation, ocean surface waters with homogenous density structure subjected to a persistent surface wind stress undergo a steady motion governed by a balance between the Coriolis force from their motion and wind-induced vertical flux

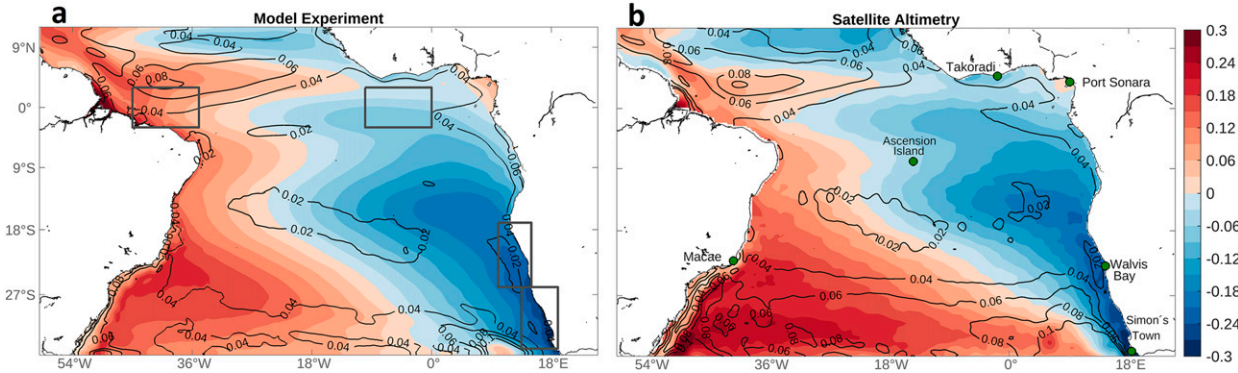


FIG. 2. Mean state (color shading; m) and variability (contours; m) of the DSL in the (a) model and (b) satellite altimetry data in the 1993–2018 period. To obtain the DSL, the sea surface height averaged over the entire model domain was subtracted in each grid point and time step. Variability is computed from the standard deviation of 5-day-averaged values. The outlined boxes in (a) show regions over which areal averages of the modeled and observed DSL are compared in Fig. S6 in the online supplemental material. Green circles in (b) indicate the locations of tide gauges utilized to validate the model results in Fig. S5 in the online supplemental material.

of horizontal momentum (Ekman 1905). The eastward ( $U_E$ ) and northward ( $V_E$ ) vertically integrated wind-driven volume transports per unit length are obtained as

$$U_E = \frac{\tau^y}{\rho_w f} \quad \text{and} \quad V_E = \frac{-\tau^x}{\rho_w f}, \quad (3)$$

where  $\rho_w$  and  $f$  are seawater density and the Coriolis parameter, respectively;  $\tau^x$  and  $\tau^y$  are zonal and meridional wind stress, respectively, and  $f$  is negative on the Southern Hemisphere.

Southwest African coasts are overall meridionally oriented and feature a dominant equatorward surface atmospheric flow, which is persistent throughout the year and causes offshore transport (Hart and Currie 1960; Shannon 1985; Small et al. 2015). Given the cross-shore divergence of the transport near the coast, the displaced water is replaced by the convergence of alongshore transport or by upwelled water into the Ekman layer (Mazinni and Barth 2013; Muller et al. 2014; Jacox et al. 2018). It is important to note that the Ekman balance is disturbed near the coast (see the online supplemental material). Because of the existence of the coast, the offshore transport and associated divergence are compensated by vertically upwelled transport, when the alongshore current is spatially uniform. This compensation of the divergence of the seaward flow by the vertical flow occurs across an offshore distance of about the first baroclinic Rossby radius of deformation  $R$ . (McCreary 1981; Fennel 1988, 1999; Mazinni and Barth 2013; Jacox et al. 2018). Given the fact that the Ekman balance is disturbed right at the coast, an additional surface transport emerges that is in the opposite direction to the Ekman transport and guarantees the coastal boundary condition [ $U(x=0) = 0$ ; see the online supplemental material]. The small component from the barotropic mode and variations of  $R$  from density changes due to upwelling (de Szoeke and Richman 1984) is not considered here.

Independent of the presence of the coast, a spatially heterogeneous wind field leads to a divergence or convergence of the Ekman transport that additionally forces water in the upper ocean to lift upward to the surface or pump downward to the deep ocean (Johnson 1976; Pickett and Paduan 2003; Junker et

al. 2015). In the  $f$ -plane approximation (i.e.,  $f$  is assumed to be invariant with latitude), this type of upwelling, referred to as Ekman pumping or WSCD upwelling ( $w_{\text{curl}}$ ), reads

$$w_{\text{curl}} = \frac{1}{\rho_w f} \left( \frac{\partial \tau^y}{\partial x} - \frac{\partial \tau^x}{\partial y} \right). \quad (4)$$

To compute the upwelling related to the coastal divergence ( $w_c$ ), we first compute the Ekman transport associated with surface wind field in each ocean grid point. To account for the effects of coastal orientation on the cross-shore transport (Fig. S2 in the online supplemental material), we compute the inner product between the vector of coastal Ekman transport and the line orthogonal to the local coastline, which is termed the cross-shore Ekman transport ( $U_{\text{cross-shore}}$ ; Pickett and Paduan 2003). One should keep in mind that near the coast ( $x = 0$ ) the Ekman balance is disturbed (see the online supplemental material). In theory, if the meridional component of wind is considered as alongshore wind,  $\tau^y_{x=0}/\rho_w f$  gives the integrated coastal divergence of Ekman transport.

We utilized the observational-based  $R$  with  $1^\circ \times 1^\circ$  resolution computed by Chelton et al. (1998). It is particularly large over the northern sector of the BUS where the stratification is strong and  $f$  (i.e., the Coriolis force) is relatively small (Fig. S3 in the online supplemental material). We regrid these data with respect to ASCAT data using the bilinear interpolation. If the Rossby radius of deformation near the coast was missing, the value from the nearest grid point with the same latitude was used. The grid sizes are  $0.25^\circ$  and are smaller than  $R$ .

To compute the coastal upwelling from ASCAT data, we assume that, first, the total offshore transport is compensated in a coastal distance equal to  $R$  and, second, the vertical velocity decreases exponentially with cross-shore distance (Fennel 1999; Jacox et al. 2018). Given the course resolution of ASCAT data, we assume the vertical water velocity in the coastal distance of  $R$  decreases to 10% of the velocity near the coast. Thus, the velocity in the coastal distance smaller than  $R$  is obtained using following formula [supplemental material Eq. (S23)]:

$$w_c = \frac{2.07 \times U_{\text{cross-shore}}}{R_1} e^{2.3026(x/R_1)}, \quad (5)$$

in which  $x$  and  $U_{\text{cross-shore}}$  denote the coastal distance and cross-shore Ekman transport, respectively. Note that  $x$  is considered to be negative in the east-to-west direction. At offshore distances larger than  $R$ , the vertical velocity associated with coastal Ekman transport becomes zero. This estimate is in a good agreement with theory of Benguela upwelling developed by Fennel (1999) (see the online supplemental material).

To compute the WSCD upwelling, we first calculate the wind stress curl. If a grid point on land was involved, it was set to missing data. Thus, we excluded the nearest grid points to the coast and computed the Ekman pumping elsewhere over the ocean.

The values of  $w_{\text{curl}}$  and  $w_c$  were merged to a single data set and termed as “analytical theory.” Thus, within the coastal distance between  $0.25^\circ$  and  $R$ , the WSCD and coastal upwelling are overlapping in the analytical approach. In theory, integrated coastal upwelling ( $W$ ; Table S1 in the online supplemental material) is included both WSCD upwelling and the integrated coastal divergence of Ekman transport. However, the contribution of WSCD upwelling in the nearest grid cell to the coast is neglected here. The volume of upwelled water in each grid cell is estimated as the product of upwelling velocity and gridcell area. A summary of different upwelling related variables is given in Table S1.

### 3. Results and discussion

In this section, we first compare the numerical model results with observational records to validate model outputs. Thereafter, we assess the narrow band rapid coastal upwelling and slow offshore upwelling along the southwestern coast of Africa (Fig. 1) with special interest in their temporal relationship. In this context, we examine the spatial structure of the mean state and variability of the upwelling together with its temporal evolution in the model experiment and the analytical theory.

#### a. Model validation

We first compare the mean state and variability of the modeled and the observed DSL for the 1993–2018 period (Fig. 2). It stands for the large-scale circulation in the central water masses. The variability is estimated from the standard deviation (Std) of 5-day mean values. In general, the spatial pattern of the mean-state ocean topography, shown by the DSL, is well represented by the model (Fig. 2). Important aspects of the large-scale ocean circulation such as zonal asymmetry across tropical Atlantic, which is associated with equatorial currents and equatorial ocean dynamics, the zonal contrast across the South Atlantic, which represents the strength and the structure of the South Atlantic subtropical gyre, and relatively low sea level across the southwest African coast are well reproduced by the model. The amplitude of the zonal contrast across the southern sector is slightly underestimated by the model, which means the subtropical gyre might be weaker in the model. In addition, the eastward intrusion of the high DSL over the southern sector is smaller in

the model. The main difference is the high ocean topography over the southwest Atlantic, which is weaker and located farther north in the model. With respect to the spatial pattern of the variability, the model result compares fairly well to the satellite-derived data. Large DSL variability over the western and eastern tropical Atlantic is well captured by the model. The spatial pattern of modeled and observed variability over the southern sector shares many similarities. However, the level of variability in this part is smaller in the model. The variability of sea level elevation near the southern boundary of the model domain with Agulhas current and associated rings is not well represented. Due to the numerical properties of the open boundary conditions, water from the Indian Ocean enters the model domain too far west and stays mostly away from the African coast. Relative to the observation, the simulated cold upwelled water tends to extend farther seaward off the Cape Columbine cell (Fig. S4 in the online supplemental material). This may have some connections to the reduced Agulhas rings in the model. This limitation makes the model simulation less reliable for the southernmost part of the BUS. For the rest of the domain, the general pattern of SST mean state and variability in the model and observation compare fairly well (Fig. S4). Remarkably, the large meridional SST gradient over the ABFZ is captured well by the model.

We also examine the time evolution of simulated DSL (Figs. S5 and S6 in the online supplemental material). To this end, we first compare the modeled DSL with several locally measured tide gauge records (Fig. S5). We also compare the time series of 5-day mean modeled and observed DSL across several regions (Fig. S6). Overall, the model simulation and tide gauge measurements are significantly correlated in all selected stations. The correlation between the model and tide gauge records in Walvis Bay and Simon’s Port located within the BUS is about 0.59 and 0.47, respectively (Fig. S5). The areal average of simulated and observed DSL over the western equatorial Atlantic, Atlantic Niño region, northern and southern BUS are evolved very closely (Fig. S6). Overall, the model simulation can explain a considerable part of the observed sea level variability.

We additionally compare the spatial pattern of the mean state and variability of the SST-based upwelling index (see section 2) in the model and observation (Fig. S7 in the online supplemental material). The modeled index is based on water temperature at 1.5-m depth, which is the shallowest vertical level of the model, whereas it is based on SST in the observational estimates. Despite the fact that the observed SST is more affected by local atmospheric forcing (i.e., radiation and evaporative cooling) the spatial pattern of the observed and modeled mean and variability compare relatively well (Fig. S7). Remarkable in this respect is the difference between the results of *Aqua* MODIS and Reynolds SST, which suggests a large observational uncertainty. The time series of modeled and observed SST-based index over the northern and southern BUS zones evolved very closely and represent significant correlations (Fig. S7).

#### b. The upwelling mean state

Figure 3 compares the spatial pattern of the annual mean (2008–18) vertical water velocity, as a measure for the strength

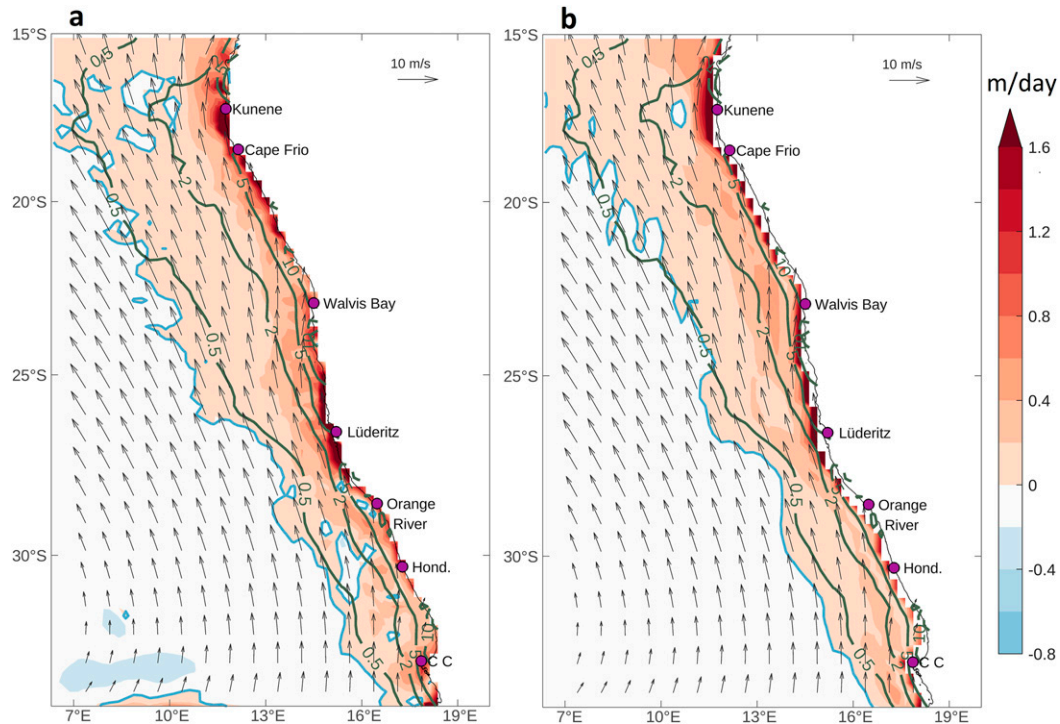


FIG. 3. Long-term mean of ocean vertical velocity (shading;  $\text{m day}^{-1}$ ) from (a) the three-dimensional ocean model and (b) the analytical theory. The surface wind field (black arrows;  $\text{m s}^{-1}$ ) and chlorophyll-a concentration (dark-green contours;  $\text{mg m}^{-3}$ ) are shown. The simulated velocity in (a) represents the vertical velocity at the base of mixed layer depth, whereas the analytical vertical velocity in (b) is estimated from ASCAT wind field records (see section 2). In (a) and (b), chlorophyll-a concentration is derived from Modis *Aqua* satellite measurements. The blue contours indicate where the vertical velocity is equal to zero.

of upwelling, obtained from the numerical model and the analytical model. The long-time mean of remotely measured Chl-a concentration and surface wind fields are overlaid. In general, the simulated and analytical vertical velocities are broadly consistent in which coastal upwelling is generally intense over the entire domain and characterized by several distinct localized upwelling cells centered close to Kunene, Cape Frio, Walvis Bay, Lüderitz, and Cape Columbine. Kunene and Lüderitz coastal cells show more intense upwelling when compared with other cells. This is in good agreement with previous studies (Lutjeharms and Meeuwis 1987; Nicholson 2010; Small et al. 2015; Jury 2017) and most likely due to the orography and the large localized land–sea temperature contrast, which can intensify the wind stress locally. Despite the fact that the analytical consideration neglects several processes such as propagation of coastally trapped Kelvin waves, topographic and coastal geometry impacts, and cross-shore geostrophic flow, coastal upwellings in the model and analytical theory have very similar structures. For the Kunene cell, the local subduction of Angola Current over the Angola–Benguela front can significantly reduce the upwelling locally, which is not considered in the analytical theory. Overall, this result illustrates the dominant role of the coastal Ekman dynamics along the southwestern coast of Africa, which can overshadow the impacts of other contributors on sufficiently long-term average. An exception is

the Kunene upwelling cell, where the analytical approach gives much stronger coastal upwelling than the numerical model. This is partly due to poleward-directed Kelvin waves that tend to reduce coastal upwelling (Fennel 1988, 1999). These Kelvin waves arising from the strong meridional gradients in the meridional wind field that are neglected in our analytical analysis.

An equatorward broadening of the open-ocean upwelling is an integral feature of the BUS (Fennel et al. 2012) that is well reproduced by the model and analytical theory. Analogously, the phytoplankton-rich belt (i.e., Chl-a concentration  $>0.5 \text{ mg m}^{-3}$ ) along the coast widens up equatorward. This provides a coherent signature with the structure of the wind-driven upwelling. It also features two relatively wide bands centered near Orange River and slightly north of Walvis Bay where the shelf widens (Fig. S1 in the online supplemental material). It is worth noting that phytoplankton biomass is also influenced by solar radiation, turbulent mixing, and nutrient flux (Lamont et al. 2019). Thus, such a consistent spatial pattern underlines the profound impacts of physical upwelling that provides the feeding opportunity for plankton communities. However, the offshore extension of the upwelling region in the analytical theory and the numerical model are slightly different. The offshore upwelling center between Lüderitz and Walvis Bay is slightly wider and stronger in the numerical model with reference to the analytical theory. This difference might be

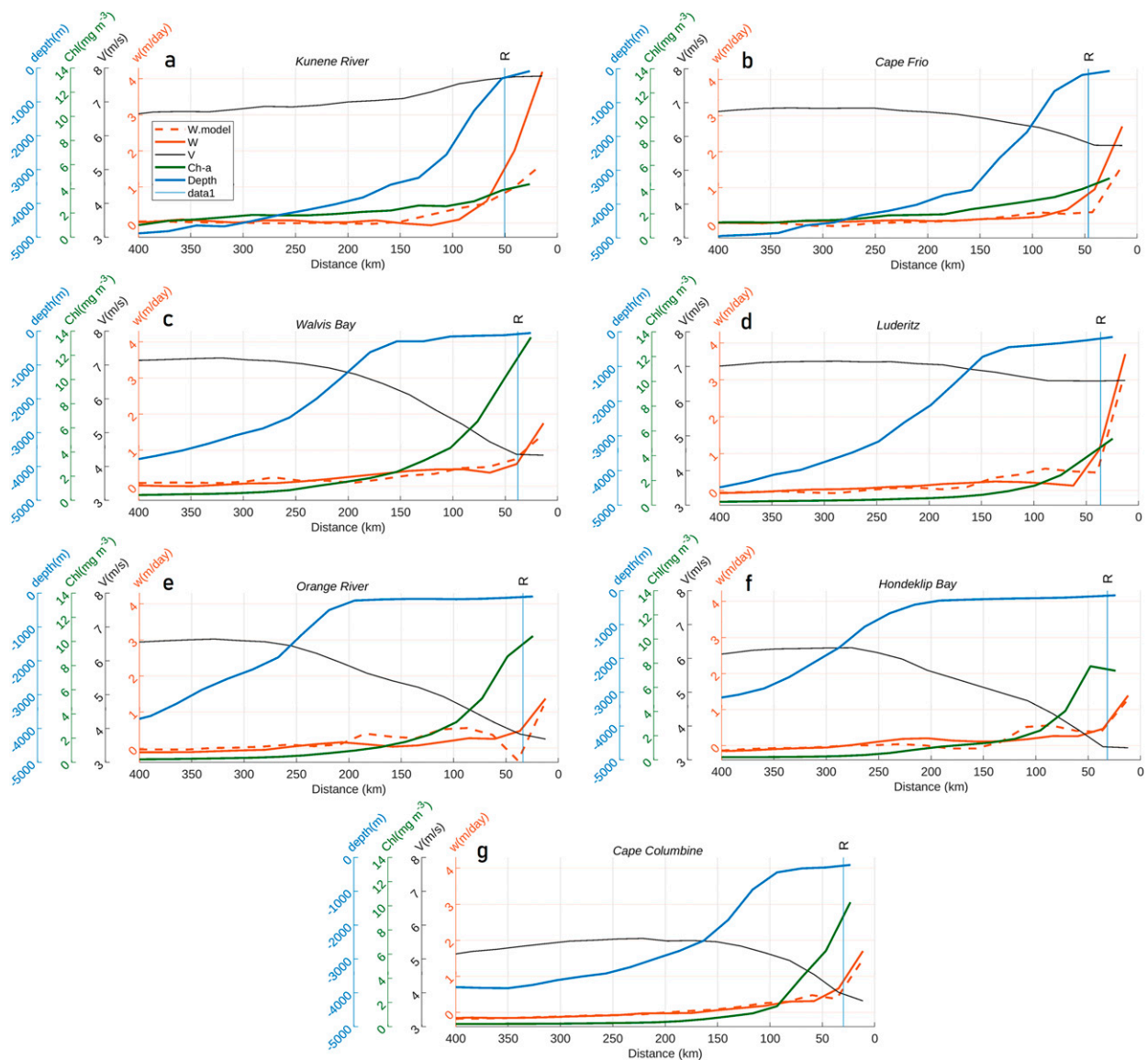


FIG. 4. Profiles of the long-term mean chlorophyll-a concentration (dark-green solid line;  $mg m^{-3}$ ), ocean depth (blue solid line;  $m$ ), simulated (orange dashed line) and analytical (orange solid line) vertical velocity ( $m day^{-1}$ ; see section 2), and meridional wind velocity (black line;  $m s^{-1}$ ) vs cross-shore distance for seven cross-shore transects: (a) Kunene, (b) Cape Frio, (c) Walvis Bay, (d) Lüderitz, (e) Orange River, (f) Hondeklip Bay, and (g) Cape Columbine (see Fig. 1). Note that the blue vertical line on the right side of each panel represents the first baroclinic Rossby radius of deformation. The simulated velocity (orange dashed line) represents the vertical velocity at the base of the mixed layer depth (see section 2). Each variable and corresponding y axis are shown with identical colors.

attributed to mesoscale and submesoscale ocean processes across the region impacting the vertical water transport (Bettencourt et al. 2012). However, this result reflects the prevailing impact of Ekman dynamics relative to other drivers of the upwelling over the long-term average across the southwest African coast.

Over the coastal distance of  $R$ , the coastal and the WSCD upwelling overlap. It is important once again to bear in mind that the wind stress curl in the nearest grid cell to the coast was set to the missing value in our analytical theory. Hence, the contribution of the coastal and WSCD upwelling can be

quantified over limited area north of Walvis Bay (Fig. S8 in the online supplemental material). North of Walvis Bay and near the northernmost sector of the BUS, a significant proportion of the nearshore upwelling is associated with the WSCD upwelling (Fig. S8). Given the horizontal resolution of ASCAT data in comparison with the local  $R$ , a reliable estimate of this contribution with adequately high spatial resolution remains a challenge.

Figure 4 displays the annual average of vertical water velocity from the model and the analytical theory, the observed Chl-a concentration, meridional surface wind velocity together



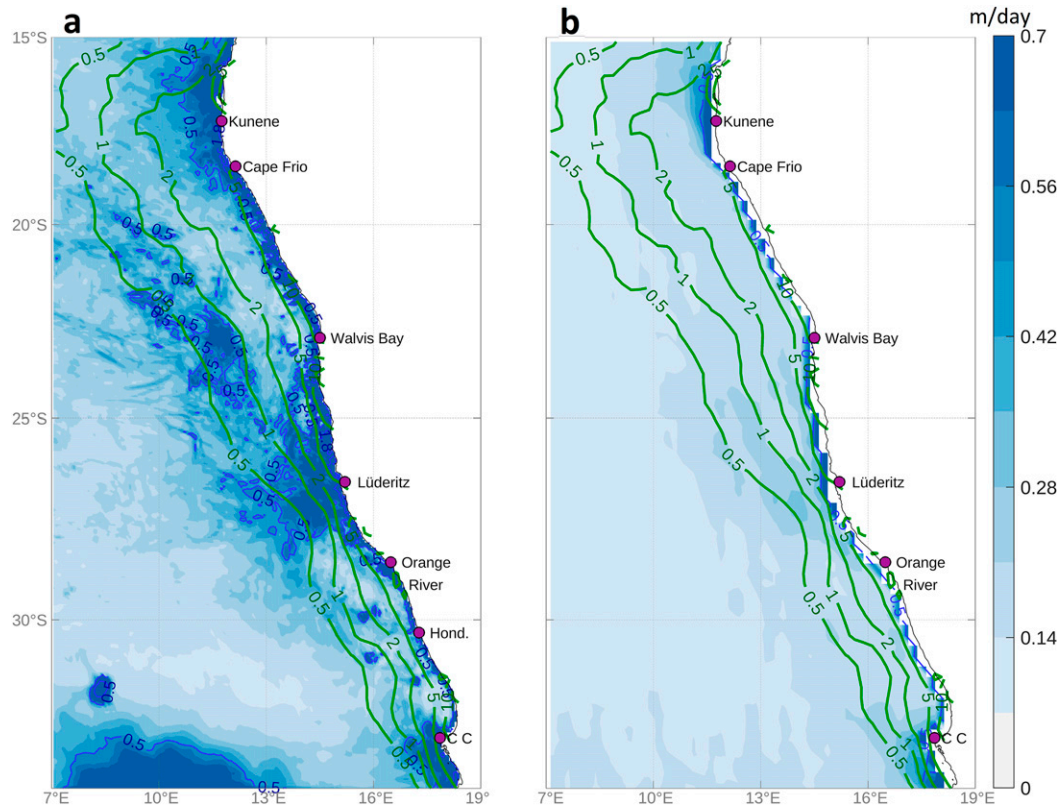


FIG. 5. Variability of (a) simulated and (b) analytical water vertical velocity ( $\text{m day}^{-1}$ ) over the 2008–18 period. In (a) and (b), variability is estimated from the Std of monthly mean values. Green contours represent the mean of Chl-a concentration ( $\text{mg m}^{-3}$ ). The simulated vertical velocity is taken at the base of the mixed layer depth.

with ocean depth in seven cross-shore transects. Sign and magnitude of the wind stress curl across the BUS are determined to a great extent by the zonal variation of the meridional wind velocity (Fennel et al. 2012). The coastal drop-off in the meridional wind velocity is the major contribution to the wind stress curl. It occurs over narrower bands in the southern BUS (Hondeklip Bay and Cape Columbine) but extends much farther offshore in the northern BUS (Cape Frio and Walvis Bay). A particular exception to that is found off the Kunene mouth, where the wind starts to decline near the coast (Fig. 4a). Despite the simplicity of the analytical approach, its outcome shares many similarities with the numerical model results. This reflects the dominant role of Ekman dynamics across the entire BUS on the long-time average. In general, the analytical theory renders larger values near the coast relative to the model output. This difference is more pronounced in the Kunene, Cape Frio, and Orange River cells. This overvaluation is potentially due to the spatially coarse resolution of the ASCAT observed wind field ( $0.25^\circ \times 25^\circ$ ), which prohibits an adequately fine-resolution estimate of coastal upwelling. As pointed out by Lamont et al. (2019), a steep inshore-offshore gradient in Chl-a concentration, associated with the seaward reduction of the vertical water velocity, is revealed in all transects. The nearshore Chl-a concentration in Walvis Bay, Orange River, and Cape Columbine, where the continental

shelf is relatively broad, appears to be large in comparison with other transects.

### c. The upwelling variability

To get an overview of the upwelling variability and to further check the consistency between the model and our analytical approach, we next examine the Std of the simulated and analytical monthly mean vertical water velocities as a measure for the variability (Fig. 5). Unlike the spatial pattern of the long-term mean (Fig. 3), the spatial maps of variability computed from the model and analytical theory show remarkable differences. The model offshore variability across the entire domain is nearly one order of magnitude larger than that derived from the analytical approach (Figs. 5a,b). The presence of submesoscale filaments and eddies across the region could be very likely behind this difference (Bettencourt et al. 2012). In fact, submesoscale dynamics is known to exert profound influences on the near-surface vertical velocity, and associated upward nutrition flux into the euphotic layer, which is not taken into account in the analytical estimate.

Offshore waters near Walvis Bay and Lüderitz and across the southwestern BUS appear to have high variability in the model. Giving the uncertainty associated with boundary values near the southern sector, it is reasonable to place a caveat on the interpretation of model outputs very close to the southern

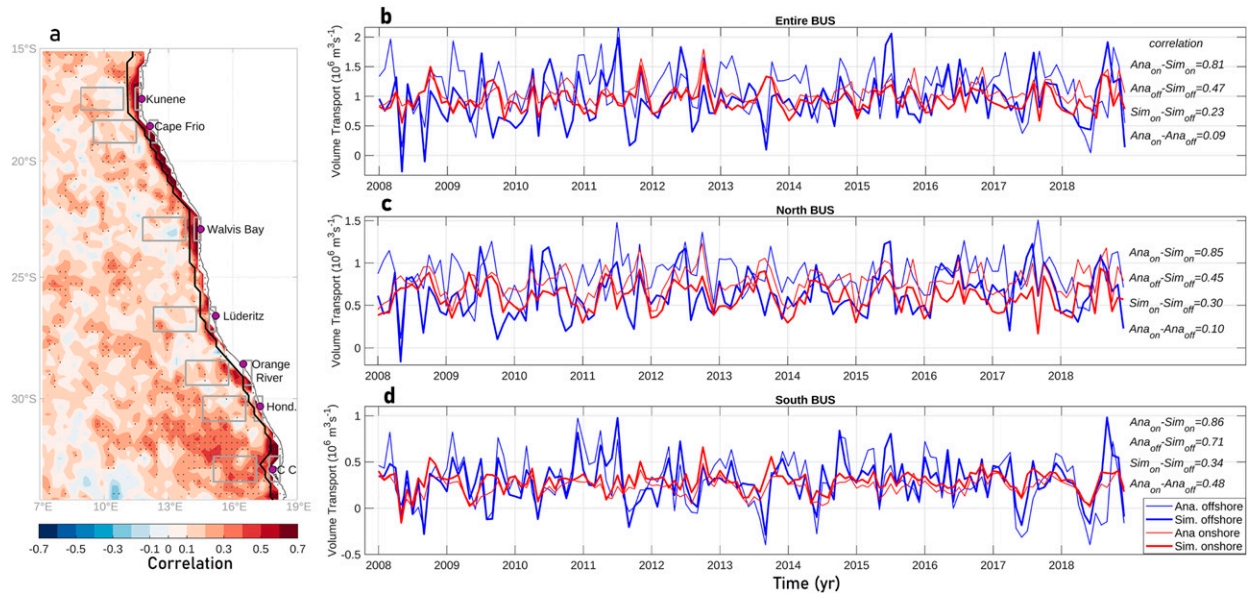


FIG. 6. (a) Spatial pattern of the linear correlation coefficients between the simulated and the analytical monthly mean vertical water velocities over the 2008–18 period. Also shown are time series of simulated (thick lines) and analytical (thin lines) coastal (red lines) and offshore (blue lines) upwelled waters across (b) the entire BUS and the (c) northern ( $26^{\circ}$ – $17^{\circ}$ S) and (d) southern ( $34^{\circ}$ – $26^{\circ}$ S) BUS subsystems. The coastal upwelled water is the total upwelled water over the coastal distance of  $R$ , which is shown by the black contour in (a). The offshore upwelled water is the total amount of upwelled water between  $R$  and 250-km offshore distance. On the right side of (b)–(d), the correlations between different time series are displayed. Note that the climatological annual cycle was not subtracted from data. The volume of upwelled water over the outlined boxes in (a) is computed and is shown in Figs. 8 and 9, below. Stippled areas in (a) show where the correlation is statistically significant at the 95% confidence interval.

part of the domain. Large variability in the proximity of the Lüderitz cell might be connected to the year-round occurrence of strong filaments with a large vertical extent and accompanied with the perennial alongshore wind (Hösen et al. 2016; Veitch and Penven 2017). Consistent with the model result, a large vertical displacement of thermocline associated with seaward propagation of submesoscale filaments, eddies, and Rossby waves near Walvis Bay was previously reported by Mohrholz et al. (2008).

In the following section, we investigate the temporal evolution of the simulated and the analytical upwelling in more detail and examine to what extent Ekman dynamics can estimate the amount of upwelled waters across the southern and northern subsystems. We begin with the linear correlation between simulated and analytical upwelling in each grid point (Fig. 6a). In general, the correlation peaks near the coast and is typically larger than 0.6 everywhere nearshore; implying that the simulated ( $w_m$ ) and analytical coastal upwelling ( $w_c$ ) evolved very closely. The large agreement between model and analytical theory near the coast further reflects the fundamental role of the coastal Ekman drift and associated upwelling near the coast. With respect to the open-ocean upwelling that is exclusively WSCD in the analytical approach (i.e.,  $w_{curl}$ ), except over the southern sector of the BUS near Cape Columbine, the correlation between the simulated and analytical upwelling is mostly weak and not statistically significant (Fig. 6a). Since the variability associated with the Agulhas rings seems to be underestimated in the model, it is difficult to make

any definite statement about the relatively high correlation over the southern part of the domain. Near Lüderitz and Walvis Bay, where a large offshore variability is simulated (Fig. 5a), the correlation is not statistically significant. This result suggests that the upwelling associated with wind stress curl is largely disrupted by the local mesoscale and submesoscale dynamics.

However, the time series of the simulated ( $V_{m-offshore}$ ) and the analytical ( $V_{curl}$ ) vertical volume transport offshore across the entire domain behave closely with the correlation of about 0.47 (Fig. 6b). Despite the interruptive effects of mesoscale and submesoscale dynamics on small scales, its impacts over the entire domain seem to be small and the basinwide vertical water transport estimated by Ekman theory is still consistent with model results. This close time evolution is also found in the upwelling across the southern and the northern subsystems (Figs. 6c,d), which render statistically significant correlations of 0.71 and 0.45, respectively. The relatively weak correlation in the northern subsystem can be partly explained by the large local submesoscale activities (Fig. 5a). Given the mesoscale corridors associated with Agulhas leakage near the southern flank, one expects the disruptive role of submesoscale dynamics over the southern BUS to be larger relative to that in the north, but the results of the numerical model reveal otherwise.

Further, the amount of upwelled water from the analytical consideration appears to be often larger than simulated upwelling over the northern BUS. This is consistent with previous

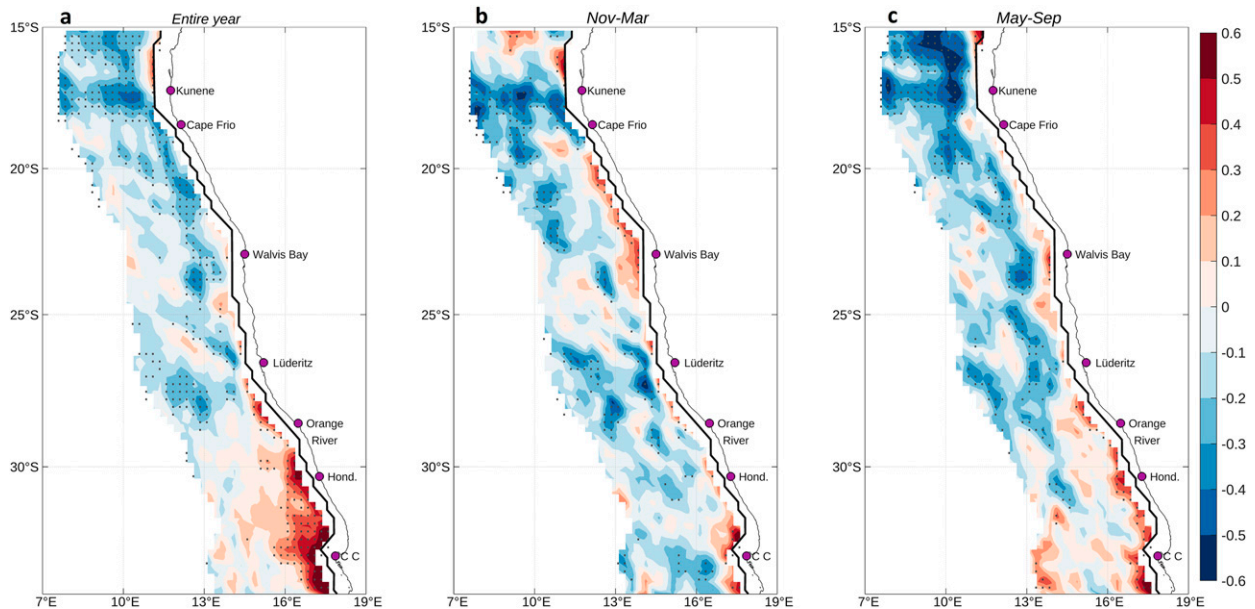


FIG. 7. Spatial pattern of the correlation coefficient between coastal ( $w_c$ ) and WSCD ( $w_{curr}$ ) upwelling obtained from analytical theory over 2008–18, showing (a) the correlation between the monthly mean values for all calendar months and the correlation for (b) November–March and (c) May–September. At each grid point, the correlation between time series of local WSCD and the coastal upwelling at the same latitude is shown. The black contours represent the coastal distance equal to  $R$ . Note that in (a)–(c) the climatological annual cycle was retained. Stippled areas show where the correlation is statistically significant at 95% confidence interval.

model studies noting that the amount of upwelled water across the eastern boundary upwelling systems is, in general, reduced when mesoscale and submesoscale activities are represented in the model (Capet et al. 2008). This effect can be attributed to the eddy-induced subductions in the upper-oceanic fronts.

#### d. Time evolution of the coastal and the open-ocean upwelling

This section is devoted to the time evolution of the coastal upwelling and the open-ocean upwelling. The latter is exclusively WSCD upwelling in the analytical approach. With a particular interest in the phase difference, we aim to determine whether these components of upwelling evolve coherently or alternating with each other. Here, we recall that local processes such as the thermal coupling between ocean and atmosphere, feedbacks between the crosswind SST gradient and the wind stress curl, orographic effects, and the coastal geometry play significant roles in these relationships (Desbiolles et al. 2014). Figure 7a displays the linear correlation coefficient between coastal (i.e.,  $w_c$ ) and WSCD (i.e.,  $w_{curr}$ ) upwelling being at the same latitude in the analytical theory.

The coastal and the WSCD upwelling are mostly anti-correlated north of Lüderitz. This feature is more obvious and statistically significant near the Kunene cell. Theoretically, it indicates that a weakening of the coastal upwelling is often concurrent with an intensification of the WSCD upwelling and vice versa. From the south of Cape Frio to the north of Lüderitz, the area with the statistically significant anti-correlation is limited to a relatively narrow band near the coast. Such an inverse relationship gradually disappears nearly

south of Lüderitz and the correlation becomes positive and statistically significant from the north of Hondeklip Bay to the south of Cape Columbine, which suggests the coastal and the WSCD upwelling often evolve nearly in phase. Part of this behavior is very likely connected to the seasonal cycle of the wind field. Concurrent with seasonally strong wind stress curl from October to April over the entire domain, the alongshore component of wind is stronger in the southern BUS but weaker in the far north (Figs. S9 and S10 in the online supplemental material). The correlation between upwelling anomalies yields the same structure but with a smaller magnitude (Fig. S11 in the online supplemental material). In particular, the reduction in the size of correlation is more pronounced off Cape Columbine and Hondeklip Bay cells where the seasonal variation of the wind is relatively strong (Fig. S9). Overall, this result suggests that the annual cycle has a contrasting impact on the relationship between coastal and WSCD upwelling in the northern and the southern subsystems.

During different seasons, the spatial structure of the correlation and, thus, the phase difference might be different. To this end, we computed the aforementioned correlation for the periods corresponding to November–March and May–September (Figs. 7b,c). The general structure does not change much. Note that the statistical degrees of freedom in Figs. 7b and 7c is different from Fig. 7a; meaning the same level of correlation does not necessarily indicate the same level of statistical confidence. From visual inspection, the magnitude of correlation coefficient near the Kunene cell appears to be larger between May and September, which is concurrent with the local intensification of the alongshore winds and weakening

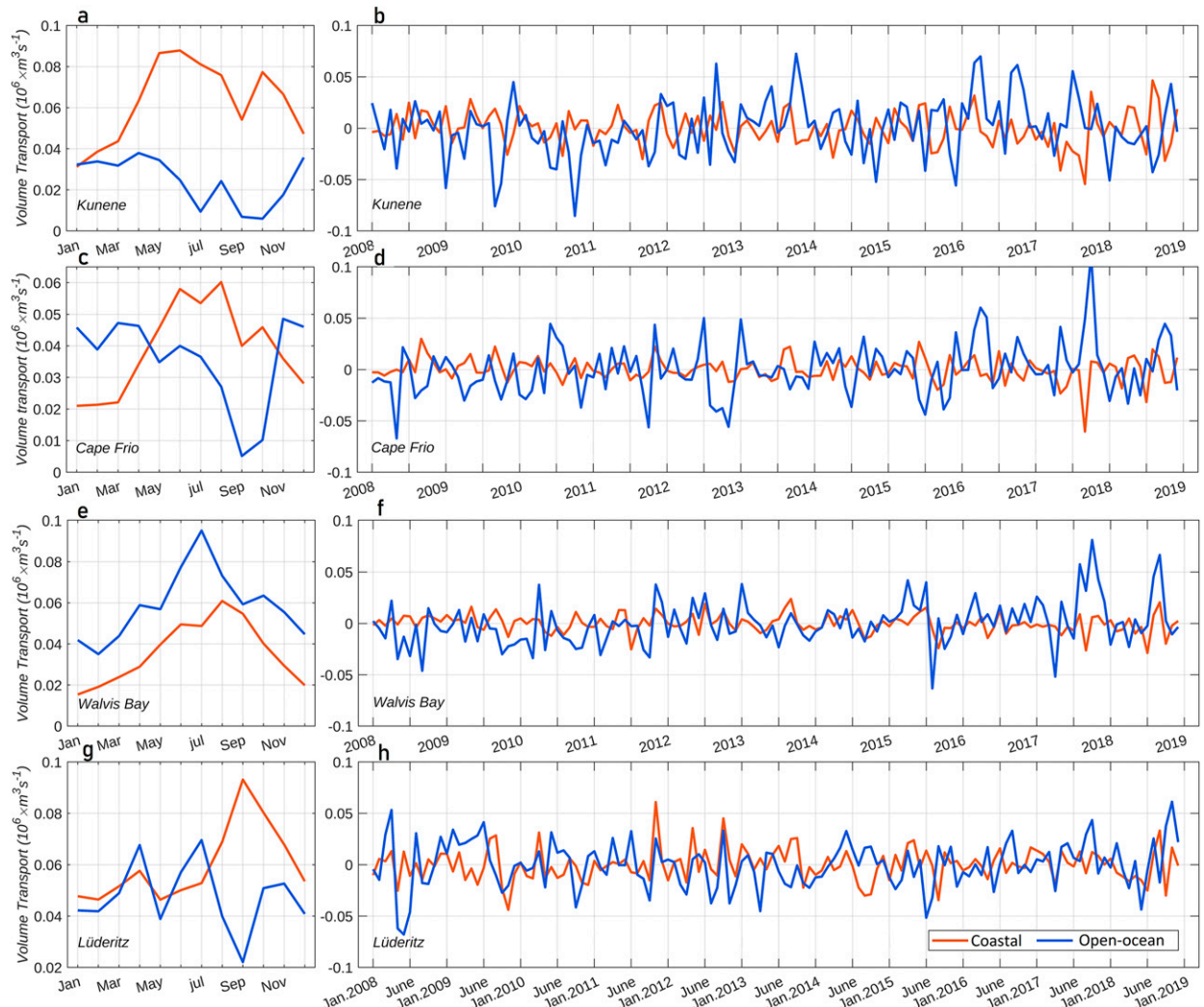


FIG. 8. (left) Climatological annual cycle and (right) time series of coastal (orange line) and offshore (blue line) amount of upwelled water anomaly ( $10^6 \text{ m}^3 \text{ s}^{-1}$ ) computed over coastal and offshore rectangular boxes near (a),(b) Kunene; (c),(d) Cape Frio; (e),(f) Walvis Bay; and (g),(h) Lüderitz. The rectangular boxes are outlined in Fig. 6a. Note that the y axes in (a), (c), (e), and (g) have different scales. In (b), (d), (f), and (h), the climatological annual cycle was subtracted in each time series.

of the wind stress curl (supplemental Figs. S9 and S10). Thus, the seesaw behavior is likely to be stronger between May and September.

The spatial pattern of the correlation in the model is very noisy (Fig. S12 in the online supplemental material). Apparently, the correlation pattern is strongly disrupted by submesoscale ocean dynamics. The inevitable question that arises is whether the mesoscale and submesoscale dynamics remarkably alter the relationship between the amount of coastal ( $w_c$ ) and WSCD upwelled water. In this way, we discuss here the time series of coastal ( $V_{m\text{-coast}}$ ) and open-ocean ( $V_{m\text{-offshore}}$ ) volume of upwelled water and their climatological annual cycle computed over seven subregions of the BUS (Fig. 8 and 9). These regions are shown by rectangular boxes in Fig. 6a.

The seasonal cycle of coastal upwelling in Kunene is characterized by a biannual cycle. The coastal upwelling peaks in

June and October, whereas the open-ocean upwelling is identified by a mirror-like pattern with its minimum in July and October (Fig. 8a). This behavior is consistent with the annual cycle of the alongshore wind and the wind stress curl (Figs. S9 and S10 in the online supplemental material). Fifty-seven percent of the coastal upwelling variability can be explained by the annual cycle (see Table 1). This feature is less pronounced for the open-ocean upwelling (14%; Table 1). This might be connected to the dominant role of the Ekman transport in the coastal upwelling whereas any changes in the strength and direction of coastal or offshore winds can significantly alter the size of WSCD upwelling. Similar to the annual cycles (Fig. 8a), coastal and open-ocean upwelling anomalies off Kunene frequently appear in opposite phases, with correlation of about  $-0.24$  (Fig. 8b). In fact, positive and negative peaks in the coastal and the open-ocean time series frequently

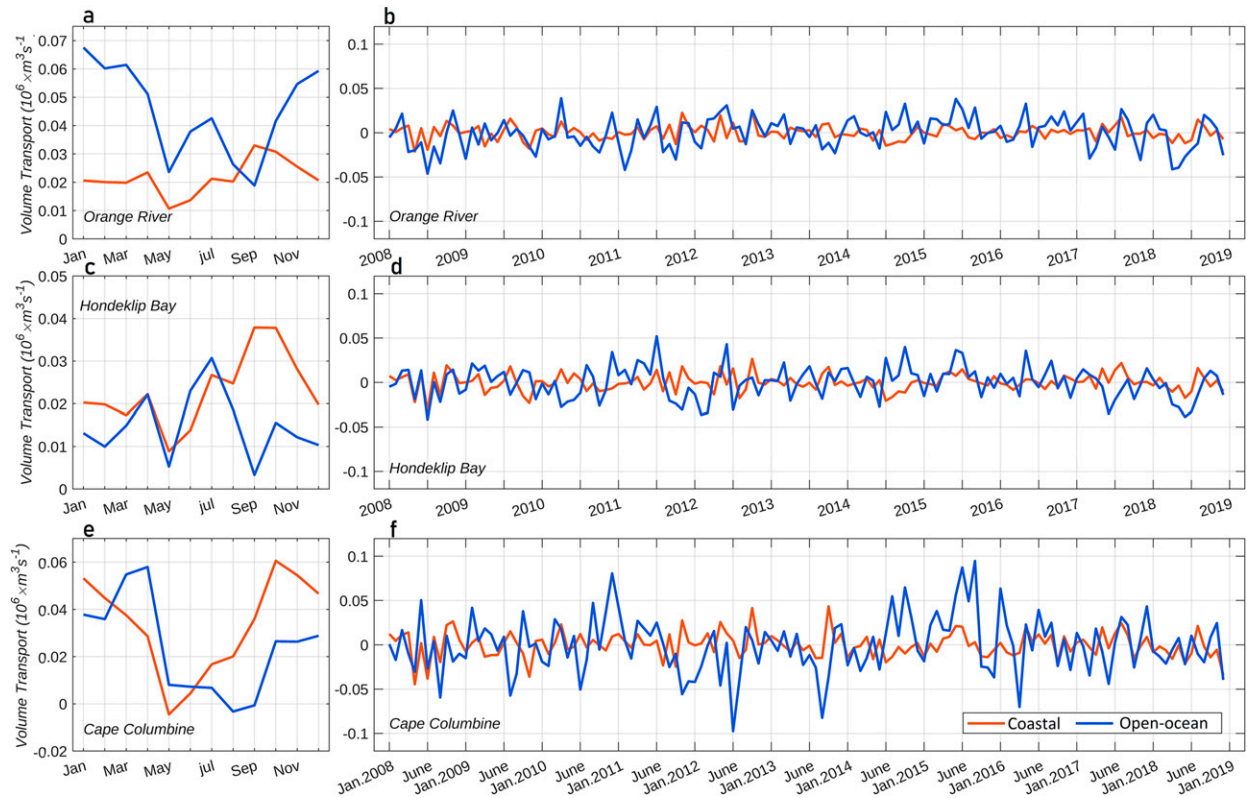


FIG. 9. As in Fig. 8, but for (a),(b) Orange River; (c),(d) Hondeklip Bay; and (e),(f) Cape Columbine. Note that the y axes in (a), (c), and (e) have different scales.

alternate with each other. The open-ocean upwelling events are mostly associated with opposite-sign anomalies in the coastal upwelling. This is in accordance with the correlation pattern obtained from the analytical consideration (Fig. 7; Fig. S11 in the online supplemental material).

Off Cape Frio, the biannual cycle is less pronounced relative to the Kunene cell (Fig. 8c). Coastal and open-ocean upwelling are often in opposite phases (Figs. 8c,d). Again, extreme events of the open-ocean upwelling are mostly concurrent with opposite-sign peaks in the coastal upwelling and vice versa (Fig. 8d). However, this opposite-phase relationship seems to be less pronounced in comparison with the Kunene cell. Again, this is in agreement with the result of analytical theory (Fig. 7a). Moreover, the correlation between coastal upwelled water in Cape Frio and Kunene is nearly 0.8. This further highlights the notion that these two upwelling cells can be considered as a single cell in terms of coastal upwelling (Lutjeharms and Meeuwis 1987; Jury 2017). However, this correlation for the offshore upwelled water and the corresponding annual cycle is relatively weak (0.27).

Fluctuations of the coastal and open-ocean upwelling are dominated by a discernible annual cycle in Walvis Bay with the explained variability of about 72% and 36%, respectively (Fig. 8e; Table 1). The coastal and open-ocean upwelling annual cycles are very similar. They peak in August and July, respectively (Fig. 8e). Strong and weak upwelling anomalies appear frequently concurrent, with the anomaly correlation

coefficient of about 0.22 (Fig. 8f). This correlation is rather ambiguous for the analytical theory. From the coast toward the ocean, the correlation changes the sign (Fig. S11 in the online supplemental material). In the vicinity of the coast, the coastal and the WSCD upwelling anomalies are positively correlated whereas they are anticorrelated farther offshore. Recall that an

TABLE 1. The explained variability (%) of monthly mean time series of coastal and open-ocean upwelled water off Kunene, Cape Frio, Walvis Bay, Lüderitz, Orange River, Hondeklip Bay, and Cape Columbine by their climatological annual cycle. The explained variability was obtained from the linear regression between the upwelling time series and their climatological annual cycle; sBUS and nBUS indicate the southern and northern BUS subsystems, respectively.

Location	Coastal upwelled water	Open-ocean upwelled water
	nBUS	
Kunene	57	14
Cape Frio	58	23
Walvis Bay	72	36
Lüderitz	45	23
	sBUS	
Orange River	36	44
Hondeklip Bay	44	16
Cape Columbine	29	66

intensive mesoscale and submesoscale dynamics is simulated off Walvis Bay that can largely modulate the offshore upwelling and influence the relationship between the coastal and the WSCD upwelling.

Significant parts of coastal and open-ocean upwelling variability in the Lüderitz cell can be explained by the seasonal variation (Table 1). The seasonal cycles are in phase between December and July and nearly in opposite phase between August and October (Fig. 8g). Similar to other stations, the annual cycle is more pronounced in the coastal upwelling (Table 1). The correlation between the upwelling anomalies is not statistically significant and the phase difference between them seems not to follow a rigid pattern (Fig. 8h).

In summary, the correlation across the coastal upwelling in different stations in the northern BUS (i.e., Kunene, Cape Frio, and Walvis Bay) is large and statistically significant, whereas the correlation derived from the open-ocean upwelling across different upwelling cells is generally weak and often not statistically significant.

Time series of the coastal and the open-ocean upwelling across Orange River, Hondeklip Bay, and Cape Columbine, located in the southern BUS, are shown in Fig. 9. The contribution of the annual cycle in total variability of the coastal upwelled water is large but slightly smaller than in the northern BUS (Table 1). The amount of the open-ocean upwelled water off Orange River is associated with a strong seasonal cycle explaining about 44% of the total variability (Fig. 9a; Table 1). The open-ocean upwelling peaks between December and March when the coastal upwelling appears to be very close to its annual average (Fig. 9a). Between August and October, the coastal upwelling peaks whereas its counterpart experiences its annual minimum (Fig. 9a). Given the large contribution of the seasonal cycle in the total variability of coastal and open-ocean upwelling off Orange River (Table 1) the phase difference between them seems to be strongly seasonally dependent. The anomaly correlation between the coastal and the open-ocean upwelling off Orange River is about 0.23.

Off Hondeklip Bay, the contribution of the seasonal cycle in the total variability of the coastal and the open-ocean upwelling is about 44% and 16%, respectively. From November to July, the coastal and the open-ocean upwelling evolve rather coherently whereas they are frequently in opposite phases between August and October (Fig. 9c). Again, the relationship between coastal and open ocean off Hondeklip Bay seems to be seasonally dependent but less pronounced in comparison with that in the Orange River cell. The correlation between the time series of upwelling anomalies is very small and not statistically significant (Fig. 9d).

For Cape Columbine, which is the southernmost selected station in this study, 66% of the open-ocean upwelling variability can be explained by the seasonal cycle alone (Table 1). The open-ocean and the coastal upwelling peaks at different times of the year (Fig. 9e). The coastal upwelling peaks in October and reaches its minimum in May. The open-ocean upwelling peaks in March–April and reaches its minimum in August–September. Overall, the correlation between the open-ocean and the coastal upwelling anomalies is about 0.20 and statistically significant, which is consistent with the result

of the analytical consideration. Given the underestimated Agulhas rings in the model, the found relationship between the coastal and the open-ocean upwelling in the vicinity of the Agulhas Bank (i.e., Cape Columbine upwelling cell) is likely to be spurious.

#### 4. Summary and conclusions

Using results of a state-of-the-art eddy-resolving ocean circulation model along with a linearized analytical approach for a simplified parameterization of wind-driven upwelling, the time and space variations of the water vertical velocity along the southwest African coast as far as Benguela upwelling system were investigated for the 2008–18 period. Satellite-derived Chl-a concentration was additionally analyzed.

In terms of the mean-state, the model and analytical approach agreed fairly well in which the intense upwelling associated with the coastal divergence of Ekman transport appeared to be the dominant factor across the entire BUS. The northward expansion of the offshore upwelling, supported by the theory of the BUS, was captured by the model and the analytical consideration. Remarkable in this respect is the spatial pattern of Chl-a concentration, which compares relatively well to the spatial pattern of the upwelling.

In agreement with previous studies, the model-based results revealed vigorous mesoscale and submesoscale activities (e.g., filaments and eddies) in the open-ocean domains with more pronounced signatures over the northern subsystem off Lüderitz and Walvis Bay (Bettencourt et al. 2012). We found these activities exert profound impacts on the wind-stress-curl-driven (WSCD) upwelling and substantially enhance its variability. Despite the interruptive impacts of mesoscale and submesoscale activities on the WSCD upwelling in the model, the consistency between model and analytical theory in terms of the total volume of upwelled water was retained. This result indicates that using near-real-time wind data in the analytical approach can provide a fast and reliable evaluation of the upwelling state for marine researchers who might need a real-time estimate of upwelling.

Also, our study shows that in terms of coastal upwelling, Kunene and Cape Frio cells can be considered as a single cell. However, WSCD upwelling across these cells showed some differences. We found an overall inverse relationship between the time evolution of the coastal and the WSCD upwelling in Kunene and Cape Frio. It indicates that the WSCD and the coastal upwelling frequently alternate each other, which implies that the weakening of one is very likely associated with the strengthening of another. Hence, the abundance of near-shore large-cell phytoplankton and open-ocean small-cell phytoplankton are expected to have a seesaw behavior that can largely regulate the marine food web.

In contrast, the coastal and open-ocean upwelling in Walvis Bay are identified with clear annual cycles and are frequently in phase. At the same time, the phase difference between the coastal and the open-ocean upwelling off Lüderitz appear to be strongly seasonally dependent. Like Lüderitz, the relationship between these two types of upwelling across the southern BUS varied seasonally. As opposed to the northern part of the

domain, the coastal and open-ocean upwelling of the southernmost sector of the BUS (i.e., Cape Columbine) closely followed each other. Overall, our results suggest that the time evolution of upwelling across the BUS is far more complicated than currently thought.

Because of the horizontal resolution of ASCAT gridded records, which is  $0.25^\circ$ , an accurate estimate of coastal wind drop-off, which is typically within a few tens of kilometers offshore in distance, still remains an issue of concern. To develop a more comprehensive understanding of this term and the connections between the coastal and WSCD upwelling, using a high-resolution regional atmospheric model simulation would be very beneficial. Our results further support the notion that local wind forcing, among other local and remote upwelling drivers (i.e., coastally trapped waves, thermocline variations, mixing), largely regulates the regional upwelling across the BUS. Thus, migration of the South Atlantic high pressure system or changes in the sea level pressure pattern over the African continent would lead to change in the magnitude, timing, and structure of upwelling across the BUS. With current satellite-based Chl-*a* data, a reliable estimate of climatic changes driven fluctuations in the plankton size structure, which is equally important as the total biomass for the marine ecologists and biologists, appears a big challenge. Hence, it seems essential to work toward a comprehensive monitoring system based on the size structure of planktons. Incontrovertible evidence for human-induced climate changes has become more frequent (IPCC 2014; Bordbar et al. 2015, 2019), which might have short- and far-reaching implications for the pelagic marine ecosystems across the BUS.

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**Data availability statement.** Data shown in this paper are available by request from the corresponding author (hadi.bordbar@io-warnemuende.de).

## REFERENCES

- Acevedo-Trejos, E., G. Brandt, J. Bruggeman, and A. Merico, 2015: Mechanisms shaping size structure and functional diversity of phytoplankton communities in the ocean. *Sci. Rep.*, **5**, 8918, <https://doi.org/10.1038/srep08918>.
- Atlas, R., R. N. Hoffman, J. Ardizzone, S. M. Leidner, J. C. Jusem, D. K. Smith, and D. Gombos, 2011: A cross-calibrated, multiplatform ocean surface wind velocity product for meteorological and oceanographic applications. *Bull. Amer. Meteor. Soc.*, **92**, 157–174, <https://doi.org/10.1175/2010BAMS2946.1>.
- Bachèlery, M. L., S. Ilig, and I. Dadou, 2016: Interannual variability in the South-East Atlantic Ocean, focusing on the Benguela Upwelling System: Remote versus local forcing. *J. Geophys. Res. Oceans*, **121**, 284–310, <https://doi.org/10.1002/2015JC011168>.
- Bakun, A., and C. S. Nelson, 1991: The seasonal cycle of wind stress curl in the subtropical eastern boundary current regions. *J. Phys. Oceanogr.*, **21**, 1815–1834, [https://doi.org/10.1175/1520-0485\(1991\)021<1815:TSCOWS>2.0.CO;2](https://doi.org/10.1175/1520-0485(1991)021<1815:TSCOWS>2.0.CO;2).
- Barlow, R., H. Sessions, M. Balarin, S. Weeks, C. Whittle, and L. Hutchings, 2005: Seasonal variation in phytoplankton in the southern Benguela: Pigment indices and ocean colour. *Afr. J. Mar. Sci.*, **27**, 275–287, <https://doi.org/10.2989/18142320509504086>.
- , D. Louw, M. Balarin, and J. Alheit, 2006: Pigment signatures of phytoplankton composition in the northern Benguela ecosystem during spring. *Afr. J. Mar. Sci.*, **28**, 479–491, <https://doi.org/10.2989/18142320609504200>.
- Beljaars, A. C. M., 1995: The parametrization of surface fluxes in large-scale models under free convection. *Quart. J. Roy. Meteor. Soc.*, **121**, 255–270, <https://doi.org/10.1002/qj.49712152203>.
- Benazzouz, A., S. Mordanea, A. Orbib, M. Chagdalia, K. Hilmib, A. Atillach, J. L. Pelegrid, and D. Hervée, 2014: An improved coastal upwelling index from sea surface temperature using satellite-based approach – The case of the Canary current upwelling system. *Cont. Shelf Res.*, **81**, 38–54, <https://doi.org/10.1016/j.csr.2014.03.012>.
- Bettencourt, J. H., C. López, and E. Hernández-García, 2012: Oceanic three-dimensional Lagrangian coherent structures: A study of a mesoscale eddy in the Benguela upwelling region. *Ocean Modell.*, **51**, 73–83, <https://doi.org/10.1016/j.ocemod.2012.04.004>.
- Bonino, G., E. Di Lorenzo, S. Masina, and D. Iovino, 2019: Interannual to decadal variability within and across the major eastern boundary upwelling systems. *Sci. Rep.*, **9**, 19949, <https://doi.org/10.1038/s41598-019-56514-8>.
- Bordbar, M. H., T. Martin, M. Latif, and W. Park, 2015: Effects of long-term variability on projections of twenty-first-century dynamic sea level. *Nat. Climate Change*, **5**, 343–347, <https://doi.org/10.1038/nclimate2569>.
- , M. H. England, A. Sen Gupta, A. Santoso, A. S. Taschetto, T. Martin, W. Park, and M. Latif, 2019: Uncertainty in near-term global surface warming linked to tropical Pacific climate variability. *Nat. Commun.*, **10**, 1990, <https://doi.org/10.1038/s41467-019-09761-2>.
- Caldwell, P. C., M. A. Merrifield, and P. R. Thompson, 2015: Sea level measured by tide gauges from global oceans—The joint archive for sea level holdings (NCEI accession 0019568), version 5.5. NOAA National Centers for Environmental Information, accessed 15 December 2018, <https://doi.org/10.7289/V5V40S7W>.
- Capet, X., F. Colas, J. C. McWilliams, P. Penven, and P. Marchesiello, 2008: Eddies in eastern boundary subtropical upwelling systems. *Ocean Modeling in an Eddying Regime, Geophys. Monogr.*, Vol. 177, Amer. Geophys. Union, 131–147.
- Chelton, D. B., R. A. DeSzoeke, and M. G. Schlax, 1998: Geographical variability of the first baroclinic Rossby radius of deformation. *J. Phys. Oceanogr.*, **28**, 433–460, [https://doi.org/10.1175/1520-0485\(1998\)028<0433:GVOTFB>2.0.CO;2](https://doi.org/10.1175/1520-0485(1998)028<0433:GVOTFB>2.0.CO;2).
- Clementi, E., and Coauthors, 2019: Mediterranean Sea analysis and forecast (CMEMS MED-Currents, EAS5 system). Copernicus Marine Environment Monitoring Service, accessed 21 January 2021, [https://doi.org/10.25423/CMCC/MEDSEA\\_ANALYSIS\\_FORECAST\\_PHY\\_006\\_013\\_EAS5](https://doi.org/10.25423/CMCC/MEDSEA_ANALYSIS_FORECAST_PHY_006_013_EAS5).

- Desbiolles, F., B. Blanke, A. Bentamy, and N. Grima, 2014: Origin of finescale wind stress curl structures in the Benguela and Canary upwelling systems. *J. Geophys. Res. Oceans*, **119**, 7931–7948, <https://doi.org/10.1002/2014JC010015>.
- , and Coauthors, 2017: Two decades [1992–2012] of surface wind analyses based on satellite scatterometer observations. *J. Mar. Syst.*, **168**, 38–56, <https://doi.org/10.1016/j.jmarsys.2017.01.003>.
- de Szoeke, R. A., and R. Richman, 1984: On wind-driven mixed layers with strong horizontal gradients—A theory with application to coastal upwelling. *J. Phys. Oceanogr.*, **14**, 364–377, [https://doi.org/10.1175/1520-0485\(1984\)014<0364:OWDMLW>2.0.CO;2](https://doi.org/10.1175/1520-0485(1984)014<0364:OWDMLW>2.0.CO;2).
- Doney, S. C., and Coauthors, 2012: Climate change impacts on marine ecosystems. *Annu. Rev. Mar. Sci.*, **4**, 11–37, <https://doi.org/10.1146/annurev-marine-041911-111611>.
- Duncombe Rae, C. M., 2005: A demonstration of the hydrographic partition of the Benguela upwelling ecosystem at 26°40'S. *Afr. J. Mar. Sci.*, **27**, 617–628, <https://doi.org/10.2989/18142320509504122>.
- Ekman, V. W., 1905: On the influence of the Earth's rotation on ocean-currents. *Ark. Mat. Astron. Fys.*, **2**, 1–52.
- Fennel, W., 1988: Analytical theory of the steady state coastal ocean and equatorial ocean. *J. Phys. Oceanogr.*, **18**, 834–850, [https://doi.org/10.1175/1520-0485\(1988\)018<0834:ATOTSS>2.0.CO;2](https://doi.org/10.1175/1520-0485(1988)018<0834:ATOTSS>2.0.CO;2).
- , 1999: Theory of the Benguela upwelling system. *J. Phys. Oceanogr.*, **29**, 177–190, [https://doi.org/10.1175/1520-0485\(1999\)029<0177:TOTBUS>2.0.CO;2](https://doi.org/10.1175/1520-0485(1999)029<0177:TOTBUS>2.0.CO;2).
- , and H. U. Lass, 2007: On the impact of wind curls on coastal currents. *J. Mar. Syst.*, **68**, 128–142, <https://doi.org/10.1016/j.jmarsys.2006.11.004>.
- , T. Junker, M. Schmidt, and V. Mohrholz, 2012: Response of the Benguela upwelling systems to spatial variations in the wind stress. *Cont. Shelf Res.*, **45**, 65–77, <https://doi.org/10.1016/j.csr.2012.06.004>.
- GEBCO Bathymetric Compilation Group, 2019: The GEBCO\_2019 Grid – A Continuous Terrain Model of the Global Oceans and Land. British Oceanographic Data Centre, National Oceanography Centre, [https://www.gebco.net/data\\_and\\_products/gridded\\_bathymetry\\_data/gebco\\_2019/gebco\\_2019\\_info.html](https://www.gebco.net/data_and_products/gridded_bathymetry_data/gebco_2019/gebco_2019_info.html).
- Gill, A. E., 1982: *Atmosphere-Ocean Dynamics*. International Geophysics Series, Vol. 30, Academic Press, 662 pp.
- Greatbatch, R., 1994: A note on the representation of steric sea level in models that conserve volume rather than mass. *J. Geophys. Res. Oceans*, **99**, 12 767–12 771, <https://doi.org/10.1029/94JC00847>.
- Griffies, S. M., 2012: Elements of the Modular Ocean Model (MOM): 2012 release. GFDL Ocean Group Tech. Rep. 7, 632 pp., [https://mom-ocean.github.io/assets/pdfs/MOM5\\_manual.pdf](https://mom-ocean.github.io/assets/pdfs/MOM5_manual.pdf).
- Hart, T. J., and R. I. Currie, 1960: The Benguela Current. *Discovery Reports*, **31**, 123–298.
- Hense, I., and A. Beckmann, 2008: Revisiting subsurface chlorophyll and phytoplankton distributions. *Deep-Sea Res. I*, **55**, 1193–1199, <https://doi.org/10.1016/j.dsr.2008.04.009>.
- Hösen, E., J. Möller, K. Jochumsen, and D. Quadfasel, 2016: Scales and properties of cold filaments in the Benguela upwelling system off Lüderitz. *J. Geophys. Res. Oceans*, **121**, 1896–1913, <https://doi.org/10.1002/2015JC011411>.
- Hutchings, L., and Coauthors, 2009: The Benguela current: An ecosystem of four components. *Prog. Oceanogr.*, **83**, 15–32, <https://doi.org/10.1016/j.pocean.2009.07.046>.
- IPCC, 2014: *Climate Change 2014: Impacts, Adaptation, and Vulnerability. Part B: Regional Aspects*, Cambridge University Press, 688 pp., <https://doi.org/10.1017/CBO9781107415386>.
- Jacox, M. G., C. A. Edwards, E. L. Hazen, and S. J. Bograd, 2018: Coastal upwelling revisited: Ekman, Bakun, and improved upwelling indices for the U.S. West Coast. *J. Geophys. Res. Oceans*, **123**, 7332–7350, <https://doi.org/10.1029/2018JC014187>.
- Johnson, J. A., 1976: Modifications of coastal currents and upwelling by offshore variations in wind stress. *Tellus*, **28**, 254–260, <https://doi.org/10.3402/tellusa.v28i3.10287>.
- Junker, T., M. Schmidt, and V. Mohrholz, 2015: The relation of wind stress curl and meridional transport in the Benguela upwelling system. *J. Mar. Syst.*, **143**, 1–6, <https://doi.org/10.1016/j.jmarsys.2014.10.006>.
- , V. Mohrholz, L. Siegfried, and A. van der Plas, 2017: Seasonal to interannual variability of water mass characteristics and currents on the Namibian shelf. *J. Mar. Syst.*, **165**, 36–46, <https://doi.org/10.1016/j.jmarsys.2016.09.003>.
- Jury, M. R., 2017: Coastal upwelling at Cape Frio: Its structure and weakening. *Cont. Shelf Res.*, **132**, 19–28, <https://doi.org/10.1016/j.csr.2016.11.009>.
- Koseki, S., H. Giordani, and K. Goubanova, 2019: Frontogenesis of the Angola–Benguela frontal zone. *Ocean Sci.*, **15**, 83–96, <https://doi.org/10.5194/os-15-83-2019>.
- Lamont, T., R. G. Barlow, and R. J. Brewin, 2019: Long-term trends in phytoplankton chlorophyll a and size structure in the Benguela upwelling system. *J. Geophys. Res. Oceans*, **124**, 1170–1195, <https://doi.org/10.1029/2018JC014334>.
- Large, W. G., J. C. McWilliams, and S. C. Doney, 1994: Oceanic vertical mixing: A review and a model with a nonlocal boundary layer parameterization. *Rev. Geophys.*, **32**, 363–403, <https://doi.org/10.1029/94RG01872>.
- Lass, H. U., and V. Mohrholz, 2008: On the interaction between the subtropical gyre and the subtropical cell on the shelf of the SE Atlantic. *J. Mar. Syst.*, **74**, 1–43, <https://doi.org/10.1016/j.jmarsys.2007.09.008>.
- Louw, D., K. Anja, V. Mohrholz, N. Wasmund, T. Junker, and A. Eggert, 2016: Seasonal and inter-annual phytoplankton dynamics and forcing mechanisms in the northern Benguela upwelling system. *J. Mar. Syst.*, **157**, 124–134, <https://doi.org/10.1016/j.jmarsys.2016.01.009>.
- Lübbecke, J. F., C. W. Böning, N. S. Keenlyside, and S. P. Xie, 2010: On the connection between Benguela and equatorial Atlantic Niños and the role of the South Atlantic Anticyclone. *J. Geophys. Res.*, **115**, C09015, <https://doi.org/10.1029/2009JC005964>.
- , P. Brandt, and M. Dengler, 2019: Causes and evolution of the southeastern tropical Atlantic warm event in early 2016. *Climate Dyn.*, **53**, 261–274, <https://doi.org/10.1007/s00382-018-4582-8>.
- Lutjeharms, J. R. E., and J. M. Meeuwis, 1987: The extent and variability of South-East Atlantic upwelling. *Afr. J. Mar. Sci.*, **5**, 51–62, <https://doi.org/10.2989/025776187784522621>.
- Mazzini, P. L. F., and J. A. Barth, 2013: A comparison of mechanisms generating vertical transport in the Brazilian coastal upwelling regions. *J. Geophys. Res. Oceans*, **118**, 5977–5993, <https://doi.org/10.1002/2013JC008924>.
- McCreary, J. P., 1981: A linear stratified ocean model of the coastal undercurrent. *Philos. Trans. Roy. Soc. London*, **302A**, 385–413, <https://doi.org/10.1098/rsta.1981.0176>.
- Mohrholz, V., C. H. Bartholomae, A. K. van der Plas, and H. U. Lass, 2008: The seasonal variability of the northern Benguela undercurrent and its relation to the oxygen budget on the shelf. *Cont. Shelf Res.*, **28**, 424–441, <https://doi.org/10.1016/j.csr.2007.10.001>.



- , A. Eggert, T. Junker, G. Nausch, T. Ohde, and M. Schmidt, 2014: Cross shelf hydrographic and hydrochemical conditions and their short term variability at the northern Benguela during a normal upwelling season. *J. Mar. Syst.*, **140**, 92–110, <https://doi.org/10.1016/j.jmarsys.2014.04.019>.
- Muller, A. A., V. Mohrholz, and M. Schmidt, 2013: The circulation dynamics associated with a northern Benguela upwelling filament during October 2010. *Cont. Shelf Res.*, **63**, 59–68, <https://doi.org/10.1016/j.csr.2013.04.037>.
- , Ch. J. Reason, V. Mohrholz, M. Schmidt, and A. Eggert, 2014: Computing transport budgets along the shelf and across the shelf edge in the northern Benguela during summer (DJF) and winter (JJA). *J. Mar. Syst.*, **140**, 82–91, <https://doi.org/10.1016/j.jmarsys.2014.02.007>.
- NASA Goddard Space Flight Center, 2018: Moderate-resolution Imaging Spectroradiometer (MODIS) Aqua Chlorophyll Data, 2018: Reprocessing. NASA OB.DAAC, accessed 20 February 2019, <https://doi.org/10.5067/AQUA/MODIS/L3B/CHL/2018>.
- Nicholson, S. E., 2010: A low-level jet along the Benguela coast, an integral part of the Benguela current ecosystem. *Climatic Change*, **99**, 613–624, <https://doi.org/10.1007/s10584-009-9678-z>.
- Pickett, M. H., and J. D. Paduan, 2003: Ekman transport and pumping in the California current based on the U.S. Navy's high-resolution atmospheric model (COAMPS). *J. Geophys. Res.*, **108**, 3327, <https://doi.org/10.1029/2003JC001902>.
- Reynolds, R. W., T. M. Smith, C. Liu, D. B. Chelton, K. S. Casey, and M. G. Schlax, 2007: Daily high-resolution blended analyses for sea surface temperature. *J. Climate*, **20**, 5473–5496, <https://doi.org/10.1175/2007JCLI1824.1>.
- Ricciardulli, L., and F. J. Wentz, 2016: Remote sensing systems ASCAT C-2015 daily ocean vector winds on 0.25 deg grid, version 02.1. Remote Sensing Systems, accessed 1 June 2019, <http://www.remss.com/missions/ascats>.
- Rubio, A., B. Blanke, S. Speich, N. Grima, and C. Roy, 2009: Mesoscale eddy activity in the southern Benguela upwelling system from satellite altimetry and model data. *Prog. Oceanogr.*, **83**, 288–295, <https://doi.org/10.1016/j.pocean.2009.07.029>.
- Rykaczewski, R. R., M. David, and J. Checkley, 2008: Influence of ocean winds on the pelagic ecosystem in upwelling regions. *Proc. Natl. Acad. Sci. USA*, **105**, 1965–1970, <https://doi.org/10.1073/pnas.0711777105>.
- Schmidt, M., and A. Eggert, 2016: Oxygen cycling in the northern Benguela upwelling system: Modelling oxygen sources and sinks. *Prog. Oceanogr.*, **149**, 145–173, <https://doi.org/10.1016/j.pocean.2016.09.004>.
- Shannon, L. V., 1985: The Benguela ecosystem. Part I: Evolution of the Benguela, physical features and processes. *Oceanogr. Mar. Biol.*, **23**, 105–182.
- , and G. Nelson, 1996: The Benguela: Large scale features and processes and system variability. *The South Atlantic*, G. Wefer et al., Eds., Springer, 163–210, [https://doi.org/10.1007/978-3-642-80353-6\\_9](https://doi.org/10.1007/978-3-642-80353-6_9).
- Siegfried, L., M. Schmidt, V. Mohrholz, H. Pogrzeba, P. Nardini, M. Böttlinger, and G. Scheuermann, 2019: The tropical–subtropical coupling in the southeast Atlantic from the perspective of the northern Benguela upwelling system. *PLOS ONE*, **14**, e0210083, <https://doi.org/10.1371/journal.pone.0210083>.
- Small, R. J., E. Curchitser, K. Hedstrom, B. Kauffman, and W. G. Large, 2015: The Benguela upwelling system: Quantifying the sensitivity to resolution and coastal wind representation in a global climate model. *J. Climate*, **28**, 9409–9432, <https://doi.org/10.1175/JCLI-D-15-0192.1>.
- Thomas, A. C., T. P. Strub, R. Weatherbee, and C. Corinne James, 2012: Satellite views of Pacific chlorophyll variability: Comparisons to physical variability, local versus nonlocal influences and links to climate indices. *Deep-Sea Res. II*, **77–80**, 99–116, <https://doi.org/10.1016/j.dsr2.2012.04.008>.
- Veitch, J., P. Penven, and F. Shillington, 2010: Modeling equilibrium dynamics of the Benguela current system. *J. Phys. Oceanogr.*, **40**, 1942–1964, <https://doi.org/10.1175/2010JPO4382.1>.
- Veitch, J. A., and P. Penven, 2017: The role of the Agulhas in the Benguela current system: A numerical modeling approach. *J. Geophys. Res. Oceans*, **122**, 3375–3393, <https://doi.org/10.1002/2016JC012247>.
- Wasmund, N., G. Nausch, and A. Hansen, 2014: Phytoplankton succession in an isolated upwelled Benguela water body in relation to different initial nutrient conditions. *J. Mar. Syst.*, **140**, 163–174, <https://doi.org/10.1016/j.jmarsys.2014.03.006>.
- Wentz, F. J., J. Scott, R. Hoffman, M. Leidner, A. Atlas, and J. Ardizzone, 2015: Remote sensing systems Cross-Calibrated Multi-Platform (CCMP) 6-hourly ocean vector wind analysis product on 0.25 deg grid, Version 2.0. Remote Sensing Systems, accessed 1 July 2019, [www.remss.com/measurements/ccmp](http://www.remss.com/measurements/ccmp).
- Zhaoyun, C., Y. Xiao-hai, J. Young-heon, J. Lide, and J. Yuwu, 2012: A study of Benguela upwelling system using different upwelling indices derived from remotely sensed data. *Cont. Shelf Res.*, **45**, 27–33, <https://doi.org/10.1016/j.csr.2012.05.013>.