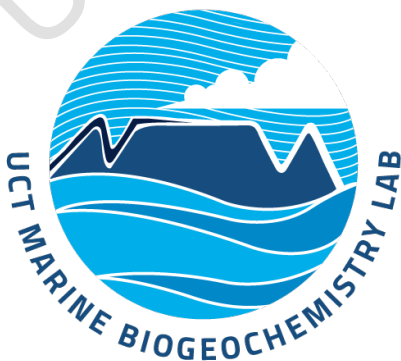


Retention processes in the Southern Benguela Upwelling System

Jonathan James Rogerson

A thesis presented for the degree of
Doctor of Philosophy



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Abstract

The Benguela Upwelling System is one of four major Eastern Boundary Upwelling Systems (EBUS) in the global ocean. Here, equatorward winds drive the upwelling of cold, deep, nutrient-rich waters along the coast, which drives high levels of biological productivity. A consequence of this high productivity is the prevalence of hypoxic conditions that co-occur with elevated nearshore and shelf bound retention of nutrients. This thesis focuses on the Southern Benguela Upwelling System (SBUS) and explores the role that oceanic fronts play at the seasonal, interannual and event-scales in restricting the offshore advection of material and elevating the residence times of matter along the shelf. Furthermore, the role of variable, near-coastal, summer alongshore winds on retention is also studied. Focus is given to how these winds impact the nearshore current velocities and the offshore positions of fronts. A physical ocean model configuration of the SBUS in conjunction with a gradient-based edge detection algorithm and a Lagrangian tracking code are used and evaluated against a combination of in situ and satellite data products. The model is able to resolve the low-frequency seasonal and interannual variability in sea surface temperature (SST) and surface circulation features. The results show alongshore fronts in summer to be effective barriers to the offshore advection of matter as they are defined by strong SST gradients and in general, are long and cohesive structures. In winter, fronts tend to be short and filamentous in nature, which limits their ability to restrict the offshore movement of material. Furthermore, fronts in summer tend to be found closer to the coast than those in winter. Transport of material on the SBUS shelf is governed by the surface equatorward jets and poleward undercurrents along the shelf bottom. Interannual variability in residence times for material on the shelf is driven by natural modes of variability in the alongshore winds that impact the shelf SST patterns which in-turn shape the spatial patterns and intensities of fronts. At the event-scale, variability in the alongshore wind profiles generate unique responses in the nearshore current velocities such that enhanced retention is associated with wind relaxation events. The transport dynamics, cross-shore mixing regimes, spatial patterns of fronts, especial coastal topography and bathymetry as well as the influence of the Agulhas Current create a unique oceanographic environment in the SBUS. Drivers of retention in other EBUS are important to understand, especially when considering biogeochemical processes or how these systems might be impacted by climate change in the future.

Declaration

I declare that this thesis is my own work and sources have been acknowledged and referenced using the Harvard convention. I understand that plagiarism is wrong and aside from the assistance and guidance of my academic supervisors the ideas presented are my own unless otherwise acknowledged.

Presentation of work

- Presented at: Ocean Sciences Meeting 2022. Talk title “Investigation of retention dynamics along the Southern Benguela Upwelling System: a modelling approach”
- Manuscript in preparation for Chapter 3: “Investigation of frontal dynamics along the Southern Benguela Upwelling System”. After revisions, title has been amended to “Frontal features and mixing regimes along the shelf region of the Southern Benguela Upwelling System” and being submitted to the Journal of Continental Shelf Research.

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*Dedicated to Teddy, Dawn and Skye;
keep wagging your tails my little
friends*

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List of Abbreviations

- Absolute Dynamic Topography – (ADT)
- Benguela Current Large Marine Ecosystem – (BCLME)
- Benguela Upwelling System - (BUS)
- Brunt-Vaisala Frequency – (BVF)
- Cape Peninsula – (CP)
- Coastal and Regional Ocean Community model – (CROCO)
- Coastal Retention Index – (CORE)
- Columbine Front – (CF)
- Copernicus Marine Environment Monitoring Service – (CMEMS)
- Eastern Boundary Upwelling Systems – (EBUS)
- Eddy Kinetic Energy – (EKE)
- El Niño-Southern Oscillation – (ENSO)
- European Centre for Medium-Range Weather Forecasts – (ECMWF)
- Finite-time Lyapunov Exponent – (FTLE)
- General Bathymetric Chart of the Oceans – (GEBCO)
- Global Drifter Programme – (GDP)
- Harmful Algal Blooms – (HAB)s
- Interquartile Range – (IQR)
- Intertropical Convergence Zone – (ITCZ)
- Lagrangian Coherent Structures – (LCS)
- Low Oxygen Waters – (LOW)
- Mixed Layer Depth – (MLD)
- Normalised Root Mean Square Deviation – (NRMSD)
- Operational Sea Surface Temperature and Sea Ice Analysis – (OSTIA)
- Regional Ocean Modeling System – (ROMS)
- ROMS offline – (ROFF)
- Root Mean Square Deviation – (RMSD)
- Sea Surface Height – (SSH)
- Sea Surface Temperatures – (SST)
- Shelf-Break Front – (SBF)
- Southern Annular Mode – (SAM)
- South Atlantic Anticyclone – (SAA)
- South Atlantic Central Waters – (SACW)
- South Atlantic Meridional Overturning Circulation Basin-wide Array – (SAMBA)
- Southern Benguela Upwelling System – (SBUS)
- Southern Oscillation Index – (SOI)
- St Helena Bay Monitoring Line – (SHBML)
- Standard Deviation – (STD)
- The Benguela Environment and Fishery Interaction and Training – (BENEFIT)
- Trophic Transfer Efficiency in the Benguela Current – (TRAFFIC)
- Upwelling Front – (UF)

Chapter 1: Introduction

“Who are we to tame you; the endless, erotic and eternal sea”

The oceans constitute 71% of the surface area of our planet, which is ironic, given the name of our celestial abode is *Terra* or Earth. Inspiring many celebrated authors throughout the ages, the grandeur and mystique of the oceans has been a major theme featuring in famous works such as Homer’s *Odyssey*, *Moby-Dick* by Herman Melville, the Nobel Prize winning *The Old Man and the Sea* by Ernest Hemingway as well as the renowned French song *La Mer* by Charles Trenet. Furthermore, while serving as an entity of enquiry and study for scholars and poets alike; the oceans play a significant socio-economic role. According to the World Bank (<https://www.worldbank.org/>), aquatic resources supplemented 3.3 billion people’s protein intake in 2019 and an estimated 600 million jobs are linked to the Blue Economy. Although the oceans are bountiful in natural aquatic resources, the aforementioned statistics mask a great spatial disparity regarding biological productivity. This said, the broad expanses of the world’s oceans are rather akin to a desert and elevated biological productivity is restricted to a select few regions where the interaction of the large-scale winds and presence of continental margins facilitate ideal physical conditions for enhanced biological activity.

Along the eastern margins of the Atlantic and Pacific Ocean, the four major Eastern Boundary Upwelling Systems (EBUS), consisting of the Benguela, California, Canary and Humboldt Systems, are found. While only representing 1% of the world’s ocean area, the four EBUS account for 20% of global fishery yields (García-Reyes *et al.* 2015) and 10% of oceanic new production (Fréon, Barange & Arístegui 2009). In general, alongshore equatorward winds stimulate the offshore advection of surface waters close to the coast which results in deep, cold, nutrient-rich waters being upwelled into the euphotic zone and nourishing high levels of biological activity. Accompanying this, EBUS are associated with surface equatorward jets that sit atop poleward undercurrents and they both flow respectively along the continental shelf margins within each region (Strub *et al.* 2013). This somewhat simple and idealized physical oceanographic description is applicable to all of the EBUS, albeit the reality is more complex and nuanced as significant variability is present between and within each system. While the driving mechanism for upwelling is ubiquitous in all four regions, Carr & Kearns (2003) showed that net primary production was not the same for all the EBUS. A review paper by Chavez & Messié (2009) suggested that the Benguela is the most productive system, followed closely by the Humboldt and Canary systems, with the California system being identified as the least productive of the four. Therefore, what are the major large-scale and local forcing mechanisms that contribute to these differences?

A summation of the important large-scale and local dynamics that shape biological productivity in EBUS is presented for the sake of contextualisation but more in-depth analyses can be found in the cited literature. At the large-scale, a major distinguishing feature between the various EBUS is the nature of the source waters that are upwelled onto the shelves (Carr & Kearns 2003). Variability in oxygen and nutrient concentrations of source waters will inevitably play a direct role in shaping biological productivity within a region and the formation and transport of deep water masses is a function of the global thermohaline circulation. Also, large-scale modes of natural variability such as El Niño exert greater influence on the abiotic and biotic components of EBUS in the Pacific versus the Atlantic (Zebiak 1993; Fréon *et al.* 2009). In the Benguela, Benguela Niños and Niñas result in local sea surface temperature anomalies which impacts marine productivity in the northern

sub-region of the Benguela as well as precipitation patterns along Southwest Africa (Florenchie *et al.* 2003; Lubbecke *et al.* 2010; Brandt *et al.* 2024).

The variability and non-linear interactions of local forcing mechanisms within each EBUS that shape biological productivity are complex and inherently region specific. Ultimately, the main driver is the variability of the alongshore wind regimes themselves as they generate upwelling along the coastal margins and force the offshore advection of surface waters (Carr & Kearns 2003; Chavez & Messié 2009). Works such as Lachkar & Gruber (2011, 2012) have also identified several other key drivers that influence productivity such as: the width of the continental shelf, mixed layer depth and eddy activity (see Rossi *et al.* 2009 and Pegliasco, Chaigneau & Morrow 2015). Furthermore, EBUS are often partitioned into sub-regions with distinct physical dynamics. Some examples include the perennial and seasonal nature of upwelling within the Benguela and California systems (Chavez & Messié 2009), the complex topography and presence of an archipelago in the Canary system (Fréon *et al.* 2009), the interaction of cold upwelled waters of the Humboldt with warm equatorial waters (Fréon *et al.* 2009) and the influence of the Agulhas Current on the southern reaches of the Benguela (Fennel 1999; Hutchings *et al.* 2009; Brandt *et al.* 2024). All these local forcing dynamics create unique and varying physical and biogeochemical environments for phytoplankton growth that leads to regional and intra-regional variability in net primary production (Messié & Chavez 2015) and higher trophic flows (Moloney *et al.* 2005). Focus will now be given to describing the oceanographic dynamics of the Benguela Upwelling System (BUS).

The BUS is defined by a broad northward flow along Southwest Africa and is bounded by two warm ocean currents, the Agulhas Current to the south and Angola Current to the north (Fennel 1999; Shannon 2001). The system is typically delineated into two distinct regions, separated by the Lüderitz upwelling cell (Hutchings *et al.* 2009). Upwelling is perennial in the northern Benguela whereas it is seasonal in the south; predominately occurring in austral late spring (November) and summer (December-February). The two regions also differ significantly in their oxygen characteristics (Monteiro & van der Plas 2006). The presence of Low Oxygen Waters (LOW) within the northern Benguela is primarily driven by a remote source of low oxygen that is transported by a poleward flow associated with a negative wind-stress curl (Monteiro & van der Plas 2006). In the southern Benguela region, LOW variability is locally driven through the physical processes of stratification, retention and advection as well as biogeochemical processes in the form of upwelled new production (Monteiro & van der Plas 2006).

Scholarship on the oceanography of the BUS is multi-disciplinary and includes topics such as: regional physical oceanography, climate variability, climate change, seasonal forecasting, numerical modelling, severe weather, marine biogeochemistry and marine spatial management. Historically, pioneer works on the oceanography of the Benguela can be traced to Rennel (1832) in his book titled: *An investigation of the currents of the Atlantic Ocean*. The seminal contributions of Nelson & Hutchings (1983), Shannon & Pillar (1986), Shannon *et al.* (1986) and Shannon, Agenbag & Buys (1987) laid the foundation for the study of the physical oceanography of the region. A review by Reason *et al.* (2005) on the progress of South African based research in the fields of atmospheric science and physical oceanography over 2000-2005 identified a downward trend in the number of academic papers published in the field of physical oceanography but noted an increase in the number of numerical modelling studies.

Although a similar and more contemporary review on the state of South African born research on the BUS is not present, a number of international collaborative oceanography based research

programmes have been established in the last two decades. One example is the South Atlantic Meridional Overturning Circulation Basin-wide Array (SAMBA) which aims to study the heat and salt fluxes at the base of the Atlantic (34°S) (Ansorge *et al.* 2014). The Benguela Environment and Fishery Interaction and Training (BENEFIT) research programme, which was carried out between 1998 and 2007 (Hampton & Sweijd 2008), was instrumental in strengthening the foundation of the already established (since 2002) Benguela Current Large Marine Ecosystem (BCLME) programme (Shillington *et al.* 2006). The BCLME project has subsequently given rise to several other projects within the academic spheres of marine spatial planning and biogeochemistry. The creation of the Benguela Current Convention, which is an international collaborative project between South Africa, Namibia and Angola on the marine management and planning within the Benguela, was an outcome of the BCLME programme (Finke *et al.* 2020). The recent Trophic Transfer eFFiciency in the Benguela Current (TRAFFIC) project (Martin *et al.* 2024) also owes its origin to the prior BCLME and BENEFIT programmes and its primary focus is on studying the biogeochemistry and biology within the system.

The increased availability of high resolution ocean re-analysis products, increases in computing power and a growing array of open source physical ocean models can be credited with the expansion of numerical modelling studies concerning the BUS (Reason *et al.* 2005). The numerical modelling studies of Penven *et al.* (2001), Veitch, Penven & Shillington (2009), Veitch, Penven & Shillington (2010), Veitch & Penven (2017) and Veitch *et al.* (2018) serve as important scholarship on the topic. Along with a growing array of numerical modelling studies there has been advancing research dedicated to understanding the nutrient cycling (Gutknecht *et al.* 2013; Burger *et al.* 2020; Flynn *et al.* 2020; Rixen *et al.* 2021) and oxygen dynamics (Jarre *et al.* 2015; Pitcher & Probyn 2017; Pitcher & Louw 2021) in the Benguela system. The intersection between the desire to understand the physical oceanographic dynamics of the BUS coupled with the advantages and developments of numerical modelling presents a unique opportunity to investigate physical drivers in the BUS that are important for biogeochemical cycles.

Using the BUS as a case study, this thesis focusses on the Southern Benguela Upwelling System (SBUS) with the aim of exploring the physical oceanographic dynamics that contribute to and elevate the shelf-bound retention of matter and nutrients.

Critical to enhancing biological productivity is ultimately the retention of nutrients along the shallow shelf regions to facilitate phytoplankton growth; although, a caveat to elevated retention is the prevalence of hypoxia and harmful algal blooms within EBUS (Pitcher *et al.* 2010; Pitcher & Jacinto 2019). Arguably, the low fisheries yields in the BUS relative to its high primary productivity may be in part due to the retention dynamics within the region (Martin *et al.* 2024). A recent paper by Flynn *et al.* (2020) hypothesized the significance of alongshore fronts in fostering ‘nutrient trapping’ along the SBUS shelf as the fronts act to restrict the offshore advection of material. Furthermore, Pitcher *et al.* (2010) has attempted to combine the dynamics of nearshore currents as well as the alongshore wind regimes to better describe the prevalence of harmful algal blooms and hypoxic events within the SBUS. From the literature, the role of alongshore fronts and nearshore processes play a major role in impacting the retention of material on the SBUS shelf. Consequently, two main research questions emerge:

- 1) What role do fronts play in modulating the retention and cross-shelf advection of nutrients and particulate matter in the SBUS?**

2) **How does the variability in nearshore wind patterns affect the retention and cross-shelf transport of material along the SBUS?**

The two main research questions are addressed in this thesis using six chapters. The second chapter gives background literature on the BUS and contextualizes the oceanography of the SBUS. The seasonal wind regimes and modes of atmospheric variability that drive upwelling along the coast as well as variability in the coastal topography for the system are detailed in order to highlight the different upwelling regimes present in the northern and southern regions of the Benguela. Focus is then given to describing the oceanography of the SBUS. The influence of the Agulhas Current on the physical oceanography of the region, the seasonal jet/frontal dynamics along the shelf and the wind-driven ocean circulation patterns are detailed. In lieu of a dedicated methodology chapter, chapters 3, 4 and 5 are written as standalone works; each with their own dedicated methodology, literature evaluation, results, discussion and conclusion sections as well as supplementary material. The thesis uses a numerical modelling approach to address the set-out aims and research questions. A physical ocean model configuration of the SBUS is used throughout the study; with a comprehensive evaluation available in chapter 3. The various computed metrics, numerical tools and statistical tests employed in the study are detailed in the respective methodologies of each chapter. These three core results chapters are titled:

- Chapter 3) Investigation of frontal variability along the Southern Benguela Upwelling System;
- Chapter 4) Investigation of retention dynamics along the Southern Benguela Upwelling System: a modelling approach; and
- Chapter 5) The role of variable alongshore summer wind profiles on nearshore retention dynamics along the Southern Benguela Upwelling System.

Chapter 3 forms the foundation of the thesis and directly addresses the first research question by presenting a comprehensive evaluation of the model outputs and the chapter focuses on the seasonal dynamics of fronts along the shelf region as well as their physical drivers. This chapter builds the context for understanding the physical dynamics of fronts and their role in the retention and transport of material. Chapter 4 then builds upon chapter 3 by adding a series Lagrangian tracking experiments to quantitatively compute seasonal and interannual retention times of particles along the shelf. The role that alongshore frontal features play in limiting the cross-shore advection of material is explored in greater detail. Chapter 5 addresses the second research question by assimilating the Lagrangian techniques of chapter 4 and the important physical dynamics of fronts from chapter 3 in order to investigate wind-driven variability on the nearshore retention of material. The chapter notes the importance of alongshore winds in driving upwelling as well as influencing the nearshore current profiles and the offshore positions of oceanic fronts.

Chapter 6 is a key-findings and final discussion section that summarizes the thesis and broadly frames the identified retention processes within the broader academic literature and aims of the study. Three main themes are addressed: (1) the major retention processes in the SBUS, (2) the uniqueness of retention dynamics and specificities to the region and (3) suggested further avenues of research that can build and expand upon the aims and objectives of this work. A cumulative reference list for chapters 1-6 is then presented. An Appendix of mathematical derivations of the major fluid dynamic equations is presented at the end for further details on common oceanographic

terms and equations referenced throughout the thesis. Overall, this thesis represents an advancement in the scholarship of numerical modelling and physical oceanography for the SBUS and the major findings have applicability to other areas of research such as nutrient and oxygen cycling within the region.

Chapter 2: Background

“Join the navy, see the world”

The Benguela Upwelling System (BUS) is one of four major Eastern Boundary Upwelling Systems (EBUS) in the global ocean (Hutchings *et al.* 2009). Along with the California, Humboldt, and Canary systems, EBUS play an important role in global marine biological productivity as they cumulatively occupy ~1% of the ocean surface area but contribute up to 20% of the world’s fishery yields (Garcia-Reyes *et al.* 2015). In general, equatorward winds stimulate the upwelling of cold, deep, nutrient-rich waters along coast margins which drives high levels of biological productivity. Critical to enhancing biological productivity is ultimately the retention of nutrients along the shallow shelf regions to facilitate phytoplankton growth; although, a caveat to elevated retention is the prevalence of hypoxia ($O_2 < 60 \mu M$) and Harmful Algal Blooms (HABs) within EBUS which can be permanent, seasonal and/or episodic in nature (Peña *et al.* 2010; Pitcher *et al.* 2010; Breitburg *et al.* 2018). An understanding of the physical oceanographic processes that impact retention dynamics along the SBUS has implications on hypoxia and nutrient cycling within the region.

This background chapter is divided into two main sections. The first overviews the major physical oceanographic features of the BUS and focus is given to the large-scale atmospheric wind regimes, the variable coastal topography and influence of the Agulhas Current. The second section details the oceanography of the SBUS. Attention is given to the seasonal and event-scale frontal dynamics, major shelf circulation patterns and nearshore current regimes in the region as they all significantly impact the retentiveness of the shelf environment.

2.1 Oceanography of the Benguela Upwelling System

The BUS (Fig. 2.1) is defined by a broad northward flow along southwest Africa and it is bounded by two warm currents: the Agulhas Current to the south and the Angola Current to the north (Fennel 1999; Hutchings *et al.* 2009). The oceanography of the BUS is governed by the interaction of four major factors: equatorward wind stress, mesoscale atmospheric perturbations, topography and the influence of the Agulhas Current system (Hutchings *et al.* 2009; Veitch & Penven 2017).

2.1.1 Wind regimes and atmospheric variability

Forming part of the South Atlantic subtropical gyre, the BUS is influenced by several key atmospheric features: the South Atlantic and South Indian Anticyclone, Intertropical Convergence Zone (ITCZ), cloud band convergence and the continental low pressure (Hutchings *et al.* 2009). Dominating the system is the South Atlantic Anticyclone which gives rise to southerly winds along the west coast of southern Africa. During austral summer, a continental low-pressure system develops, and this leads to an intensification of the zonal pressure gradients and intensified winds (Tyson, Garstang & Swap 1996). In winter, the major atmospheric circulations shift north, following the migration of the ITCZ, which results in low-level southerly winds dominating most of the BUS; however, south of 30°S, the region is influenced by westerly winds which drive mid-latitude systems, predominantly cold fronts (Tyson *et al.* 1996).

Superimposed on the seasonal changes in the low-level winds is considerable mesoscale, synoptic, intra-seasonal and interannual variability (Shillington *et al.* 2006). On synoptic time scales, the upwelling-favorable equatorward winds are perturbed by the passing of cold fronts, coastally trapped low pressure systems and mesoscale features such as berg winds and land-sea breezes (Shillington *et al.* 2006). On the interannual scale, the dominant global mode of variability is the El Niño-Southern Oscillation (ENSO). ENSO impacts the large-scale atmospheric circulation of the South Atlantic which in turn influences the Sea Surface Temperatures (SST)s within the Benguela region (Reason *et al.* 2006; Rouault, Pohl & Penven 2010). Along the west coast of southern Africa, upwelling favorable winds were found to be weaker (stronger) during El Niño (La Niña) phases which corresponded to warmer (cooler) than average SST conditions along the coastal regions in early summer and particularly from February to May (Rouault *et al.* 2010).

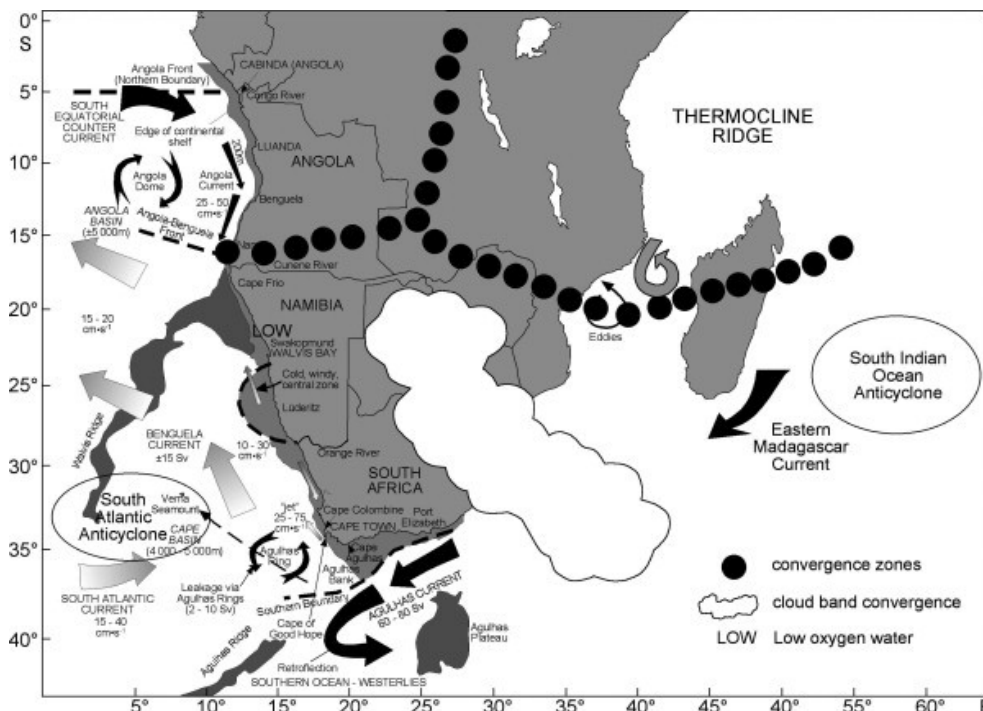


Figure 2.1: Oceanography of the BUS showing the major atmospheric, topographical and physical ocean features that impact the region (Hutchings *et al.* 2009)

Similar to El Niño in the Pacific, anomalous warm events occur in the South Atlantic off the coasts of Namibia and Angola. These events are named Benguela Niños and are generated following a relaxation in the intensity of the trade winds in the western equatorial Atlantic (Florenchie *et al.* 2003; Lubbecke *et al.* 2010). They tend not to occur in unison with their Pacific counterpart (Florenchie *et al.* 2003; Lubbecke *et al.* 2010; Rouault *et al.* 2018); although, it has been postulated that the modification of the trade winds during ENSO periods may lead to the development of Benguela Niños during late austral spring and early summer (Reason *et al.* 2006). Benguela Niños are therefore a prominent low frequency mode of variability, and they are associated with a southward displacement of the Angola-Benguela front following the intrusion of warm and saline equatorial waters (Reason *et al.* 2006). A consequence of this is anomalous rainfall patterns and a modification of regional fish stocks along Southwest Africa (Shillington *et al.* 2006). The Southern Annular Mode (SAM) is another important mode of interannual variability, influencing the atmospheric dynamics of the high latitude regions of the Southern Hemisphere as well as the

rainfall patterns over the Western Cape of South Africa (Reason *et al.* 2006; Rouault *et al.* 2010). However, unlike ENSO, the relationship between SAM and coastal SST variability off southern Africa is less coherent (Rouault *et al.* 2010). In summation, large-scale atmospheric variability arising from SAM and the interactions of ENSO as well as smaller scale temporal and spatial atmospheric perturbations impact the upwelling regimes along the southwest coast of Africa.

2.1.2 Topography and dynamics of the shelf environment

Several important topographical features such as: a series of canyons and capes, a narrow shelf environment and the broad Agulhas Bank play a significant role in impacting the oceanographic circulation in the BUS. The orientation of the shoreline is N-S from Cape Town up to Cape Frio where it then changes to NE. Three major embayments are found at Lobito (12°S), Walvis Bay (23°S) and St Helena Bay (32°S). The shelf environment can be considered the region extending from the coast seaward towards the continental shelf-break and varies between 150-600 m in depth (Boyd 1987). Furthermore, the shelf width varies significantly in the longshore direction. From Cape Point to Cape Columbine, the shelf is narrow and steep before broadening and shoaling north of Cape Columbine; with the 200 m isobath extending near to 100 km offshore. At 30°S, the shelf widens further to approximately 170 km before narrowing and steepening south of Lüderitz and remaining such further north to Cape Frio.

The currents along the shelf region of the BUS are driven by equatorward wind stress (Fennel *et al.* 1999) as well as topographically steered by the shelf-edge (Veitch *et al.* 2010). Upwelling intensity along the shelf is not contiguous and instead limited to several upwelling cells at: Columbine (33°S), Namaqua (30°S), Lüderitz (27°S), Walvis Bay (23°S), Namibia (21°S) and Kunene (18°S). These enhanced upwelling cells are associated with the alongshore wind patterns which in-turn are related to the smaller scale orientation of the coastline (Veitch *et al.* 2009). To understand the influence of topography on the upwelling dynamics, Veitch *et al.* (2010) used the Regional Ocean Modeling System (ROMS) (refer to Shchepetkin & McWilliams 2005) and conducted a sensitivity analysis whereby the shelf and coastline were smoothed between 32° and 21°S. Veitch *et al.* (2010) found that the smoothed topography simulation generated a contiguous upwelling zone in the alongshore direction, but the upwelling fluxes were lower than the peak upwelling fluxes resolved in the reference experiment. Aligning with observational measurements, Veitch *et al.* (2010) numerically demonstrated the enhanced upwelling intensity downstream of prominent capes.

The phasing of upwelling in the Benguela is linked to the seasonal shifts in the large-scale atmospheric circulations (Shillington *et al.* 2006). Using ROMS, Veitch *et al.* (2009) highlighted regional differences in the seasonal phasing of upwelling in the Benguela system. Veitch *et al.* (2009) noted that the northward migration of the South Atlantic Anticyclone during winter had no affect on the approximately perennial upwelling-favorable southerly winds north of Lüderitz; whereas to the south, the wind regime shifted and became dominated by westerlies. Furthermore, the seasonal shift in position of the large-scale atmospheric features induces a maximum wind-stress at Lüderitz resulting in the most intense upwelling within the BUS. Consequently, the upwelling cell of Lüderitz represents a physical boundary which separates the more perennial upwelling environment to the north from the strongly seasonal upwelling in the south that sees a maximum in upwelling during summer (Veitch *et al.* 2009; 2010).

The shelf environment typically has strong and narrow equatorward jets as well as poleward undercurrents along the shelf bottom and slope (Nelson 1989). Taking a depth-integrated (0-1000

m) analysis of the annual mean large-scale circulation of the Benguela, Veitch *et al.* (2009; 2010) found that the flow regime to the south of Lüderitz was dominated by a meandering north-westerly flow while to the north, the flow was poleward. Veitch *et al.* (2009) explained that the paradoxical flow pattern to the north relative to the predominantly southerly winds was driven by the Sverdrup relation. The broad Agulhas Bank forms the southmost extent of the shelf environment and is an important spawning ground for pelagic fish (Japs, Sims & Smale 1994). Northward jets transport larvae to nursery grounds such as St Helena Bay (Penven *et al.* 2001; Ragoasha *et al.* 2019). Furthermore, the oceanography of the Agulhas Bank is complex as it interacts with both the Agulhas and Benguela Current. Jackson *et al.* (2012) showed that the Agulhas Bank is heavily influenced by the Agulhas Current. Most notable is the effect the Agulhas Current has on the biogeochemistry for the bank through the stimulation of shelf-edge upwelling along the southeastern edge. Furthermore, the generation of internal waves acts to vertically displace phytoplankton and this uplift leads to elevated levels of chlorophyll concentrations and the creation of mesoscale biological hotspots. In addition, the meandering of the Agulhas Current onto the bank itself has been shown to entrain shelf particles and impact the cross-shelf transport of material (Jackson *et al.* 2012).

2.1.3 Influence of the Agulhas Current

The Agulhas Current has been shown to influence the current intensities and biogeochemistry along the Agulhas Bank (Jackson *et al.* 2012). The current also exerts a significant effect on the BUS. The Agulhas Current is one of the most energetic western boundary currents in the world and it retroflects south of the Agulhas Bank which results in the formation of rings, eddies and filaments that influence the mesoscale variability of the Benguela system (Duncombe Rae, Shannon & Shillington 1989; Duncombe Rae 1991; Shillington *et al.* 1992; Lutjeharms 2006; Shillington *et al.* 2006; Rubio *et al.* 2009; Veitch & Penven 2017). The leakage of the Agulhas into the Atlantic Ocean via the Benguela Current through the Agulhas eddy corridor is thought to be an important avenue for the transport of salt and energy from the South Indian Ocean into the South Atlantic and contribute to the global thermohaline circulation (Duncombe Rae 1991; Lutjeharms 2006; Beal *et al.* 2011). Duncombe Rae *et al.* (1989) was one of the first to document the presence of an Agulhas ring in the South Atlantic Ocean and subsequent work by Shillington *et al.* (1992) showed how Agulhas rings can interact with the shelf-edge upwelling fronts in the southern Benguela which can lead to the development of long seaward filaments. Though Shillington *et al.* (1992) noted the limited productive potential of these filaments, it was recognized that they could be important in the offshore advection of biota and nutrients.

Using ROMS, Veitch & Penven (2017) conducted a numerical experiment where the affect of the Agulhas Current was removed in order to understand its respective influence on the BUS. Veitch & Penven (2017) corroborated the earlier observational work of Duncombe Rae *et al.* (1989) as well as Shillington *et al.* (1992) by showing how Agulhas rings impact the positions of the upwelling fronts and volume transport in the Benguela. Furthermore, Veitch & Penven (2017) noted the warming effect of the Agulhas in the offshore waters and its subsequent impact on steepening the cross-shore thermal gradients across the shelf-edge that leads to an intensification of the shelf-edge jets (refer to Sec. 3.2.1). Therefore, the Agulhas Current has a profound effect on both the large-scale offshore and near-shore physical dynamics and variability in the position of the retroflexion will impact the transfer of salt and energy into the BUS (Lutjeharms 2006; Reason *et al.* 2006; Beal *et al.* 2011).

In appreciating the broad oceanography of the Benguela it is obvious that the large-scale atmospheric dynamics, ocean circulation patterns, coastal topography and upwelling regimes vary meridionally. Consequently, in studying the BUS, many authors (some examples: Boyd 1987; Shillington *et al.* 2006; Monteiro *et al.* 2006; Veitch *et al.* 2009; Veitch *et al.* 2010; Lamont *et al.* 2018) sub-divide the Benguela into a northern and southern region, separated by the Lüderitz upwelling cell. Some authors further sub-divide the system such as Lamont *et al.* (2018) who distinguishes the Agulhas Bank as being a separate region or Monteiro *et al.* (2006) who references the ‘Central Benguela’. For this thesis work the SBUS will be defined as the region extending from 28° to 36°S and the physical oceanography of the region is explored in the subsequent section.

2.2 Physical oceanography of the SBUS

The SBUS (Fig. 2.2 and 2.3) experiences seasonal upwelling during late spring (November) and summer (December – February), forced by southerly winds. The region is highly productive, with St Helena Bay supporting 40-50% of the primary productivity (Burger *et al.* 2020). In exploring the upwelling dynamics that contribute to retention along the shelf, attention needs to be given to several important hydrodynamic features: a strong shelf-edge jet between Cape Point and Cape Columbine, alongshore hydrographic fronts north of Cape Columbine, and the wind-driven ocean circulation patterns around St Helena Bay.

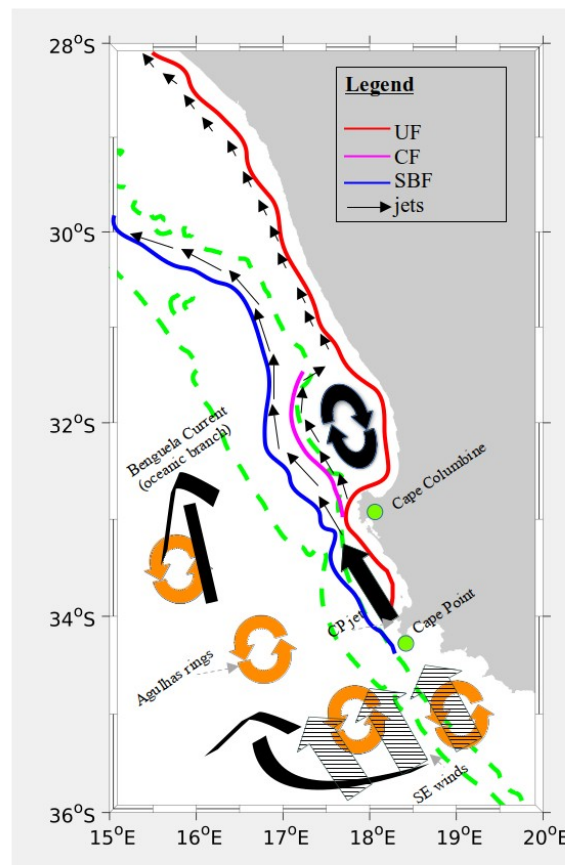


Figure 2.2: Schematic showing the major summer physical oceanographic features of the SBUS. Southerly winds stimulate upwelling along the shelf resulting in the formation of the Upwelling front (UF) and Columbine front (CF). The shift to northerly winds in winter precludes the formation of these alongshore fronts. The Shelf-break front (SBF) is a permanent hydrographic feature of the region. Green dotted lines show the position of the 200 and 500 m isobath. (Copy of Fig. (4.1))

2.2.1 Alongshore hydrographic fronts and jets

Through a scaling analysis the predominant motion in the ocean is horizontal but vertical motion plays a crucial role in linking the deep nutrient-rich ocean to the sunlit surface layer (Vélez-Belchí & Tintoré 2001). A front is a region of intense cross-shelf gradients in temperature, salinity and/or density. Applying the thermal wind relation, fronts are associated with localized equatorward jets (Brink 1983). Furthermore, frontal regions are linked with locally enhanced vertical advection and can be regions of subduction and elevated carbon export (Spall 1995; Kuebel Cervantes *et al.* 2004; Stukel *et al.* 2017).

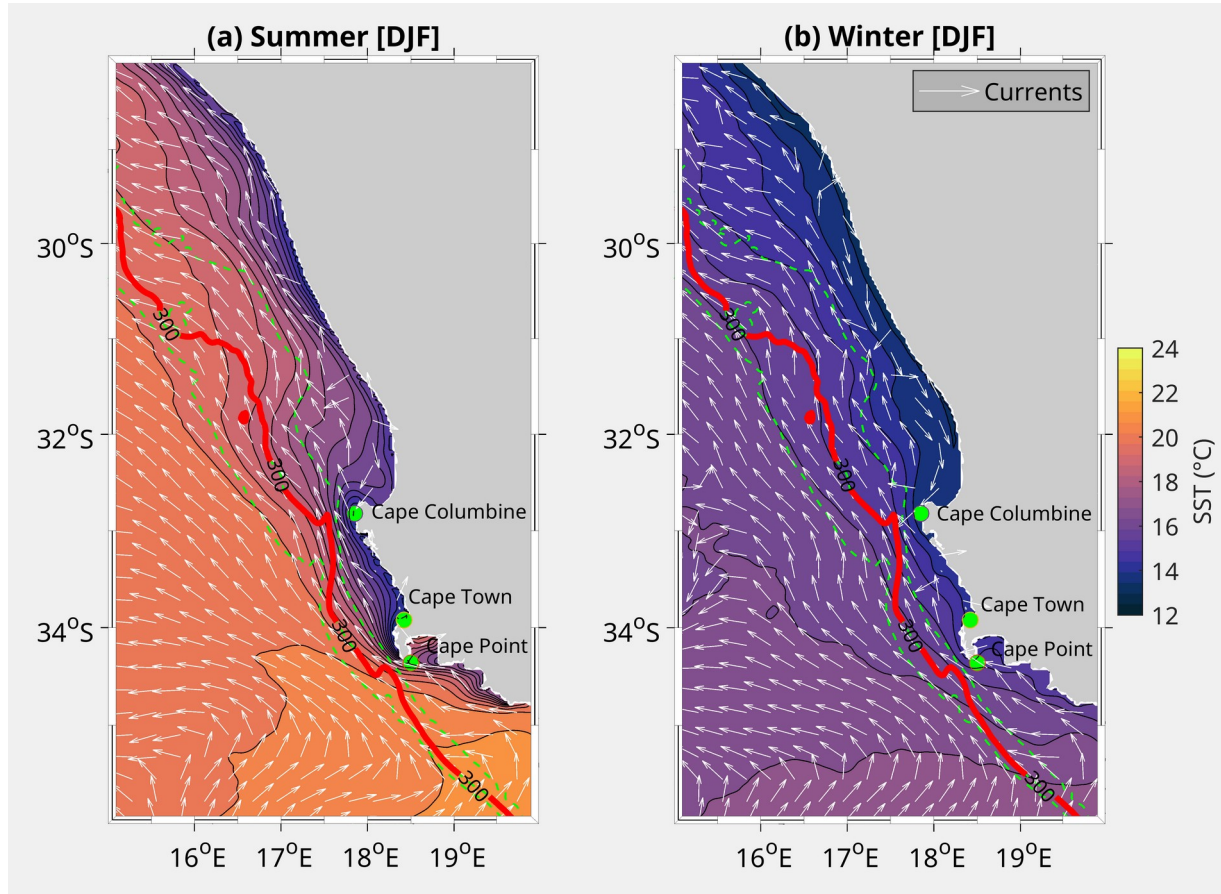


Figure 2.3: Mean summer (a) and winter (b) SST and flow regimes in the SBUS from a 16 year regional model simulation (CROCO configuration used in this work). White arrows show the normalized surface integrated (upper 50 m) flow patterns. The green dotted lines show the 100 and 500 m isobaths and the red line demarcates the 300 m isobath.

From observational measurements, Boyd & Nelson (1989) noted a strong north-westward barotropic flow along the shelf-edge off the Cape Peninsula which in literature is sometimes referenced as the “Good Hope Jet”. Adopting the naming convention of Veitch *et al.* (2018), this jet will be referred to as the Cape Peninsula (CP) jet to distinguish it from the jet features north of Cape Columbine. With currents speeds ranging from 0.4 – 1.2 m/s (Armstrong *et al.* 1987; Boyd & Nelson 1989; Nelson 1989; Veitch *et al.* 2018) the CP jet is set-up along the cross-shelf thermal gradient between the cool upwelled shelf waters and the warm offshore waters (Fig. 2.3). The position and intensity of the jet is subject to the seasonal wind patterns, but numerical modelling results show the CP jet as being a permanent feature, regardless of the prevailing wind patterns, and

flows between the 200 and 500 m isobath (Veitch *et al.* 2018). The jet was found to intensify during the upwelling season and extend further south to Cape Agulhas whereas in winter, the absence of upwelling weakened the intensity of the jet, and the flow becomes restricted between Cape Point and Cape Columbine.

In summer, the CP jet bifurcates (Fig. 2.2) into two jets north of Cape Columbine (Lamont *et al.* 2015; Veitch *et al.* 2018). The offshore branch along the 500 m isobath is primarily fed by the CP jet and is referred to as the Shelf-Break Front (SBF). The irregular nature of this shelf-edge jet can give rise to the formation of anti-cyclonic and cyclonic eddies along the shelf (Rubio *et al.* 2009). Though these shelf-born eddies are smaller than Agulhas Rings, they have vertical extents of 400 - 800 m and a mean diameter ranging from 70 – 150 km (Rubio *et al.* 2009). As these eddies propagate westwards into the open ocean, they entrain shelf particles and play a role in the exchange of heat, salt, and biogeochemical properties (Rossi *et al.* 2008; Rubio *et al.* 2009; Gruber *et al.* 2011; Hernández-Carrasco *et al.* 2014).

The nearshore jet flowing along the 200 m isobath is referred to as the Columbine Front (CF) (Fig. 2.2) and is fed to a lesser extent by the CP jet and is enhanced by local upwelling off Cape Columbine (Penven *et al.* 2000; Veitch *et al.* 2018). Interconnecting with the work of Nelson (1981) and Penven *et al.* (2000), the CF is the oceanic appendage of the Columbine plume which is formed off Cape Columbine during upwelling events. In winter, the double jet feature is not visible as the CP jet becomes more restricted to the shelf region of the Cape Peninsula and only feeds the inner shelf region off Cape Columbine (Veitch *et al.* 2018). Furthermore, the dominance of westerly winds precludes the formation of an upwelling plume, and the associated CF. It is misconstruing to say that the SBF and CF merge, as suggested by Lamont *et al.* (2015), and instead, the single winter jet feature off Cape Columbine can be envisaged as the shoreward migration of the SBF in the absence of the CF and is termed as the ‘merged CF’ in the work of Flynn *et al.* (2020).

Inshore of the CF, Armstrong *et al.* (1987) observed a weak jet closer to the coast and this highly variable and seasonal front is termed ‘the upwelling front’ (UF) (Fig. 2.2). The UF is an alongshore feature, typically being long and continuous in nature; however, cold filaments of upwelled water from the coastal zone can spread offshore and warp the linear structure of the UF, generating protruding appendages that extend the offshore extent and influence of the front (Armstrong *et al.* 1987; Brink 1987; Shillington *et al.* 1992). The formation of the UF begins with an initially warm and well stratified surface layer with no prominent cross-shore density gradients present. The onset of upwelling-favourable southerly winds results in the offshore advection of surface waters through the process of Ekman transport. Cold, deep, nutrient-rich waters are upwelled into the surface layer, resulting in the formation of prominent cross-shore thermal/density gradients. As the cooler more dense upwelled waters horizontally converge with the warmer and less dense surface waters a boundary region develops between the two water masses. Frontogenesis intensifies as the cross-shore density gradients become more defined. Following geostrophic adjustment, a defined baroclinic jet forms at the boundary interface (Hoskins 1982; Spall 1995; McWilliams 2021). The outcome of this frontogenesis is the formation of an UF that acts to separate colder and more nutrient-rich waters on the coastal side of the front from warmer surface oceanic waters on the oceanic side.

During spring and early summer, intense upwelling can result in the UF merging with the CF (Lamont *et al.* 2015) and if the UF is forced far enough offshore, it can cause the advection of phytoplankton blooms past Cape Columbine and into Saldanha Bay (Probyn *et al.* 2000). In the

SBUS, alongshore hydrographic fronts have been numerically (Penven *et al.* 2000) and observationally (Lamont *et al.* 2015; Flynn *et al.* 2020) demonstrated to elevate the coastal retention of particles. Using observational data along the St Helena Bay Monitoring Line (SHBML), Lamont *et al.* (2015) observed the subduction of upwelled shelf waters below the alongshore fronts and proposed a mechanism whereby alongshore hydrographic fronts reduce offshore advection and losses of phytoplankton and nutrients. This led to the development of the idea of ‘nutrient trapping’ by Flynn *et al.* (2020) as a means to explain the high levels of biological productivity and hypoxia observed in the SBUS.

2.2.2 Wind-driven ocean circulation patterns

The spatial patterns of nearshore winds have a strong influence on the dynamics of upwelling (Blanke *et al.* 2002; Capet, Marchesiello & McWilliams 2004). Upwelling can be driven by alongshore wind-stress which results in offshore Ekman transport as well as cyclonic wind-stress curl which is forced by a nearshore wind drop-off. However, Ekman transport is an order of magnitude greater than curl-driven upwelling in coastal areas (Lamont *et al.* 2018) and is more important to ecosystem function as Ekman transport is more efficient in the vertical advection of deeper waters into the euphotic zone (Capet *et al.* 2004; Bordar, Mohrholz & Schmidt 2021). Furthermore, the phasing of alongshore winds can influence the near-surface current velocities close to the coast (Fawcett *et al.* 2008) as well as generate recirculation patterns in the lee of prominent capes (Penven *et al.* 2000).

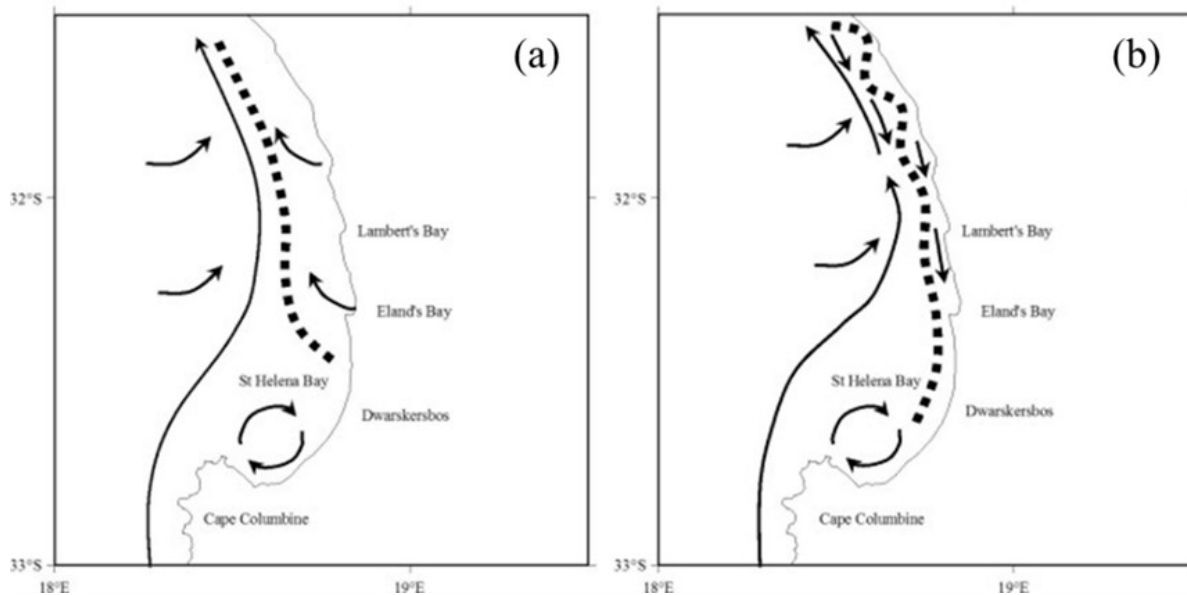


Figure 2.4: Solid black lines show the flow patterns while the checkered line denotes the accumulation of phytoplankton as well as the position of the upwelling front. (a) conditions during upwelling (b) for relaxation. (Adapted from Pitcher *et al.* 2010)

Nelson (1989) observed surface (<40 m) baroclinic jets in the nearshore region along the shelf and noted that during periods of wind relaxation, a poleward flow would develop. A mechanism for this phenomenon was numerically demonstrated by Gan & Allen (2002a, b), who attributed the poleward flow to the development of a negative pressure gradient that is set-up by the interaction of the wind-induced coastal current with alongshore variations in the coastline and bottom topography. Using mooring data downstream of Cape Columbine, Fawcett *et al.* (2008) described the sensitivity

of the near-surface (upper 60 m) circulation patterns to the prevailing alongshore winds. During late summer and winter, near-surface currents were predominantly southward while early in the upwelling season, the currents were directed northward in response to southerly winds. These differing nearshore flow regimes can actually be seen in Fig. (2.3), as the currents are northward in summer whereas they are directed south in winter. Furthermore, Fawcett *et al.* (2008) distinguished between low (> 1 day) and high (< 1 day) frequency currents. Low frequency currents reversed phase in periods of 2-5 days, in keeping with the phasing of the upwelling winds, and were barotropic in their flow. High frequency currents were forced by diurnal winds associated with land-sea breezes. Fearon *et al.* (2020) showed that the land-sea breeze is resonant with the inertial response at 30°N/S . Consequently, the inertial oscillations generated nearshore baroclinic currents with speeds of 0.3 m/s and the 180° phase shift was associated with currents at 10-15 m (Fawcett *et al.* 2008). The large shear created by these phase shifts enhances diapycnal mixing along the shelf and this implicates the cross-shore pressure gradients that drive alongshore geostrophic currents (Fearon *et al.* 2020). Though inertial oscillations can be important for diapycnal mixing, they are inevitably superimposed on the low frequency currents which ultimately govern the mean northward or southward flow.

St Helena Bay is an important nursery ground for pelagic larvae (Penven *et al.* 2001) and is linked with the high retention of particles due to the recirculation structures resulting from the interaction of the topography as well as equatorward winds (Penven *et al.* 2000; Lett *et al.* 2006). Equatorward winds stimulate the formation of an upwelling plume in the lee of Cape Columbine producing a cyclonic eddy (Fig. 2.2 and 2.3) (Penven *et al.* 2000). The associated circulation pattern of the cyclonic eddy acts to isolate the nearshore region from the offshore domain, elevating particle retention on the coastal side of the plume (Penven *et al.* 2000; Pitcher & Nelson 2006; Pitcher *et al.* 2010). Combining the work of Penven *et al.* (2000) as well as Fawcett *et al.* (2008), Pitcher *et al.* (2010) suggested a mechanism for the enhancement of HABs in St Helena Bay during the upwelling season. Equatorward winds stimulate upwelling, resulting in the formation of an UF, and northward currents along the shelf (Fig. 2.4a). A relaxation in the wind results in a shoreward migration of the UF and the development of a poleward flow which will transport blooms south and into St Helena Bay (Fig. 2.4b).

Chapter 3: Investigation of frontal variability along the Southern Benguela Upwelling System

*“All fronts are equal but some are more equal than others”
- Adapting an adage from G. Orwell’s “Animal Farm”*

Summary

Mesoscale/submesoscale fronts are ubiquitous features in all Eastern Boundary Upwelling Systems (EBUS) and are associated with prominent sea surface temperature (SST) gradients and enhanced vertical advection. The physical dynamics of mesoscale fronts impact biogeochemical cycles in EBUS. Using a physical ocean model of the region in conjunction with a gradient-based edge-detection algorithm, the frontal dynamics and shelf-based circulation patterns in the Southern Benguela Upwelling System (SBUS) were investigated. Evaluated against a combination of *in situ* and satellite data products the model was able to resolve the low-frequency interannual and intra-seasonal variability in SST and surface circulation features. An artefact of the ocean boundary physical forcing conditions, the model resolves a very deep winter mixed layer depth and has warm temperature biases at the surface and at depth along the shelf; which impact the seasonal positions and intensities of the alongshore jets and fronts. The non-linear interaction and seasonal shift in the wind regime and stratification play an important role in driving frontal variability. In summer, fronts are usually long and continuous alongshore features that are defined by strong SST gradients and associated with prominent Lagrangian Coherent Structures (LCS). In winter, fronts are found further offshore and are more filamentous and numerous compared to summer. They are defined by weaker SST frontal gradients, shorter frontal lengths and are not associated with obvious LCS. In summer, the main driver of frontal variability is the alongshore winds whereas stratification plays a more significant role in winter. Strong stratification in summer acts to limit submesoscale variability. Cross-shore circulation features on the shelf show greater intensity during summer and likely play an important role in the transport and retention of material.

3.1 Introduction

Eastern Boundary Upwelling Systems (EBUS) are oases of marine biological productivity as they cumulatively occupy ~1% of the ocean surface area but contribute up to 20% of the world’s fishery yields (García-Reyes *et al.* 2015). In EBUS, alongshore equatorward winds stimulate offshore Ekman transport which generates upwelling along coastal margins. As a consequence of these upwelling-favourable winds, cold, nutrient-rich deep waters are upwelled to the surface, sustaining high biological productivity. Off southwest Africa, the Benguela Upwelling System (BUS) is one such EBUS and the region is bounded by two warm currents; the Agulhas Current to the south and the Angola Current to the north (Fennel 1999; Hutchings *et al.* 2009). Owing to differences in the upwelling regimes, the BUS is typically divided into two regions, separated by the Lüderitz upwelling cell (Veitch, Penven & Shillington 2009). The northern region experiences perennial

upwelling while the south experiences seasonal upwelling, predominantly in summer (December - February), forced by southerly winds (Shillington *et al.* 2006; Monteiro *et al.* 2006; Hutchings *et al.* 2009). In this section, focus will be given to the southern region, referred to as the Southern Benguela Upwelling System (SBUS).

The SBUS is highly productive, with St Helena Bay supporting 40-50% of the primary productivity (Burger *et al.* 2020). Upwelling intensity along the shelf is not contiguous and instead limited to upwelling cells at the Cape Peninsula (34°S), Cape Columbine (33°S) and Namaqua (30°S). These enhanced upwelling cells are associated with the alongshore wind patterns which in-turn are related to the smaller scale orientation of the coastline (Veitch *et al.* 2009). Several alongshore hydrographic fronts exist north of Cape Columbine and are intrinsically linked with the positions of the surface equatorward jets (see Fig. 2.2). Set-up along the cross-shelf sea surface temperature (SST) gradient between the cool upwelled shelf waters and the warm offshore waters, a prominent equatorward jet, flowing between the 200 and 500 m isobath, exists between Cape Agulhas and Cape Columbine (Armstrong *et al.* 1987; Boyd & Nelson 1989; Nelson 1989; Veitch *et al.* 2018). The polyonymous jet is herein referred to as the Cape Peninsula (CP) jet, with the Benguela Jet or Good Hope Jet being used equivalently in literature. The position and intensity of the jet is influenced by the seasonal wind patterns, with *in situ* observations (Van der Lingen & Hugget 2003) and numerical modelling results (Veitch *et al.* 2018) showing the CP jet as being a permanent oceanographic feature, regardless of the prevailing wind patterns.

In summer, the CP jet bifurcates into two jets north of Cape Columbine (Lamont *et al.* 2015; Veitch *et al.* 2018). The offshore branch along the 500 m isobath is primarily fed by the CP jet and is referred to as the Shelf-Break Front (SBF). The nearshore jet flowing along the 200 m isobath is referred to as the Columbine Front (CF) and is fed to a lesser extent by the CP jet and is enhanced by local upwelling off Cape Columbine (Penven *et al.* 2000; Veitch *et al.* 2018) and the associated Columbine plume (Nelson 1981; Penven *et al.* 2000). In winter (June - August), the double jet feature is not visible as the CP jet becomes more restricted to the shelf region of the Cape Peninsula and only feeds the inner shelf region off Cape Columbine (Veitch *et al.* 2018). Furthermore, the dominance of westerly winds precludes the formation of an upwelling plume, and the associated CF. From a climatology based on 12 years of more-or-less monthly cruise data, Lamont *et al.* (2015) suggested that the SBF and CF merge to form a single jet feature in winter.

Inshore of the CF, Armstrong *et al.* (1987) observed a weak jet closer to the coast and this highly variable front is termed the 'upwelling front' (UF). The UF is an alongshore feature, typically being long and continuous in nature; however, cold filaments of upwelled water from the coastal zone can spread offshore and warp the linear structure of the UF; generating protruding appendages that extend the offshore extent and influence of the front (Armstrong *et al.* 1987; Brink 1987; Shillington *et al.* 1992). During spring and early summer, intense upwelling may even result in the UF merging with the CF (Lamont *et al.* 2015). Along the shelf-edge, the irregular nature of the SBF can give rise to the formation of anti-cyclonic and cyclonic eddies (Rubio *et al.* 2009). Though these shelf-born eddies are smaller than Agulhas Rings, they have vertical extents of 400 - 800 m and a mean diameter ranging from 70 - 150 km (Rubio *et al.* 2009). As these eddies propagate westwards into the open ocean, they entrain shelf particles and play a role in the exchange of heat and salt (Rubio *et al.* 2009) as well as the off-shelf transport of biomass (Rossi *et al.* 2008) and nutrients (Gruber *et al.* 2011) which can lead to a reduction in biological productivity.

Mesoscale fronts are a ubiquitous feature in all EBUS (Wang, Castelao & Yuan 2015; Oerder *et al.* 2018), with typical horizontal length scales of $\sim 10 - 100$ km (Thomas, Tandon & Mahadevan 2008; McWilliams 2021). As well as being associated with prominent SST gradients, fronts are regions of enhanced vertical advection (Garvine 1974; Mahadevan & Archer 2000; Vélez-Belchí & Tintoré 2001; Kuebel Cervantes *et al.* 2004). In the Peruvian Upwelling System, Hauschildt *et al.* (2021) showed how fronts can have a negative impact on the primary productivity for a region through the subduction of recently upwelled nutrients. Similar conclusions were drawn by Stukel *et al.* (2017) in the California Current who also noted the potential of frontal regions as being places for the enhanced export of carbon into the deep. Associated with the elevated biomass at fronts, Woodson & Litvin (2015) found increased fishing yields associated with higher frontal activity in the California Current. Within the SBUS, fronts have been observationally (Probyn *et al.* 2000; Barange & Pillar 1992; Pitcher & Nelson 2006; Lamont *et al.* 2015; Flynn *et al.* 2020) and numerically (Penven *et al.* 2000) demonstrated to enhance the retention of particles and nutrients. Flynn *et al.* (2020) proposed the idea of ‘nutrient trapping’ as a mechanism to explain the enhancement of hypoxia along the SBUS shelf in late summer. For nutrient trapping, it is hypothesized that the alongshore hydrographic fronts act to restrict the off-shelf transport of nutrients and biology. The resulting sinking and accumulation of organic matter generates high levels of respiration resulting in elevated bottom oxygen consumption along the shelf.

This chapter aims to explore the frontal dynamics along the SBUS using a physical ocean model of the region. The objective is to understand the spatial and temporal variability of fronts in conjunction with their oceanographic properties. The chapter begins with a description of the model configuration and edge detection algorithm used. From there, the results section is divided into three parts: (1) model evaluation, (2) frontal edges detected in model vs observations (3) description of the frontal dynamics at the seasonal and monthly climatological scale. The discussion that follows briefly addresses the model’s performance and capabilities in the SBUS. From there, focus is given to contrasting the physical oceanographic characteristics of fronts in summer and winter. The discussion ends with exploring the drivers of frontal variability, focusing on the roles of wind and stratification.

3.2 Methodology

The physical ocean model CROCO (Coastal and Regional Ocean Community model) based on ROMS (Shchepetkin & McWilliams 2005, Penven *et al.* 2008) was used to study the circulation and frontal dynamics in the SBUS. Built upon ROMS_AGRIF (Debreu *et al.* 2012), CROCO is a free-surface, terrain-following coordinate model that is well suited for regional applications and it solves the incompressible primitive equations based on the Boussinesq and hydrostatic approximations. The code used is that of CROCO v1.1 version available at <http://www.croco-ocean.org> (Debreu *et al.* 2012).

3.2.1 Model Configuration

The model was configured as a single domain (Fig. 3.1) with a 3 km ($1/36^\circ$) horizontal spatial resolution consisting of 60 vertical sigma levels. For the vertical S-coordinate stretching parameters; 5 was used for the surface and 0 for the bottom while the topographical smoothing factor was set to 0.25. The oceanic boundary conditions were provided by GLORYS (<https://glorys>), which is an eddy-resolving physical ocean reanalysis product with a spatial resolution of $1/12^\circ$ (Lellouche *et al.* 2021). The atmospheric bulk forcing used the ERA5 3-hourly

product (<https://www.ecmwf.int>) which is the fifth generation of the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis and has a spatial resolution of $0.25^\circ \times 0.25^\circ$ for atmospheric variables (Hersbach *et al.* 2020). For the topographical data the General Bathymetric Chart of the Oceans (GEBCO) 2020 grid was used (<https://www.gebco.net>) and it provides a global grid at 15 arc-second intervals. The boundary conditions were handled following the work of Marchesiello, McWilliams & Shchepetkin (2001). The K-profile parameterisation was used for the vertical mixing regime (Large *et al.* 1994). To prevent excessive mesoscale activity, the relative winds as opposed to the absolute were applied to the surface ocean (Renault, McWilliams & Mason 2017; Renault *et al.* 2020). The tidal regime for the region is semi-diurnal with a spring tide amplitude of 2 m (Branch & Griffiths 1988) and negligible tidal currents (Fawcett *et al.* 2008). Winter rainfall patterns along the west coast of South Africa promote an arid environment with few major rivers (Botai, Botai & Adeola 2018). Consequently, no tidal or river forcing was present; analogous to previous modelling works conducted in the region by Penven *et al.* (2001) and Veitch, Penven & Shillington (2010). The simulation was run from January 2002 to December 2018 with outputs saved every 1-day. The first two years were reserved for spin-up and results are presented for January 2004 to December 2018.

3.2.2 Model evaluation and observational data sets

Numerical models are extensively used in ocean sciences for hypothesis testing and a key step in scientific rigour is the evaluation of a model's accuracy or "skill" (Oreskes, Shrader-Frenchette & Belitz 1994; Stow *et al.* 2009). In evaluating a numerical model against observational data sets, performance criteria need to be set (Rykiel Jr 1996). This typically involves statistical metrics and qualitative assessments of the models behaviour in order to determine a temporal and spatial domain for which there is sufficient confidence in the application of the model's outputs to address the desired scientific questions. Attention will be given to the model's ability to resolve the physical oceanography of the SBUS at the monthly and seasonal temporal scales with statistical analyses conducted for the zoomed domain shown in Fig. (3.1).

Tab. (3.1) summarizes the various observational data sets used to evaluate the model configuration. SST was evaluated against the Operational Sea Surface Temperature and Sea Ice Analysis (OSTIA) product which is a Level 4 (L4) product that combines multi-sensor satellite and *in situ* data (Good *et al.* 2020). From the daily fields, monthly SST averages were computed for OSTIA. Similarly, the model SST was computed at the same temporal scale and then interpolated onto the OSTIA grid for statistical analysis. Geostrophic currents and sea surface height (SSH) were evaluated against the derived geostrophic velocities and absolute dynamic topography (ADT) from the L4 Copernicus Marine Environment Monitoring Service (CMEMS) multi-satellite altimetry product.

$$EKE = \frac{[u_g'^2 + v_g'^2]}{2} \quad (3.1)$$

EKE was calculated using Eq. (3.1) where u_g' and v_g' are the zonal and meridional geostrophic velocity anomalies; computed by subtracting the long-term mean from the respective component fields. The geostrophic velocity components and EKE were computed daily on the native model and observational grids and then the model outputs were interpolated onto the CMEMS grid for comparison.

In situ data from the M4 South Atlantic Meridional Overturning Circulation Basin-Wide Array (SAMBA) mooring (Kersalé *et al.* 2018), located at a depth of 300 m, was used to assess the surface and bottom currents as well as bottom temperatures along the shelf edge. On the shelf itself, across-shelf temperature sections and mixed layer depth (MLD) were assessed relative to the 12 hydrographic stations comprising the St Helena Bay Monitoring Line (SHBML) (Lamont *et al.* 2015). The MLD for the model and observations was calculated following the method of Lorbacher *et al.* (2006) which uses local maxima in the curvature of density or temperature profiles to identify the MLD. To account for differences in the vertical resolution between the SHBML and the model, both were vertically interpolated onto a new vertical grid with the following depth levels, [5, 10, 15, 20, 25, 30, 40, 50, 60, 70, 80, 90, 100, 125, 150, 175, 200, 250, 300, 400, 500, 600, 700, 800, 900, 1000, 1100, 1200, 1400, 1600]. Two moorings at 70 m and 20 m along the shelf (Tab 1. for location) were used to evaluate the bottom temperatures close to the coast (Pitcher *et al.* 2014).

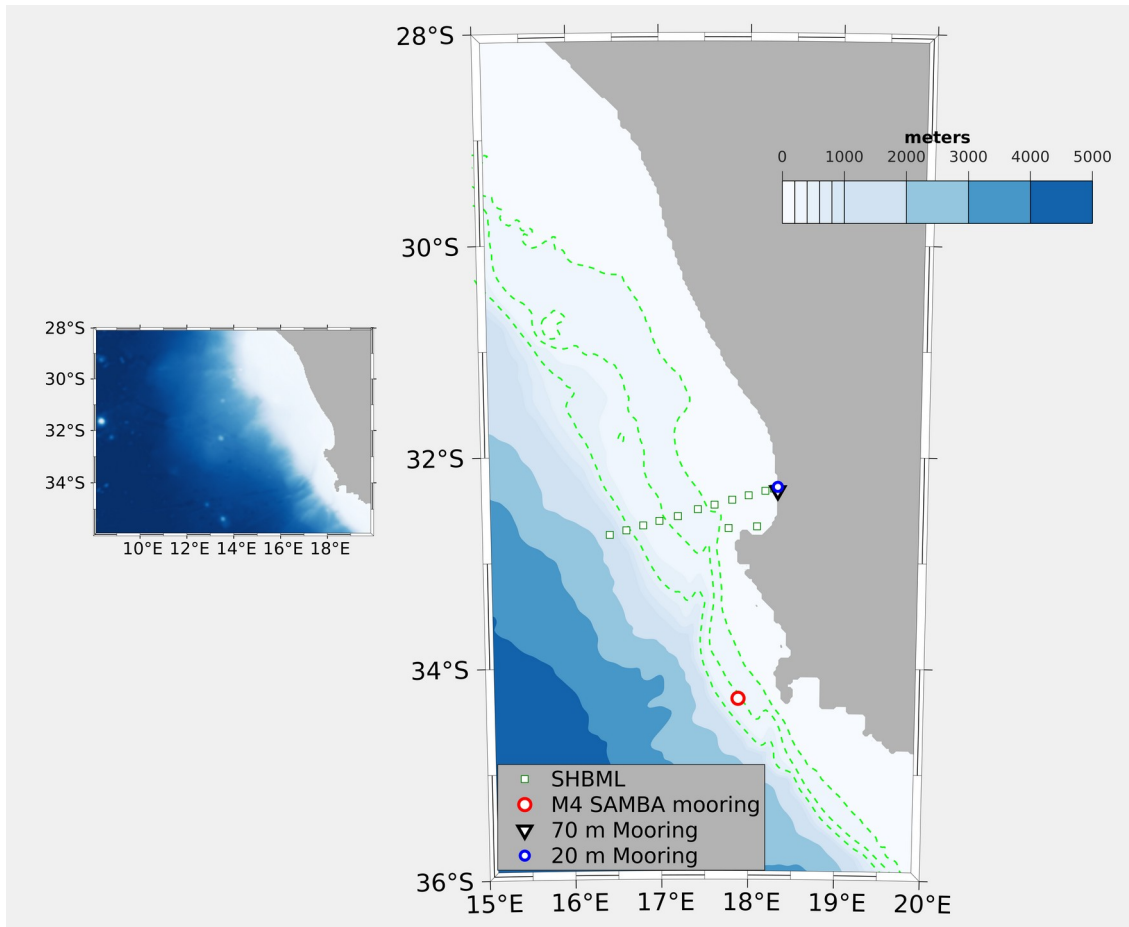


Figure 3.1: Full model (left) and zoomed domain (right) showing bathymetry and location of moorings and sampling stations. Green dotted lines show the position of the 200, 300 and 500 m isobath

Stow *et al.* (2009) proposed several useful statistical metrics when conducting model evaluation such as: correlation coefficient, centred Root Mean Square Deviation (RMSD) and mean bias (Supp. 3.1S). The correlation coefficient measures the tendency of predicted and observed values to vary together and has a range of -1 to 1; where a negative score indicates an inverse relationship and an ideal score is close to 1. However, a correlation near 1 can be achieved between observed and predicted values even when they differ by a constant factor. The RMSD and bias metric give an indication of the size of the discrepancies between predicted and observed values. Scores close to

zero indicate a close match though a score of zero is nearly impossible. Unlike correlation coefficient, it is difficult, if not impractical, to intercompare RMSD scores between different data sets of different variables. The measure of RMSD is in the native units of the variable of interest and is further dependent on the scale of the numbers used. This inherently means that proposing a blanket RMSD criterion for evaluation purposes is nonsensical and rather the RMSD should be interpreted independently for each data set. A compromise is to compute the Normalised RMSD (NRMSD), where the RMSD is divided by the mean of the observations to give a percentage of variability and make the score unitless.

Table 3.1: Observational data sets used in the model evaluation

Name	Spatial resolution	Temporal resolution	Reference
OSTIA	5 km	Monthly: 2004 – Dec 2018	Good <i>et al.</i> (2020)
CMEMS	0.25°x0.25°	Daily: 2004 - 2017	https://doi.org/
SHBML	Sections	Monthly: 2004 – Dec 2011	Lamont <i>et al.</i> (2015)
SAMBA M4 mooring	-34.294 S; 17.880 E	Daily: Jul 2014 – Oct 2018	Kersalé <i>et al.</i> (2018)
St Helena Bay 70 m mooring	-32.329 S; 18.183 E	Daily: Feb 2012– 8 Nov 2012	Pitcher <i>et al.</i> (2014)
Elands Bay 20 m mooring	-32.292 S; 18.318 E	Daily: Nov 2008 – Nov 2011	Pitcher <i>et al.</i> (2014)

No one statistical score will give an all encompassing representation of a model's accuracy and hence the incorporation of multiple metrics allows for a more diverse interrogation of a model's performance. Furthermore, some statistical metrics can complement each other such as residual measures when interpreting the correlation coefficient score. Consequently, graphical methods such as the Taylor Diagram (Taylor 2001) and Target Diagram (Jolliff *et al.* 2009) are useful as they allow multiple metrics to be summarised and interrelated in a single plot.

Table 3.2: Summary of statistical tests and performance ratings for model evaluation

	Excellent	Good	Reasonable	Poor
Correlation coefficient	1-0.9	0.9-0.8	0.8-0.6	<0.6
Ratio of standard deviations	1-0.9, 1-1.1	0.9-0.8, 1.1-1.2	0.8-0.6, 1.2-1.4	<0.6, >1.4
NRMSD	0 – 0.05	0.05 – 0.1	0.1-0.3	>0.3
Ratio of means	1-0.9, 1-1.1	0.9-0.8, 1.1-1.2	0.8-0.6, 1.2-1.4	<0.6, >1.4

The statistical metrics chosen for the analysis are shown in Tab. (3.2). In addition to the correlation coefficient and NRMSD the ratio of model to observation for standard deviation and mean is utilised. The mean bias metric incurs the same intercomparison and evaluation issues as the RMSD and a ratio will still permit the assessment of spread between the model and observations. Rather than having a 'yes' or 'no' proposition for the evaluation criteria based on threshold values, a rating scale was developed (Tab. 3.2). The breadth of available statistical metrics (Stow *et al.* 2009) and absence of accepted standards to evaluate earth system models (Rykiel Jr 1996) inherently makes these evaluation criteria subjective and open to debate. However, by adopting a rating scale, a model need not be rejected if it fails any one criterion but rather it can be acknowledged as a

shortcoming. In addition, a model may accurately resolve the qualitative behaviour of a system without the quantitative accuracy. Therefore, skill metrics were viewed in tandem with spatial maps to see if major circulation patterns and features were resolved. Consequently, the strengths and weaknesses of the model were gauged and a domain of applicability constructed as this would better inform the types of hypotheses that could be tested and in-turn dictate the credibility of the results.

3.2.3 Front detection algorithm

Oceanic fronts are narrow regions of enhanced horizontal gradients in water properties and the task of front detection is analogous to edge detection in image processing (Oram, McWilliams & Stolzenbach 2008; Belkin *et al.* 2021). At present, a number of front detection algorithms have been developed for detecting sea-surface frontal features from satellite data. Varying in complexity, two main families of algorithms exist: a simple gradient-based approach (Canny 1986; Oram *et al.* 2008; Belkin & O'Reily 2009; Mauzole, Torres & Fu 2020) and a more advanced histogram analysis (Cayula & Cornillon 1992; 1995; Nieto, Demarcq & McClatchie 2012). Choosing a front detection algorithm is not trivial as no algorithm is perfect and each have their respective weaknesses. Gradient-based detection algorithms are highly sensitive to noise in the gradient field as the process of differentiation amplifies any noise present in the data. While for the histogram algorithms, the Cayula & Cornillon (1992; 1995) have a tendency to generate discontinuities along fronts, even with the implementation of contour following algorithms. Some of these issues can be resolved with pre-processing imaging techniques (Miltiadou *et al.* 2018) but the most effective way to utilize any front detection algorithm is with proper calibration.

For the study, the gradient-based Canny algorithm (Canny 1986) was used. The front detection was applied to the SST field along the shelf region of the SBUS, herein defined as the area extending from the coast to the 500 m isobath, within the zoomed domain shown in Fig (1). Similar modelling studies investigating frontal dynamics in high-resolution numerical models such as Ren *et al.* (2021) in the South China Sea and Mauzole *et al.* (2020) in the California Current System have used gradient-based edge detection algorithms. The Canny method has two important steps: (1) the pre-processing filter and gradient estimation and (2) edge localization.

Pre-processing filter and gradient estimation

The first step is to apply an appropriate filter to smooth the input SST data and remove small scale features. A two-dimensional Gaussian filter, which is the directional derivative of the two dimensional Gaussian distribution is applied.

$$G = \exp\left[\frac{-(x^2+y^2)}{2\sigma^2}\right] \quad (3.2)$$

In Eq. (3.2), x and y denote the longitudinal and latitudinal position of the data while σ is the width of the Gaussian distribution and the controlling variable in the equation. Adjusting σ alters the length scale of the resulting gradients where increasing σ increases the number of neighbouring points used to determine the gradient at a single point. Oram *et al.* (2008) highlighted the relationship between the gradient length scale and the spatial resolution of the input data on the length scale of the detected frontal edges. Fig. (3.2) shows an example of different values of σ applied to a summer SST gradient field for the model. A σ value of one corresponds to a desired

edge scale of 3 km, while three and five correspond to 9 km and 15 km respectively. Without a filter (Fig. 3.2a), the gradients are sharp and it is difficult to distinguish individual frontal features. The value of σ in Fig. (3.2c) and (2d) is too high as it over smooths the gradient field; making the frontal features too broad. A value of one in Fig. (3.2b) makes the gradients less sharp but still allows for some individual frontal features to be identified. However, there is still some noise present and thus a σ value of two was used for the study which corresponds to a desired edge scale of 6 km and is similar to that used by Ren *et al.* (2021). Furthermore, the chosen length scale does permit the identification of sub-mesoscale ($\sim 1\text{-}10$ km) frontal features which are associated with elevated lateral dispersion and vertical velocities (Thomas *et al.* 2008; Ruiz *et al.* 2019; McWilliams 2021).

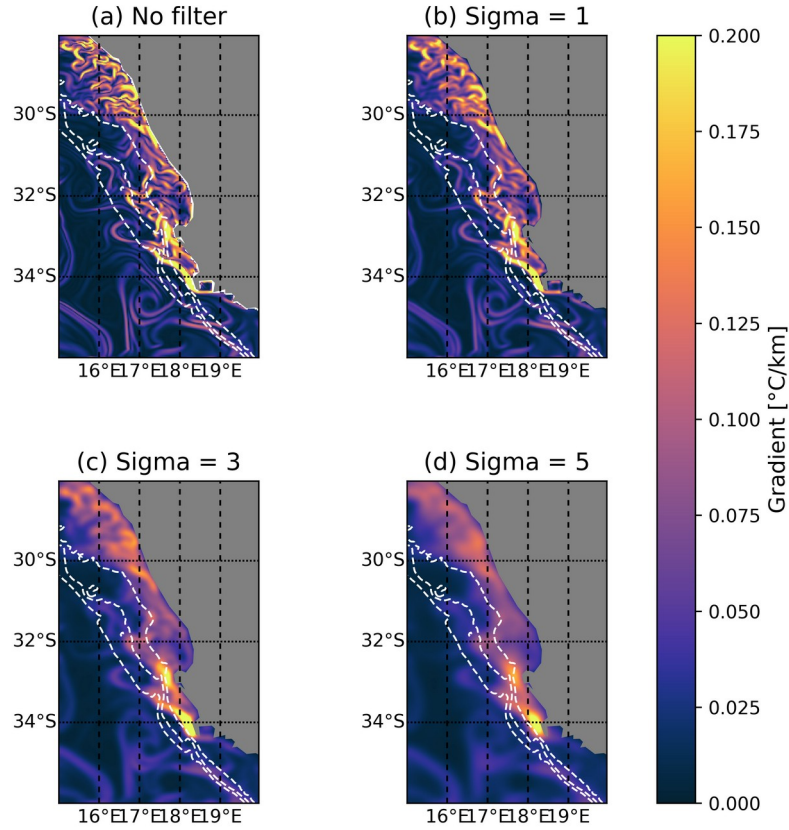


Figure 3.2: Snapshot of summer SST gradient field for different values of sigma

For a given SST field, T , after applying the Gaussian filter (Eq. 3.2), the next step in the Canny algorithm was to compute the x and y gradients of T . The discrete differential Sobel operator uses two 3×3 kernels which were convolved with T to produce T_x and T_y (Eq. 3.3).

$$T_x = \begin{bmatrix} +1 & 0 & -1 \\ +2 & 0 & -2 \\ +1 & 0 & -1 \end{bmatrix} * T ; T_y = \begin{bmatrix} +1 & +2 & +1 \\ 0 & 0 & 0 \\ -1 & -2 & -1 \end{bmatrix} * T \quad (3.3)$$

The magnitude (T_{mag}) of the gradient could then be computed as:

$$T_{mag} = \sqrt{T_x^2 + T_y^2} \quad (3.4)$$

Edge Localisation

Once the Gaussian filter and gradient components were computed, the next step in the Canny algorithm was to identify significant edges in the gradient maps, T_{mag} (Eq. 3.4), using threshold values. Thresholding is performed with hysteresis, where gradient magnitudes greater than the upper threshold are given a value of one, as are all connected points of gradient magnitude greater than the lower threshold, while all other values are set to zero. The choice of thresholds will implicate the nature of the detected fronts. For SST, which varies spatially and temporally, absolute threshold values were not set as this would affect the front detection in different seasons. Instead, as many authors do (example: Oram *et al.* 2008; Ren *et al.* 2021), a global histogram of gradient magnitudes is constructed from T_{mag} with the lower and upper thresholds corresponding to the 80th and 90th percentiles. In so doing, each individual gradient map corresponding to a particular day is assessed relative to its own gradient distribution, implying that the threshold values for detection will change for each day. Consequently, at the seasonal scale, the more intense cross-shore SST gradients along the shelf in summer will result in the absolute values of detected frontal features being significantly greater than those in winter (Supp. 3.2S). Once the edges have been identified, a binary edge map is created.

Front detection comparison against observational data

To evaluate whether the frontal positions in the model were correctly resolved, the frontal edges were compared against those computed for the OSTIA product at the seasonal scale. The comparison would be a qualitative assessment to see if the frontal structures were similar between the model and observations.

Frontal diagnostics

From the edge maps, the number of detected fronts (N), average length (treated as the number of pixels) and mean offshore distances were calculated. When computing the mean offshore distance of fronts for a given edge map, the distance of each individual point corresponding to a single front was computed parallel to the coast and then averaged to produce a mean distance ($\bar{x}_i$). Using Eq. (3.5), the mean of the average distances from all the fronts was then calculated to be the mean offshore distance.

$$\bar{x}_i = \frac{1}{N} \sum_{i=1}^N x_i \quad (3.5)$$

The advantage of using a numerical model was that the edge maps could be co-located with additional oceanographic variables, mainly vertical velocity, Brunt-Vaisala Frequency (BVF) and SST gradient; which were averaged in a similar manner to mean distance (Eq. 3.5). BVF is a measure of stratification, with units of [seconds⁻²], and was computed at a depth of 30 m (approximate base of observed climatological mean MLD from SHBML) from the daily model outputs using Eq. (3.6), where ρ is density.

$$BVF = -\frac{g}{\rho_0} \frac{\partial \rho(z)}{\partial z} \quad (3.6)$$

3.2.4 Spatial and frontal nomenclature for the SBUS

To refine the regional nomenclature description of the SBUS, the shelf region north of Cape Columbine was sub-divided into three sections with St Helena Bay (32.82°S: 31.5°S), Mid-Shelf (31.5°S: 30.3°S) and Namaqua (30.3°S: 28°S) being defined (Fig. 3.3).

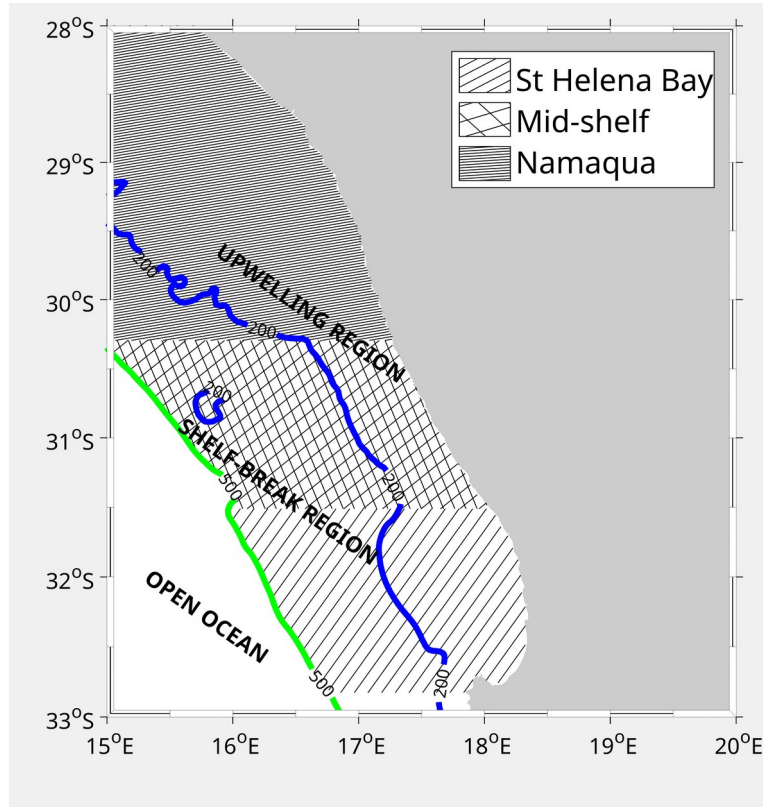


Figure 3.3: The three main spatial regions of St Helena Bay, Mid-shelf and Namaqua are shown with different hatching patterns. The blue line shows the 200 m isobath. The Upwelling Region of the shelf is defined as extending from the coast up to the main contiguous 200 m isobath. Between the 200 – 500 m isobath (green line) is the Shelf-break region. The 500 m isobath forms the boundary between the shelf and open ocean environments of the SBUS.

The shelf was delineated into two subdomains, where the predominantly on shelf subdomain region extending from the coast to the 200 m isobath was considered as the “Upwelling region” and the subdomain from 200 m – 500 m was referred to as the “Shelf-break region”. For the analysis it is assumed that the UF resides within the “Upwelling region” and the SBF within the “Shelf-break region” such that the average frontal diagnostics computed for these subdomains are representative of the physical dynamics of the respective fronts. From the work of Armstrong *et al.* (1987) and Brink (1987), the UF in the SBUS is conventionally defined as a seasonal frontal feature that forms in summer as a result of upwelling-favourable winds. Consequently, no true UF can be present in winter due to the absence of upwelling-favourable winds. To avoid the creation of more jargon, the term UF during winter is broadened to simply encompass the major jet feature(s) present within the defined Upwelling region of the shelf.

3.2.5 Finite Time Lyapunov Exponent

For time dependant non-autonomous systems, such as hydrodynamic models, the identification of transport barriers can be achieved by studying the finite-time Lyapunov exponent (FTLE) fields (Haller 2001; Haller 2002; Shadden, Leiken & Marsden 2005; Haller & Beron-Vera 2012; Van Sebille *et al.* 2018). Beginning with Eq. (3.7), let a particle's position be defined as:

$$\dot{\mathbf{x}} = \mathbf{v}(\mathbf{x}, t) \quad (3.7)$$

where $\mathbf{v}=(u, v)$ and $\mathbf{x}=(x, y)$. Integrating Eq. (3.7), gives the flow map: $F_t^T: \mathbf{x}(t) \rightarrow \mathbf{x}(t+T)$ which maps fluid particles trajectory from the initial time t to the integration time length T . From the flow maps, the finite-time version of the Cauchy-Green (C) strain tensor can be computed.

$$C(\mathbf{x}, t: T) = [\nabla_H F_t^T(\mathbf{x})]^T \nabla_H F_t^T(\mathbf{x}) \quad (3.8)$$

From Eq. (3.8), $[A]^T$ is the transpose of the matrix and C measures the magnitude of the growth in separation of infinitesimal perturbations in the initial position space. The FTLE is a measure of the maximum exponential divergence of nearby trajectories over a finite time-interval (Shadden, Dabiri & Marsden 2006) and is given by Eq. (3.9):

$$FTLE = \frac{1}{|T|} \ln \sqrt{\lambda_{\max}(C)} \quad (3.9)$$

where $\lambda_{\max}$ is the maximum eigenvalue of C over the chosen finite integration time T .

Using the scripts of Onu, Huhn & Haller (2015), a two-dimensional auxiliary grid was generated, with a spatial resolution 5 times that of the surface velocity data in order to discretize the computation of the FTLE field for the SBUS (Fig. 3.1). The robustness of FTLE is that it is independent of time and accounts for the integrated effect of the flow. From Haller (2002) and Shadden *et al.* (2005), Lagrangian Coherent Structures (LCS) are identified as ridges in the FTLE field and for forward (backward) integration correspond to stable (unstable) manifolds. These material lines in a flow, which are linearly stable or unstable for longer periods than neighbouring points, reveal the underlying flow geometry and correspond to distinct barriers that cannot be crossed by ideal tracers. (Haller & Yuan 2000; Haller 2001). It is expected then that prominent LCS should be associated with intense frontal features. For $T > 0$, the LCS are repelling such that trajectories on either side of the stable manifold separate faster than arbitrary points whereas $T < 0$ reveals unstable manifolds which correspond to regions of enhanced accumulation and are important for transport but were not considered in the work.

Applying the thresholds of Dauch *et al.* (2018) and following a similar methodology to Giudici *et al.* (2021), the LCS were identified as ridges > 0.5 from the FTLE fields after being normalised by the maximum FTLE value. The choice of threshold does not effect the position of the LCS, only their width. In general, longer integration times reveal more of the LCS but limits to the integration time include the computational expense and spatial domain limits. Consequently, an integration time of 30 days was deemed appropriate to capture the LCS, with the seasonal summer and winter

average FTLE fields computed over 1 January 2014 – 31 December 2017 using a 10-day sliding window to account for variability in the initial velocity states.

3.3 Results

3.3.1 Model Evaluation

Overall model performance

In evaluating the model's performance seasonally as well as interannually, a combination of *in situ* and satellite data products (Tab. 3.1) were used to assess the model's performance relative to the subset of the domain shown in Fig. (3.1) using the skill metrics and rating system in Tab. (3.2). Collectively, Fig. (3.4) and Tab. (3.3) summarize the model's performance relative to the statistical tests and scores shown in Tab. (3.2) for the various observational data sets.

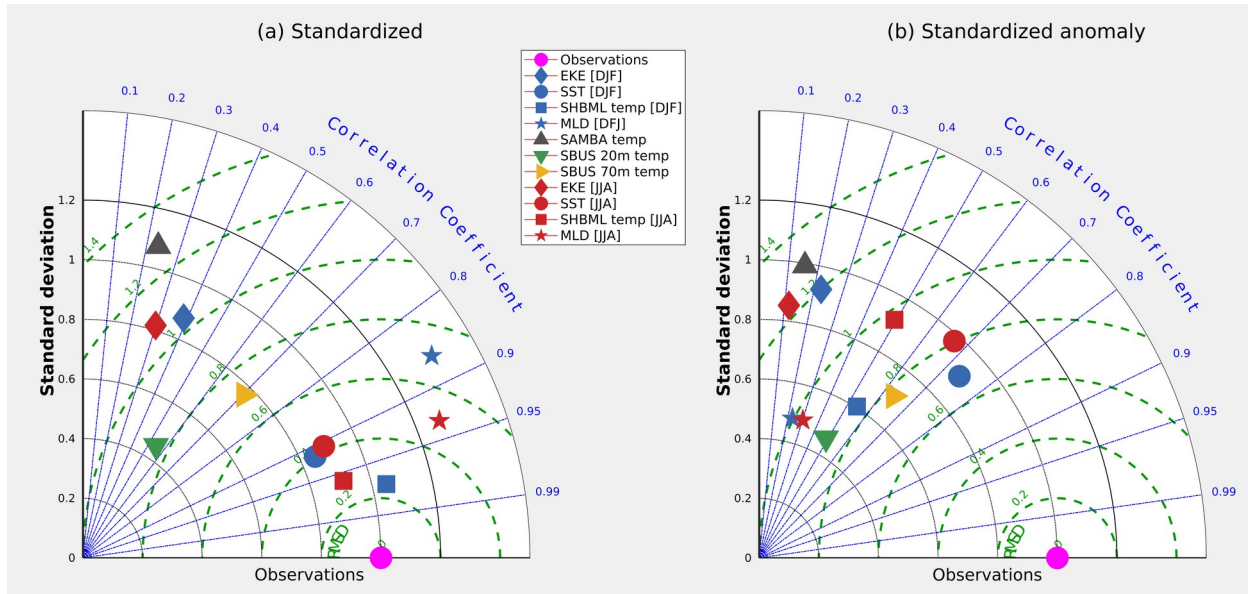


Figure 3.4: Taylor diagram for model and observational data sets for (a) Standardized and (b) Standardized anomaly. Radial distance represents the ratio of simulated to observed standard deviations, and azimuthal angle represents model-data correlation. Green arcs represent centered root mean square difference between model and data. The observations are shown on the bottom x-axis with the model's performance relative to the observations shown by the scatter. In the legend 'EKE' is that computed against the CMEMS altimetry product, 'SST' relative to OSTIA and 'MLD' against the SHBML.

The Taylor Diagrams in Fig. (3.4) show the performance of the model relative to the observational data sets. For comparative purposes, the model statistics are standardized by the seasonal mean and standard deviation of the observations (Fig. 3.4a) with the standardized anomaly statistics computed similarly once the monthly and seasonal (for MLD and SHBML temperature) means were subtracted (Fig. 3.4b). Juxtaposing the two Taylor Diagrams, the model performance is poorer for the anomalies; however, this is to be expected as the spatial resolution of the model cannot resolve all variability present. The dichotomy in the statistical scores between the absolute and anomaly fields raises an important point of contention as to which to use for the model evaluation scoring. For transparency, it was necessary to generate both sets of statistics but for simplicity, the model will be evaluated against the absolute fields (Fig. 3.4a). Absolute fields are more favourable for the model scoring as the assimilation of forcing data is inherently derived from observations.

From Fig. (3.4a), the model shows excellent correlations and good standard deviation (STD) ratios for the summer and winter SST. Similarly, excellent correlations and STD ratios are seen for the SHBML in both seasons. Relative to the mooring data, the correlations are poor to reasonable with only the STD ratio for the SAMBA data being excellent while the 20 and 70 m moorings are again poor to reasonable. This suggests that the model is able to resolve the temperature dynamics, both in the surface and vertical plane, but in the near coastal region, the dynamics are not as well captured and this may be due to the low spatial resolution of the atmospheric forcing along the coast. The EKE correlations are poor in both seasons but the STD ratios are good which suggests that the model is able to capture similar magnitudes of variability to the observations for mesoscale features. For MLD, the STD ratios are reasonable for summer and winter with the correlations being good and excellent respectively. Noticeably, the RMSD differs between summer and winter suggesting that the patterns of MLD are better resolved in winter.

In Fig. (3.4b), showing the anomalies, the statistical scores worsen for SST, SHBML and MLD relative to Fig. (3.4a). The decline in correlation showcases the difficulty in capturing variability while the increase in RMSD and decline in STD indicate that the intensity of variability differs between the model and observations. The similar STD scores for SAMBA as well as the summer and winter EKE relative to Fig. (3.4a) is a positive as it shows the model's ability to resolve similar magnitudes of variability about the mean to the observations. For MLD the STD ratios decline while the difference in RMSD between summer and winter for the SHBML suggest that the variability is better captured in summer than winter.

Table 3.3: Summary of model scores across the different observational data sets

	Observations	EKE [DJF]	SST [DJF]	SHBML temp [DJF]	MLD [DJF]	SAMBA temp
Correlation	1.00 - Excellent	0.39 - Poor	0.92 - Excellent	0.97 - Excellent	0.87 - Good	0.23 - Poor
STD ratio	1.00 - Excellent	0.87 - Good	0.85 - Good	1.05 - Excellent	1.35 - Reasonable	1.08 - Excellent
NRMSD	0.00 - Excellent	2.11 - Poor	0.04 - Excellent	0.07 - Good	0.34 - Poor	0.10 - Reasonable
Mean ratio	1.00 - Excellent	0.89 - Good	1.02 - Excellent	1.14 - Good	1.06 - Excellent	1.17 - Good
	SBUS 20 m temp	EKE [JJA]	SST [JJA]	SHBML temp [JJA]	MLD [JJA]	SBUS 70 m temp
Correlation	0.54 - Poor	0.30 - Poor	0.91 - Excellent	0.96 - Excellent	0.93 - Excellent	0.70 - Reasonable
STD ratio	0.45 - Poor	0.82 - Good	0.89 - Good	0.91 - Excellent	1.28 - Reasonable	0.77 - Reasonable
NRMSD	0.11 - Reasonable	2.36 - Poor	0.04 - Excellent	0.07 - Good	0.23 - Reasonable	0.05 - Good
Mean ratio	1.16 - Good	0.93 - Excellent	1.01 - Excellent	1.07 - Excellent	2.40 - Poor	1.22 - Reasonable

Tab. (3.3) complements Fig. (3.4) with the addition of the mean ratios and NRMSD which cannot be scored from the Taylor Diagrams. In Tab. (3.3), the STD ratios and correlation scores are shared

with Fig. (3.4a); with the remaining statistics computed from the unstandardized data. For the temperature dynamics, the model does an excellent job capturing the area averaged SST mean state and residuals for summer and winter. This is also seen for the SHBML where the NRMSD is good while the mean ratios are good and excellent for summer and winter respectively. Not seen in Fig. (3.4a), Tab. (3.3) shows some positives in the model's performance relative to the mooring data. For the SAMBA and 20 m mooring, the mean ratio is good while the reasonable NRMSD suggest some ability to resolve variability and patterns in the nearshore environment; which is difficult to capture. For the area averaged EKE, the model mean ratio is good and excellent for summer and winter respectively while the NRMSDs are poor. This shows that while the model is able to capture the magnitude of EKE, it does not necessarily capture the exact spatial characteristics of the sub to mesoscale features that generate the turbulent field that is measured by EKE. The opposite is seen for winter MLD where the mean state is poorly resolved but the reasonable NRMSD suggests that some of the MLD spatial characteristics are captured. Although, the high mean ratio of 2.40 for winter shows the MLD is significantly deeper than in the observations. Furthermore, in summer, the NRMSD is poor but the mean ratio is excellent.

Across all the different observational data sets, the model is able to score a 'good' or 'excellent' for one or more statistical metrics. Noting the weakness in the nearshore region and the winter MLD, the model does an excellent job in resolving the spatial and temporal temperature dynamics, especially for the surface, as well as capturing the intensity of EKE. Reducing a model's performance down to a single value allows for a quick assessment but inherently masks spatial biases and discrepancies. Therefore, the overall evaluation shows that the model is capable of broadly representing the physical dynamics of the SBUS but spatial comparisons are needed to refine this conclusion.

Climatological State Evaluation

Beginning with SST, the model is able to capture the mean summer (December, January, February: Fig. 3.5) and winter (June, July, August: Fig. 3.6) patterns. In summer, intense SST gradients between Cape Point and Cape Columbine define the CP jet while along the shelf, distinct upwelling cells are visible off the Cape Peninsula, Cape Columbine and Namaqua and the warm offshore waters are associated with Agulhas leakage (Fig. 3.5a, b).

In the model, the offshore extent of cool shelf waters reside closer to the shore, limited to the 200 m isobath, whereas in the observations, the cool upwelled waters permeate the entire shelf (Fig. 3.5a, b). The average summer bias for the domain is 0.35 °C with considerable spatial variability present as a strong 0.72 °C positive SST bias exists over the shelf and a small negative bias of < 0.1 °C is seen in the open ocean regions (Fig. 3.5c). Between the 200 and 500 m isobath the positive bias is similar, suggesting that the alongshore intensity of the shelf-edge jet should not be significantly affected. However, along the coast, a negative bias exists off the Cape Peninsula which may lead to a weakening of the jet feature in that region. From Fig. (3.5d), the average correlation is 0.74 for the domain and over the shelf it is 0.76.

In winter, the absence of upwelling-favourable winds and reduced solar insolation leads to a weaker CP jet and no upwelling cells along the shelf (Fig. 3.6a, b). Between the 200 m isobath and the coast, the model shows slightly warmer waters relative to the observations (Fig. 3.6c). A positive bias of 0.27 °C exists over the shelf and a small negative offshore bias of ≈ -0.1 °C is present. From Fig. (3.6c), the transition from negative to positive bias along the shelf corresponds with the

approximate location of the shelf-edge jet. Consequently, the intensity of the jet is likely to be weaker. Comparing Fig. (3.6d) with Fig. (3.6c), the regions of lowest correlations along the shelf correspond well with the transition region of negative to positive bias. It may be that the weakening of the shelf-edge jet results in fewer baroclinic instabilities and the associated mesoscale features and frontal filaments associated with the jet. This results in the correlation for the shelf region being lower than in summer, at 0.66. The dichotomy in the performance of the model between the shelf and offshore region contrast with the initial ‘excellent’ statistical score in Tab. (3.3). As expected, the spatial comparisons gave a more refined and in-depth evaluation. From Fig. (3.5) and Fig. (3.6), the model is able to resolve major seasonal SST features as well as reasonably capture the low frequency intra-seasonal and interannual variability; however, the prominent shelf based SST biases warrant consideration when examining the frontal diagnostics.

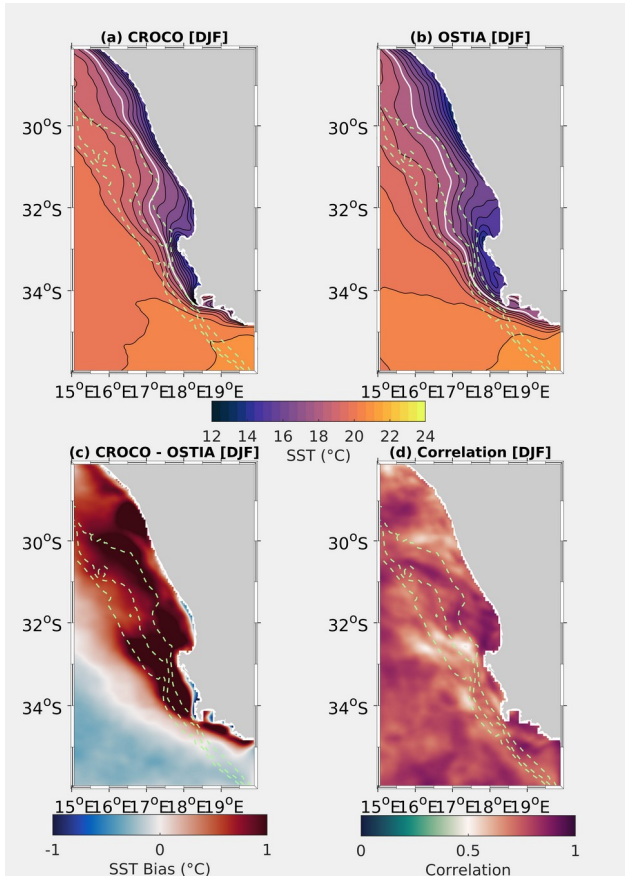


Figure 3.5: Mean summer state over Jan 2004 - Dec 2018 in the (a) model, (b) OSTIA. (c) shows the mean bias while (d) is the monthly anomaly correlations. Green dotted lines demarcate the 200, 300 and 500 m isobaths while the white line shows the 18 °C isotherm

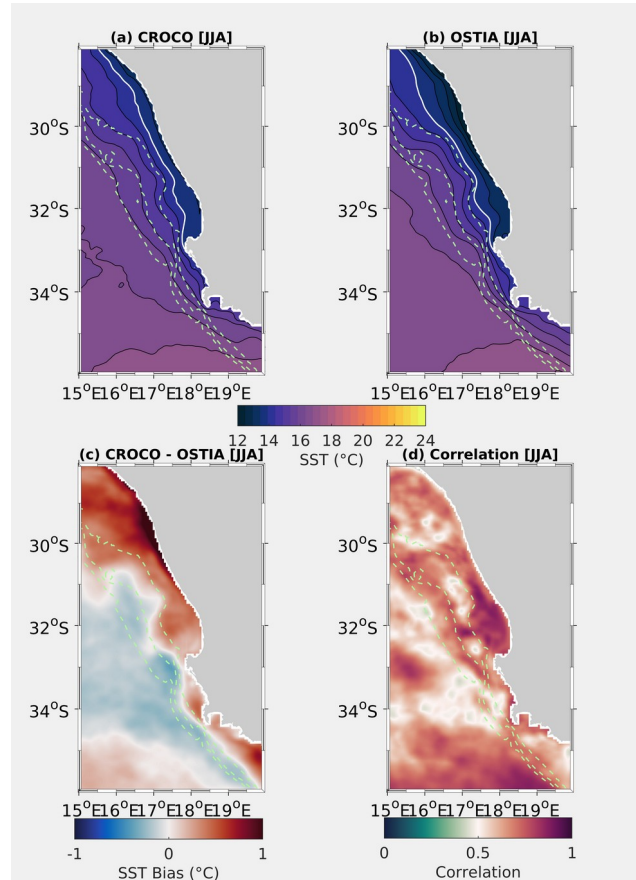


Figure 3.6: Mean winter state over Jan 2004 - Dec 2018 in the (a) model, (b) OSTIA. (c) shows the mean bias while (d) is the monthly anomaly correlations. Green dotted lines demarcate the 200, 300 and 500 m isobaths while the white line shows the 14°C isotherm

In Fig. (3.7), the major seasonal shelf and oceanic circulation features are evident such as the intense shelf-edge jets, a cyclonic circulation within St Helena Bay and high offshore EKE associated with Agulhas leakage. A marked deviation from the satellite observations is evident in the open ocean region of the southwest corner of the domain. From the EKE (Fig. 3.7 top), SSH patterns (middle) and geostrophic currents (bottom), it appears to be tight anti-cyclonic feature which is not present in the observations. As the focus of the study is on the shelf region of the

SBUS, this anomalous open ocean circulation pattern does not significantly impact the shelf region of interest; however, it does place a limitation for interrogating open-ocean dynamics north of Cape Columbine.

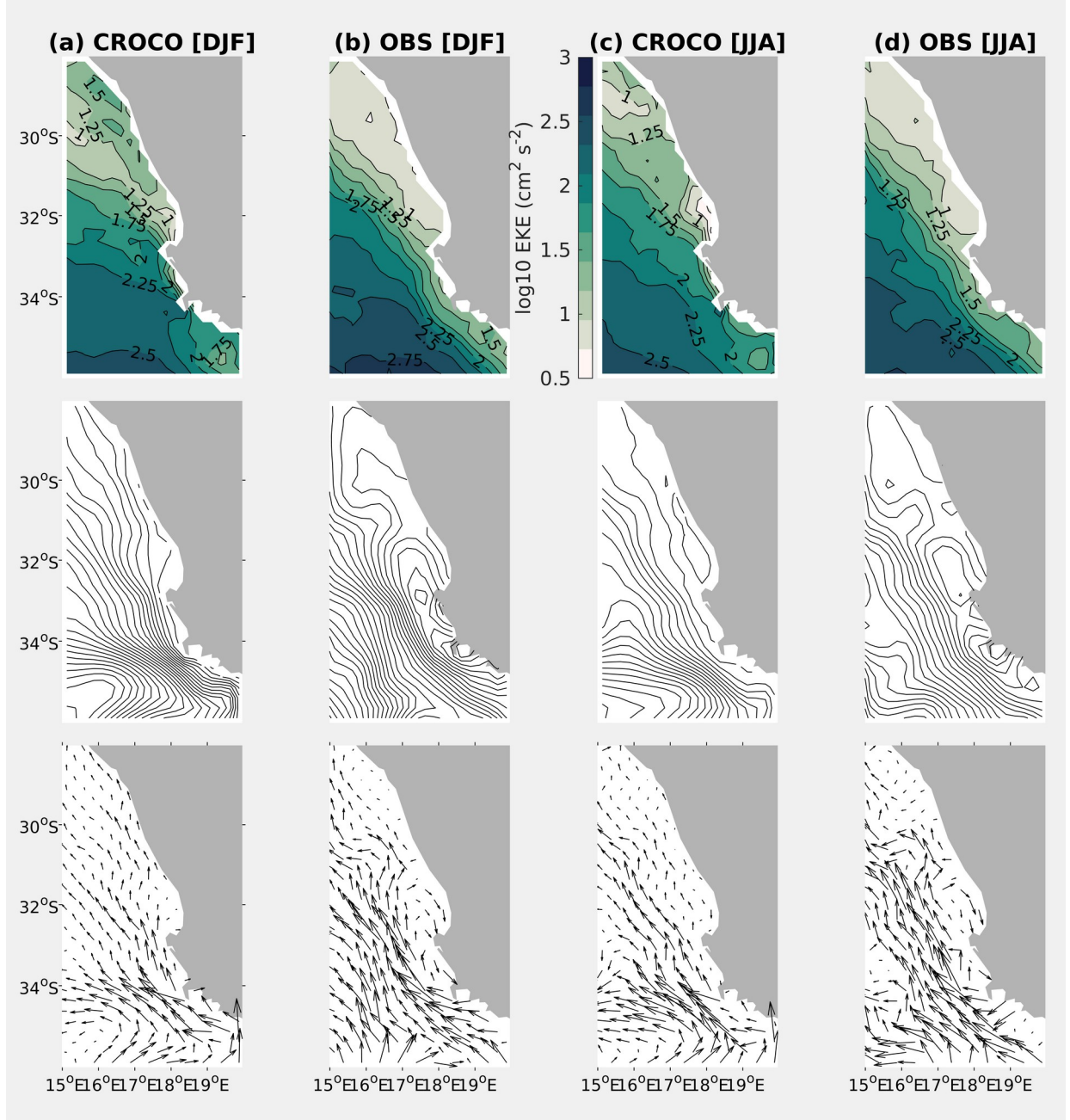


Figure 3.7: Model vs observations for average summer and winter: EKE (top row) , SSH (middle) and geostrophic currents (bottom) in the SBUS. Summer is shown for the first and second column while winter is in the third and fourth.

Contrasting the spatial patterns of EKE along the shelf, the model shows higher EKE than the observations in both seasons. Off Cape Point and Cape Columbine, the higher EKE in the model suggests a more energetic Cape Jet in closer proximity to the coast which is explained by the summer SST biases (Fig. 3.5c). The patterns of SSH confirm a more intense summer jet, whereas in winter, the model does resolve the offshore migration of the jet but not to the same degree as the

observations which may explain the higher EKE off Cape Point within the model. North of Cape Columbine, the isolines indicate a surface cyclonic circulation within St Helena Bay that is prevalent in both summer and winter within the observations. The model captures the feature in summer, although it's meridional extent is less than that seen in the observations, whereas in winter, the feature is well resolved within the bay, having a more narrow and elongated structure. The geostrophic velocities indicate weaker currents along the shelf for both summer and winter in the model compared to the observations.

Situated north of Cape Columbine (Fig. 3.1), the across-shelf sampling stations of the SHBML allow for the seasonal interrogation of the across-shelf temperature patterns along the SBUS shelf. Fig. (3.8) shows the average summer (Fig. 3.8a, b) and winter (Fig. 3.8d, e) across-shelf temperature sections of the SHBML for the model and observations. While the area-averaged evaluation metrics (Tab. 3.3) suggest that the model does an excellent job, a significant bias is present for both summer and winter. In summer, the model (Fig. 3.8a) has uptilting isotherms towards the coast, indicating an equatorward flow in both the offshore and nearshore region. The observations similarly show uptilting isotherms offshore but not so evidently closer to the coast. The slope of the isotherms/isopycnals appear to be steeper for the observations, particularly over the shelf-edge and mid-shelf region, suggesting that the modelled currents are weaker. This is consistent with the geostrophic currents seen along the shelf in Fig. (3.7) as well as the warm bias seen in Fig. (3.8c). The warm bias (2 -3 °C), prevalent in the upper 100 m, intensifies and deepens towards the coast with it's maximum subsurface extent existing approximately 50 km offshore.

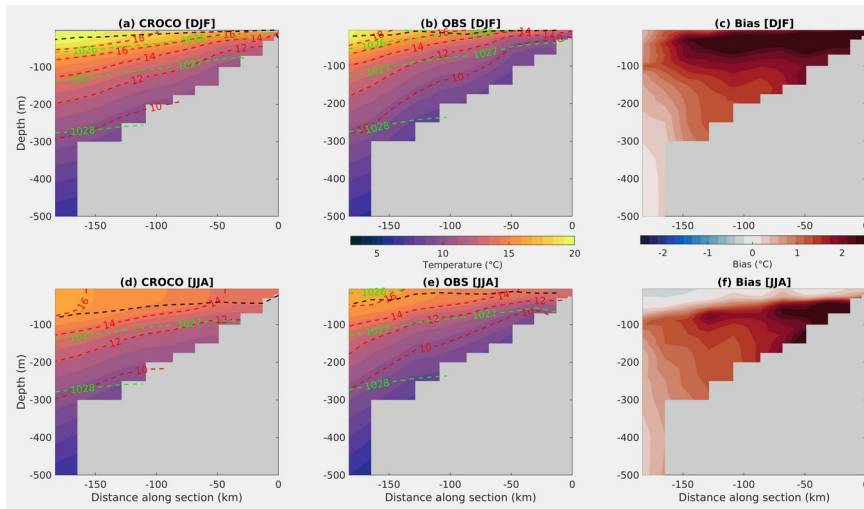


Figure 3.8: Temperature sections across the SHBML for the average summer conditions in the (a) model, (b) observations and (c) bias (model - obs). The average winter state is shown in (d) model, (e) observations and (e) bias. Green dotted lines indicate isobars and the red lines show the isotherms. The black dotted line is the MLD.

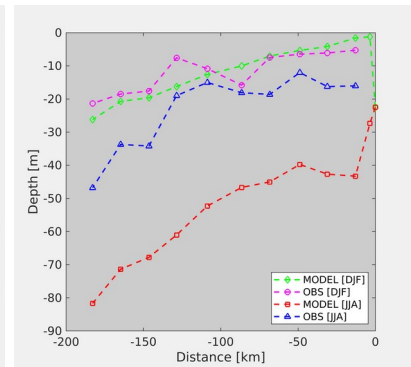


Figure 3.9: Average summer and winter MLD's computed across SHBML

In winter, the isotherms and isopycnals in Fig. (3.8d, e), show surface equatorward currents in both the model and the observations while closer to the coast, the downtilting of the isotherms indicate a poleward flow. The slope of the isotherms/isopycnals is greater in the observations, suggesting a less intense surface equatorward jet as well as poleward undercurrent in the model. The deepening of the MLD during winter months is overestimated in the model (Tab. 3.3 and Fig. 3.8d, e) and it's position is roughly consistent with the transition from a small bias (negative in the offshore domain and positive over the shelf) at the surface to an intense warm bias at depth. Consequently, the

temperature biases observed in the SHBML sections (Fig. 3.8c, f) not only impact the strengths of the surface and poleward flows but also the depth of the MLD. Across the SHBML, the MLD is deeper in winter than summer and shoals towards the coast (Fig. 3.9) similar to other EBUS. Despite the large and particularly sub-surface positive temperature biases, the model is able to capture the equatorward jet features, as well as the nearshore poleward counter current.

Seasonal and interannual evaluation against moorings

The SAMBA M4 mooring, located along the shelf-edge off Cape Point (Fig. 3.1), gives ADCP data throughout the water column as well as bottom temperatures. From the current roses in Fig. (3.10), the model does a good job in capturing the direction and intensity of the surface jet in the upper 50 m although it is marginally weaker than the observations. This is most probably linked to the temperature bias along the shelf-edge, especially for winter, where the mooring's position corresponds well with the juxtaposition of the cool offshore bias and warmer nearshore bias (Fig. 3.6c). Below 250 m, a predominant poleward flow is evident in the model and the direction is nicely resolved but the intensity is weaker. Similar to the SHBML, the model has an average positive bias at depth (Supp. 3.3S) of 1.3 °C, recorded at 300 m. Consequently, the warm bias is likely affecting the steepness of the isotherms in the region, resulting in a weaker poleward return flow.

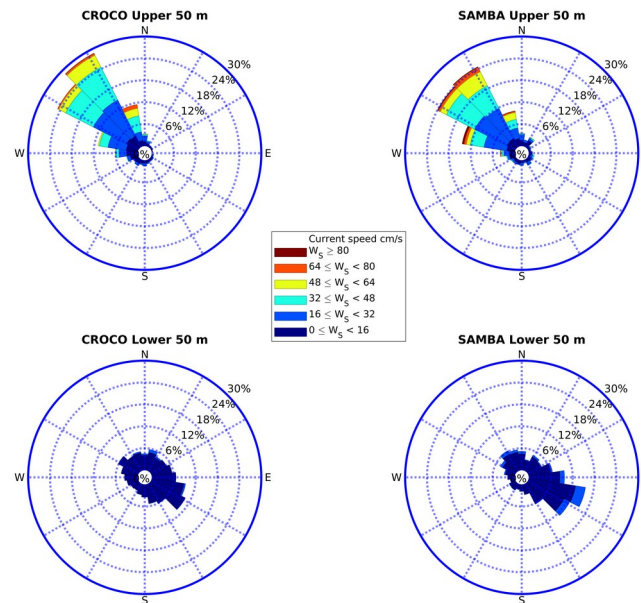


Figure 3.10: Current roses for the seasonal average currents in the upper and lower 50 m for the model and SAMBA M4 mooring. The spokes point in the direction of the current.

Further north along the shelf, the two shallow moorings north of Cape Columbine, at depths of 20 m and 70 m, have a similar warm bias (Fig. 3.11) of 1.9 and 2.2 °C respectively. For the 20 m mooring, the model consistently comes in and out of agreement with the observations, capturing the variability best in the late spring (October-November) and summer months whereas in winter, the model shows a consistent warm bias but does not always capture the observed variability. For the 70 m mooring, the model shows a consistent warm bias throughout the sampling period but does capture a lot of the variability during summer and winter. The initial skill metrics from Tab. (3.3) corroborate with Fig. (3.11) as the 70 m mooring had a reasonable correlation whereas the 20 m was

poor. For both moorings, the model did not capture the same degree of variability seen in the observations. Perhaps higher spatial and temporal resolution alongshore winds would resolve greater variability. Furthermore, the added variability in the nearshore region such as diapycnal processes as well as land-sea breezes (Fearon *et al* 2020) make resolving the 20 m dynamics more difficult for the model.

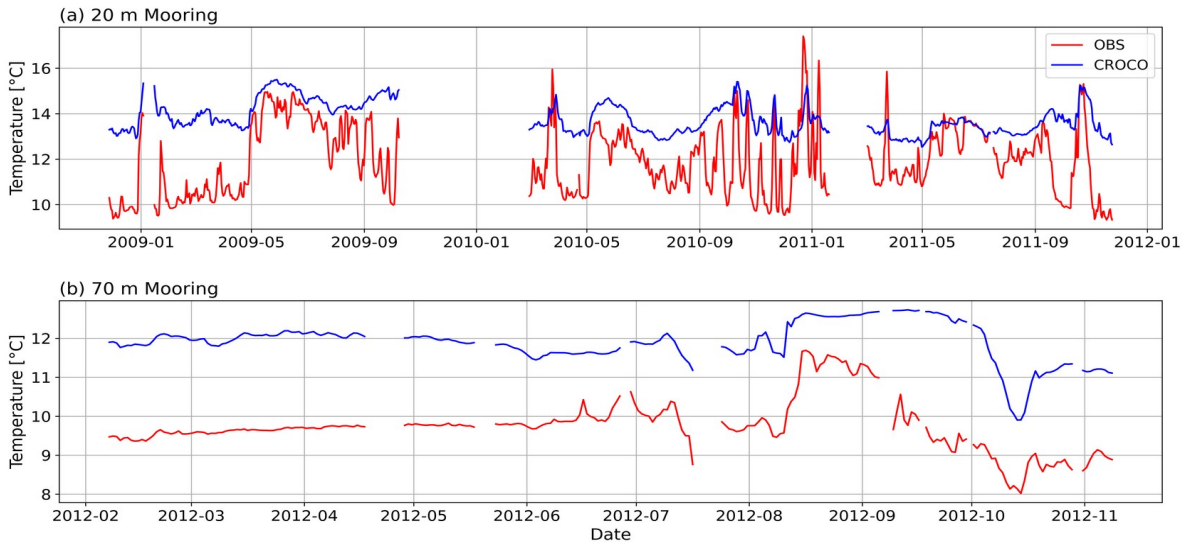


Figure 3.11: Mooring data for (a) 20 m (b) 70 m in St Helena Bay

Despite persistent temperature biases, both at the surface and with depth, the model is able to capture the key features applicable to this study including the equatorward and poleward flow regimes along the shelf as well as the cyclonic circulation within St Helena Bay. Additionally, it is able to capture low frequency intra-seasonal and interannual variability of SST as well as some of the nearshore temperature variability along the shelf.

3.3.2 Front detection comparison against observational data

The previous section concluded that the model is able to credibly capture and resolve the major physical oceanographic features of the SBUS, especially the intra-seasonal SST patterns (Fig. 3.5 and 3.6) and circulation dynamics (Fig. 3.7) along the shelf. Differences in the cross-shore SST patterns between the model and observations will impact the offshore positions of the various alongshore jets. Therefore, it is important to understand if and whether detected frontal features in the model differ significantly to those from OSTIA.

In Fig. (3.12), intense SST gradients (Eq. 3.4) are associated with the prominent shelf-edge jets as well as upwelling cells along the coast during summer. In summer, the broad spatial patterns of SST gradients are resolved along the shelf within the model (Fig. 3.12a) such as an intense CP jet as well as upwelling cells off Cape Columbine and Namaqua. Along the coast, more intense gradients are seen in the model compared to the observations, especially in St Helena Bay and off Cape Point, as a consequence of the nearshore cool bias and shelf warm bias (Fig. 3.5c). Calculating the average frontal gradients using the detected edges from Fig. (3.13a, b), the model had fronts with gradients of 0.11 ± 0.047 °C and the observations of 0.083 ± 0.024 °C. This means that the model resolves more intense fronts than the satellite product. Comparing the frontal patterns in Fig. (3.13a, b) for summer, the model simulates similar frontal patterns relative to the observations. North of Cape

Columbine, a long, near continuous front is found close to the coast although, in the observations, no front is detected for the St Helena Bay region and this may arise due to the cool nearshore bias present in the model. Furthermore, the model resolves the front associated with CP jet, although the position of the jet is found in closer proximity to the coast than in the observations.

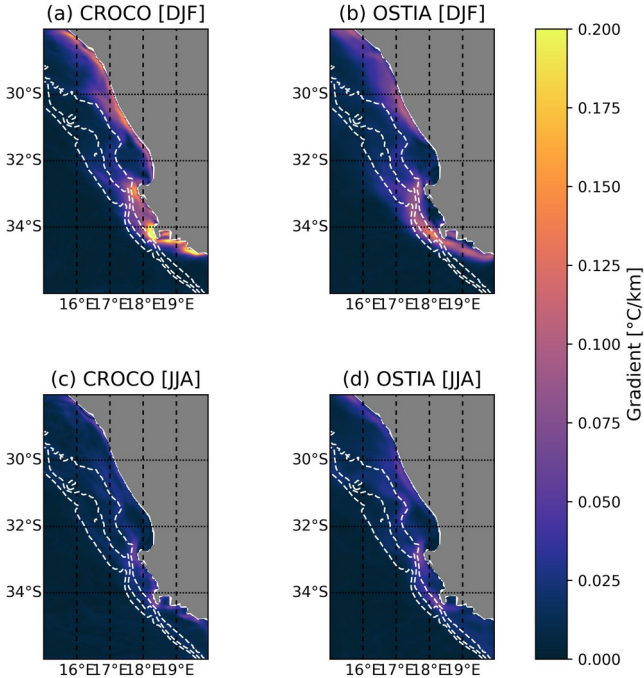


Figure 3.12: Average summer and winter SST gradient for the model and observations.

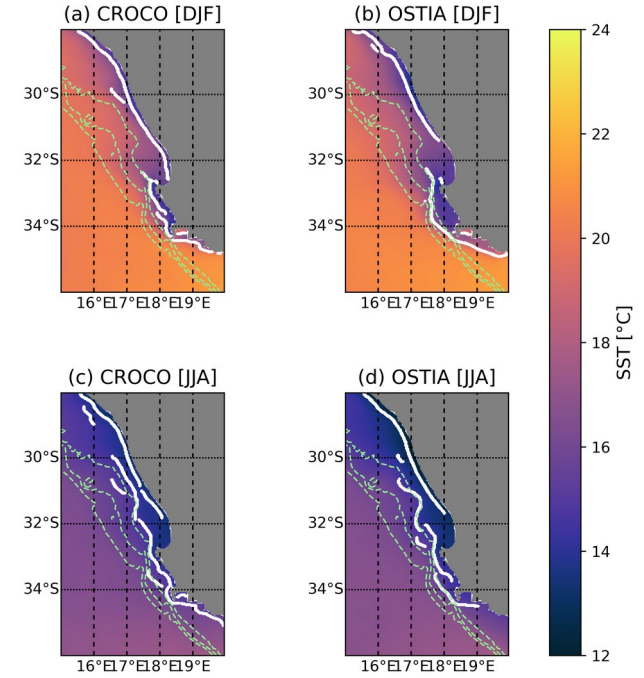


Figure 3.13: Average SST for summer and winter for the model and observations. White solid lines show the location of the frontal edges while the green dotted line indicate the position of the 200, 300 and 500 m isobaths

In winter, the absence of favourable upwelling winds and reduced insolation leads to weaker SST gradients for the region (Fig. 3.12c, d). The model shows moderate SST gradients in the nearshore off Cape Point and Cape Columbine as well as elevated SST gradients along the coast north of 31°S, although the intensity of these gradients are marginally weaker than the observations. This is confirmed when calculating the average frontal intensity which for the model was 0.035 ± 0.013 °C and for the observations, 0.041 ± 0.013 °C. In Fig. (3.13c, d), the model captures the position and morphology of the nearshore fronts along the Cape Peninsula and those north of Cape Columbine. More fronts are captured, relative to summer, although it is important to stress that the intensity of these fronts are significantly weaker. From Fig. (3.12) and Fig. (3.13), the model captures the seasonal frontal structures, resolving the frontal dynamics best along the shelf north of Cape Columbine. A facet of the cool nearshore bias in summer along the coast is more intense fronts within the model. Conversely, the warm nearshore biases for winter (Fig. 3.6c) result in less intense fronts. Acknowledging this fact, the gradient-based Canny algorithm can be applied to detect significant frontal features on the SBUS shelf from the model outputs.

3.3.3 Seasonal cross-shore mixing and frontal patterns in the SBUS

A stark contrast exists in the frontal behaviour and dispersion patterns between summer and winter (Fig. 3.14). During summer (Fig. 3.14a), a high and near continuous alongshore band of high frontal

probability runs the entire length of the coast as a result of the predominantly southerly winds that stimulate coastal upwelling. A distinct moderate band of contiguous probability exists north of St Helena Bay and a bulge in the frontal probability near 30°S corresponds to the region of intense offshore flow associated with the topographical steering of the surface jets. An offshore extension of high frontal probability off Cape Columbine corresponds to the presence of the Columbine upwelling plume. Within St Helena Bay, a near spherical region of low frontal probability is evident in the centre of the bay, confined between the upwelling plume and intense upwelling along the coast. The surface expression of the cyclonic circulation within the bay (Fig. 3.7) likely acts to isolate the centre of the bay; limiting frontal activity. Further offshore, between the 200 and 500 m isobath, the moderate probability corresponds with the position of the intense shelf-edge jet as it bifurcates north of Cape Columbine.

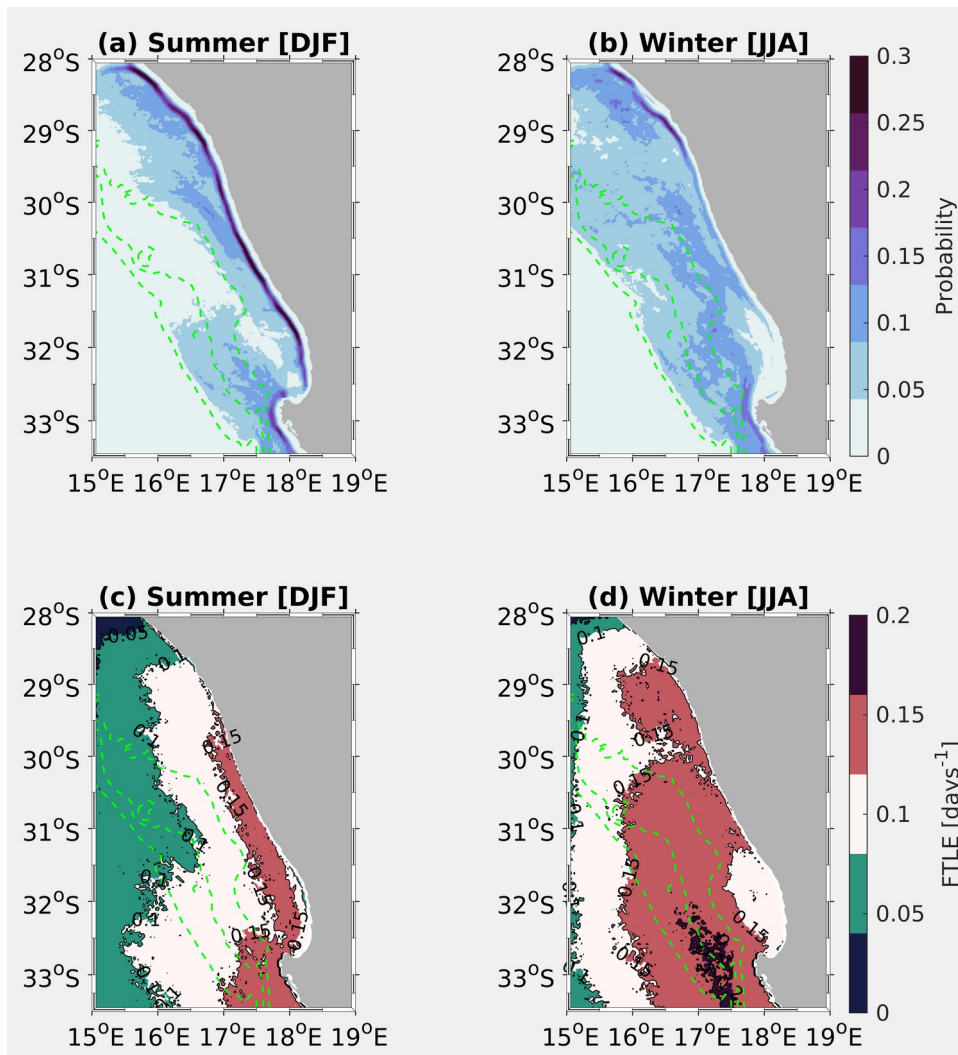


Figure 3.14: Seasonal summer and winter averages of (a-b) frontal probability and (c-d) FTLE (computed over Jan 2014 – Dec 2017). Green dotted lines demarcate the 200, 300 and 500 m isobaths.

The patterns of FTLE (Sec. 3.2.5) in Fig. (3.14c and d) highlight regions of enhanced exponential divergence in nearby particles. In Fig. (3.14c), the cross-shore patterns of FTLE broadly correspond with the summer frontal probabilities in Fig. (3.14a). Along the coast, a prominent tongue of contiguous high dispersion extends till just north of 30°S and thereafter, dispersion only remains

high very close to the coast. The offshore flow of the surface jets near 30°S possibly reduce the mesoscale activity within the upwelling region further north, resulting in weaker particle dispersion. In St Helena Bay, the region of high dispersion corresponds well with the high frontal probabilities along the coast (Fig. 3.14a). Further offshore, the lower FTLEs within the bulge of the 200 m isobath correspond with the location of low frontal activity seen in Fig. (3.14a). This supports the idea that the cyclonic circulation is isolating the centre portion of the bay which may elevate particle retention and accumulation in that region. Moving offshore, the shelf-edge jet is associated with weaker particle dispersion compared to the coastal region. This is likely due to wind induced offshore Ekman transport along the coast and the more intense coastal jets elevating particle dispersion.

The winter frontal patterns (Fig. 3.14b) differ to summer, most significant being the absence of the alongshore band of high frontal probability parallel to the shore and the offshore increase in frontal probabilities across most of the shelf. It is important to appreciate that the detected SST fronts in winter are defined by weaker SST gradients than those in summer (Supp. 3.2S). In winter, the predominantly northerly winds preclude upwelling along most the coast; however, an elevated band of high frontal probability exists north of 29°S, associated with the persistence of weak southerly winds in the region (Fig. 3.13a). In St Helena Bay, it is interesting that the low frontal probability region observed in the centre of the bay during summer (Fig. 3.14a) has increased in size and shifted to being closer towards the coast. The absence of intense upwelling along the coast and weakened plume off Cape Columbine allows the cyclonic circulation within the bay to expand in size (Fig. 3.7) and isolate a greater portion of the bay to frontal activity which may foster greater particle accumulation. Another major feature of winter is the offshore increase in frontal probability, especially in the shelf-break region and along the 200 m isobath. The offshore movement of fronts is related to the weakening cross-shore SST gradients and the broadening of homogeneous SSTs along the coast (Fig. 3.6).

From the winter dispersion patterns (Fig. 3.14d), the regions of high FTLE correspond well with the regions of high frontal probability, as seen in summer. A key difference; however, is the high dispersion rates exist over a much greater portion of the shelf and a distinct region of lower dispersion is seen within St Helena Bay. An increase in the dispersion rate across much of the shelf impacts the shelf-bound retention of particles because a similar cross-shore dispersion rate will allow particles to easily mix from the coast to the shelf-edge. Noting that the frontal probabilities increase offshore, it is fair to assume that the fronts in winter are not playing a major role in restricting the offshore advection of particles along the SBUS shelf. The weakened cross-shore SST gradients reduce the intensity of the alongshore fronts and jets (Fig. 3.7) and likely impact the cohesiveness of these structures. The lower dispersion rate in St Helena bay (Fig. 3.14d) arises because of the cyclonic circulation within the bay that acts to isolate the region; resulting in less frontal activity and particle dispersion. Consequently, the isolation of the bay can lead to elevated particle retention in the region.

In summary, it appears that the cohesive alongshore fronts and jets in summer are playing a major role in limiting the cross-shore extent of high dispersion between Cape Columbine and ~ 30°S to only the coastal margins. The seasonal erosion of intense cross-shore SST gradients in winter weakens the intensity of the alongshore jets. This promotes a much larger band of high dispersion over a greater region of the shelf suggesting greater cross-shelf losses of material.

Seasonal and Climatological Frontal Diagnostics

The seasonal shift in the intensities of the cross-shore SST gradients, alongshore wind regimes and stratification impacts the frontal environment and mixing patterns in the SBUS during summer and winter (Fig. 3.14). It is necessary then to understand how the physical oceanographic properties of fronts shift between summer and winter. Frontal diagnostics (Sec. 3.2.3) are interrogated at the monthly climatological scale (Fig. 3.15) and cross-correlations are employed to explore the relationships between the various frontal diagnostics relative to the alongshore wind forcing and stratification. (Fig. 3.16).

In Fig. (3.15) the average oceanographic properties of fronts within the shelf-break and upwelling region (Fig. 3.3) are collectively treated as being representative of the physical dynamics of the SBF and UF respectively. Both fronts show a similar seasonal cycle, being found closer to the coast in summer and further offshore in winter. Furthermore, they have a similar seasonal range in offshore distance of ~ 30 km. Along the shelf-edge, the prominent cross-shore SST gradients between the cool shelf waters and warmer open ocean waters set-up the SBF. The front remains quasi-stationary for most of the year and the largest shift in position occurs during October/November when the front retreats shoreward corresponding to the onset of southerly winds (Fig. 3.15a).

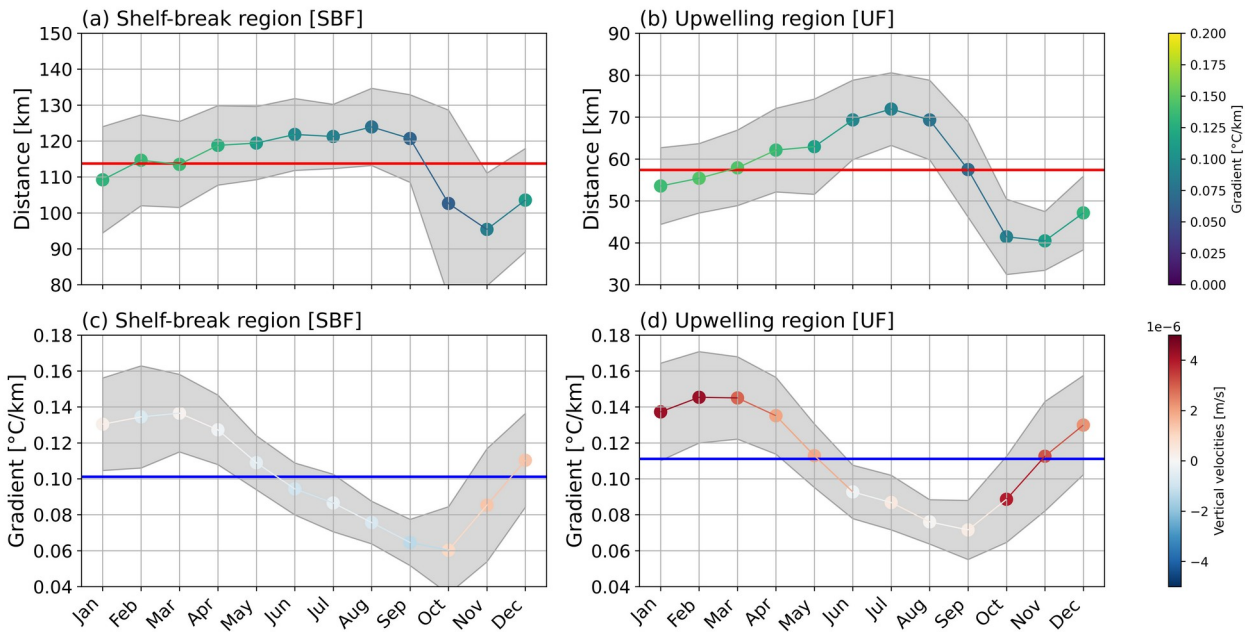


Figure 3.15: Monthly climatology for the shelf-break (a, c) and upwelling region (c, d) for the SBUS. The top row shows the evolution between distance offshore and SST gradient while the bottom row is the SST gradient and vertical velocities. The red horizontal line is the mean distance offshore of the respective front and the blue line is the mean gradient. The grey shaded region is ± 1 std.

As seen in Fig. (3.14), the UF is found closest to the shore in summer due to upwelling along the coast. The absence of upwelling-favourable winds precludes the formation of a traditional UF along the coast (Fig. 3.14b). Using the broadened definition from Sec. (3.2.4), the front shows an offshore migration in winter (Fig. 3.15b) associated with the absence of coastal upwelling and a homogenisation in the near-coastal SSTs. As expected, both the UF and SBF front are defined by stronger SST gradients in summer than winter, with similar ranges in magnitude seen for both

fronts. A stark contrast between the SBF and UF is seen for the monthly mean vertical velocities (Fig. 3.15c, d). The SBF experiences near persistent downwelling, only showing weak positive velocities from October to December, and this may be related to negative wind-stress curl along the shelf-edge. The UF shows a distinct seasonal shift from upwelling in summer and spring to downwelling conditions in winter, driven by the seasonal cycle of the alongshore wind patterns.

To investigate the relationships between the various frontal diagnostics and the major physical oceanographic drivers, the summer and winter correlations were computed from the daily model outputs. Only the 95% statistically significant relationships are retained (Fig. 3.16). For most of the diagnostics, though the correlations are statistically significant, they are very small but a number of prominent relationships are present.

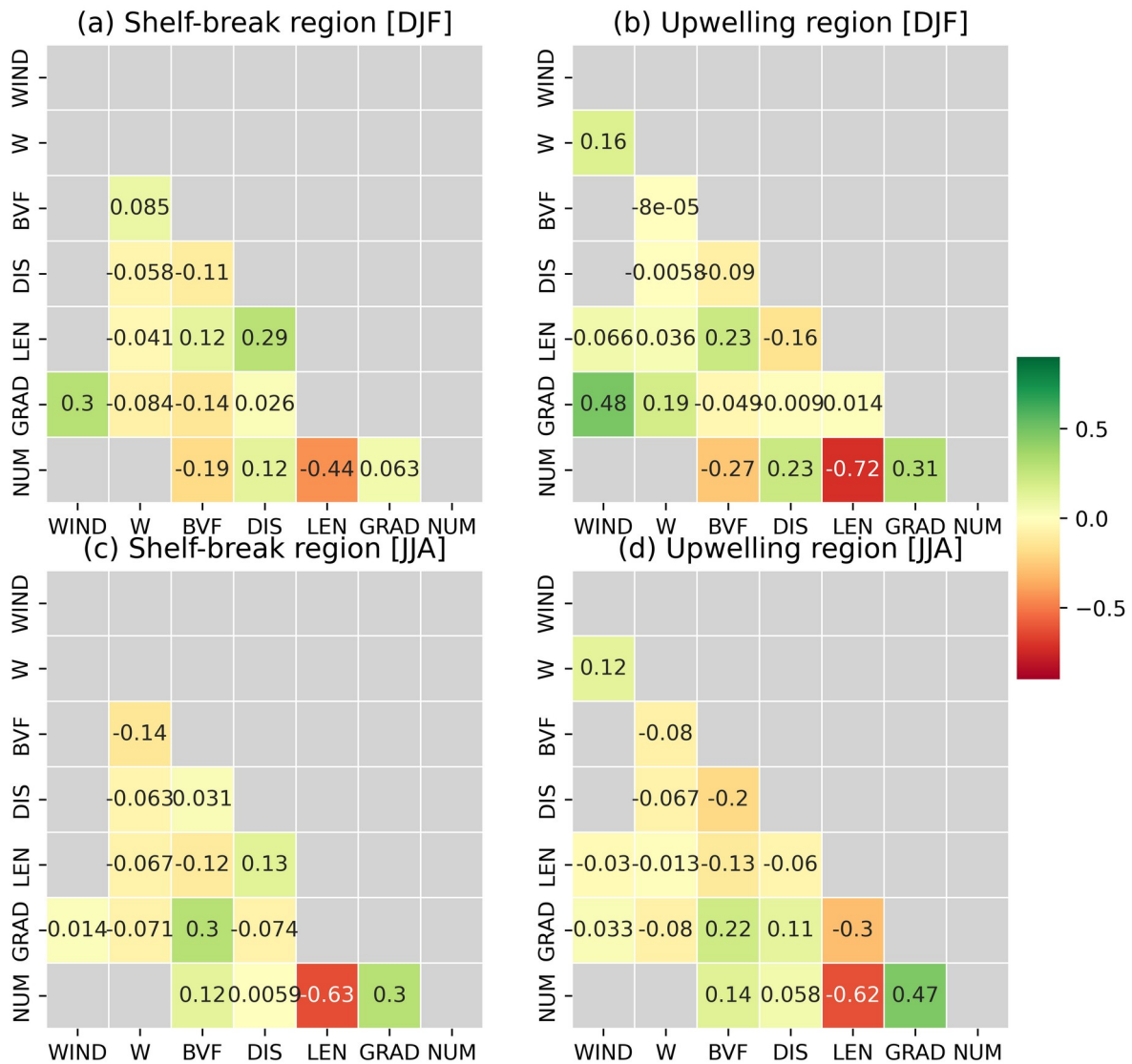


Figure 3.16: Correlations of anomalies for frontal diagnostics for summer and winter averaged within the upwelling and shelf region. WIND - alongshore wind component, W - vertical velocity, BVF - stratification, DIS - frontal distance offshore, LEN - length of the fronts, GRAD - SST frontal gradient and NUM - number of detected fronts. Grey blocks represent statistically non-significant relationships.

In summer (Fig. 3.16a, b), an expected strong negative relationship exists between the number of fronts and frontal length in the upwelling and shelf-break regions, with the upwelling region having the greatest correlation. The coupling between frontal gradient and alongshore wind is seen in the positive correlations within the upwelling (0.48) and shelf-break regions (0.3). The lower correlation in the shelf-break region is associated with the fact that the fronts are primarily driven by the prominent cross-shore SST gradients along the shelf-edge associated with Agulhas leakage and there is likely a positive relation between the strength of the alongshore winds and wind-stress curl; although this dynamic was not explored. For the upwelling region (Fig. 3.16b), evidence for the role of alongshore winds driving upwelling is seen in the small positive relationship with vertical velocity. A negative relationship exists between stratification and number of fronts detected in both the upwelling (-0.27) and shelf-break (-0.19) region. This relationship suggests that a weakening of stratification will subsequently enhance frontal activity along the shelf. Given the strong correlation between number of fronts and frontal length there is a converse correlation between stratification vs frontal length. Consequently, stronger stratification promotes the formation of longer fronts.

Though the correlations do not exceed ± 0.3 , it is interesting that the shelf-break region has a positive relationship between offshore distance and frontal length whereas the relationship is negative (-0.16) in the upwelling region. Longer and more cohesive fronts should elevate the shelf-bound retention of particles. The decrease in frontal length offshore in the upwelling region is likely due to the higher proportion of small scale jets and filaments which interrupt the generation of longer fronts. Conversely, in moving offshore in the shelf-break region, the constant source of warm water associated with Agulhas leakage, which helps drive the SBF, creates a more contiguous SST gradient along the shelf-edge which promotes the formation of longer fronts.

In winter (Fig. 3.16c, d), as expected, the most prominent correlation is again the negative relationship between frontal length and number of fronts. Of significance is the positive relationship between frontal gradient and stratification in the upwelling (0.22) and shelf-break (0.3) region. It is important to note that in summer, the most prominent correlation for gradient was against the alongshore wind component. Though the absolute correlations are smaller than in summer, a positive shift occurs in the relationship between stratification and number of fronts detected. Noting that the model resolves a significantly deeper winter MLD than the observations (Fig. 3.9); the significance of stratification in driving frontal variability within the model may be exasperated. Consequently, a better resolved winter MLD may elevate the importance of the alongshore wind component in affecting the intensity of fronts. In the upwelling region. A negative relationship is present for gradient and frontal length while the relationship between gradient and the number of fronts detected is positive and rather strong in both the upwelling and shelf-break region. This relationship seems intuitive as elevated frontal activity should result in the formation of more intense fronts.

From Fig. (3.16), a prominent, but rather intuitive, negative relationship exists between frontal length and the number of fronts detected. It is unclear; however, whether there is an absolute change in the magnitude of the variables between summer and winter. From Tab. (3.4), summer experiences statistically fewer but longer fronts than winter; although these fronts are defined by more intense SST gradients. The greatest seasonal change in frontal length is experienced in the upwelling region while the seasonal shift in the number of fronts is most prominent in the shelf-break region. Tab. (3.4) corroborates the seasonal frontal patterns in Fig. (3.14a, b) and furthers the idea that more cohesive and intense fronts in summer will have greater retentive abilities than the shorter and

weaker fronts in winter. Though the seasonal means are statistically different, qualitatively, the seasonal difference in frontal lengths are small. As alluded to from the dispersion patterns in Fig. (3.14c, d), it suggests that fronts in winter are not playing a major dynamic role in restricting the offshore advection of particles, regardless of their length or numbers.

Table 3.4: T-test assuming equal variance for summer and winter comparison of frontal length, number of fronts and SST frontal gradients

Upwelling region						
	Frontal length [Pixels]		Number of fronts		SST gradients [$^{\circ}\text{C}/\text{km}$]	
	Mean	p-value	Mean	p-value	Mean	p-value
Summer [DJF]	56.2	< 0.001	17.8	< 0.001	0.14	< 0.001
Winter [JJA]	48.8		20.3		0.085	
Shelf-break region						
	Frontal length [Pixels]		Number of fronts		SST gradients [$^{\circ}\text{C}/\text{km}$]	
	Mean	p-value	Mean	p-value	Mean	p-value
Summer [DJF]	46.7	< 0.001	7.7	< 0.001	0.12	< 0.001
Winter [JJA]	44.8		14.7		0.085	

Lagrangian Coherent Structures along the SBUS

Prominent ridges in a FTLE field can be treated as LCS which act as flow barriers and separate dynamically distinct regions. Representative snap shots of summer and winter LCS are shown in Fig. (3.17) and Fig. (3.18) respectively. In summer (Fig. 3.17), two distinct LCS are present along the shelf. Parallel to the coast, a long and near contiguous LCS exists north of Cape Columbine within the upwelling region. This nearshore structure is related to the prominent equatorward surface jets along the coast (Fig. 3.8) and neatly divides the shelf region in two. Further offshore, along the shelf-edge off Cape Columbine, another LCS is present which corresponds to the shelf-edge jet. The meridional extent and width of the LCS's differ between the two snap shots; related to the temporal variability in the respective velocity fields. Overlaying the frontal edges, they correspond well with the positions of the LCSs along the shelf; however, this is somewhat expected as fronts and jets are intrinsically related. The narrow and cohesive LCS complement the contiguous alongshore fronts suggesting that the alongshore fronts in summer can act as cohesive flow barriers that restrict the offshore advection of particles. The nearshore fronts likely have the greatest impact on the shelf-bound retention of particles as they are defined by more intense SST gradients and are longer (Tab. 3.4) than the fronts along the shelf-edge.

In winter (Fig. 3.18), the identified ridges in the normalized FTLE fields do not generate narrow and distinct LCS along the shelf like in summer. This does not suggest that LCS do not exist in winter; rather, they are not as stable or cohesive than they are in summer over the same temporal integration length. Looking closely, a cyclonic structure is present along the shelf-edge in Fig. (3.18a), likely due to an offshore eddy, and some narrow ridges exist along the coast within St Helena Bay (Fig. 3.18b) which could be considered as LCS. The near absence of distinct LCS along the shelf does not significantly impact the interpretation of Fig. (3.18) as it furthers the point that the fronts and jet structures in winter are not as cohesive or stable as in summer.

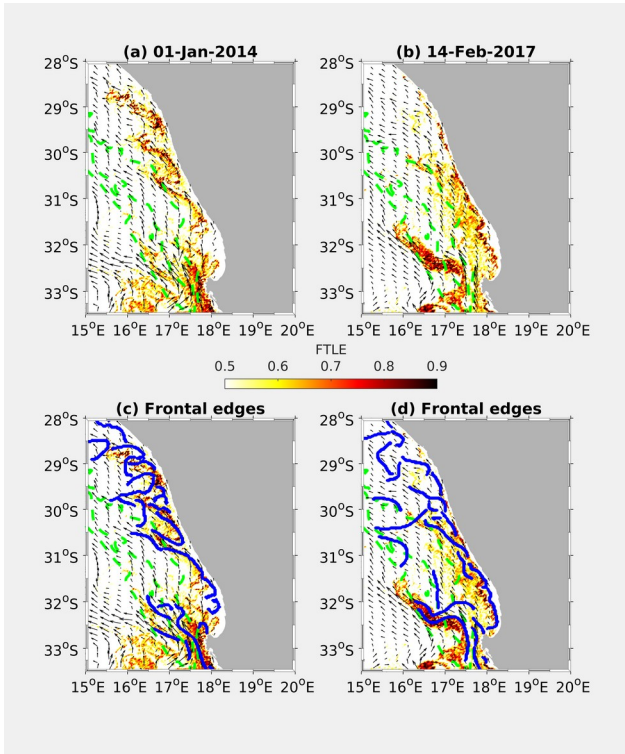


Figure 3.17: Snap shots of summer LCS, defined as ridges within the normalized FTLE field, overlaid with the surface velocity field after $T = 30$ days (a,b). In (c,d), the detected frontal edges are shown as blue contours relative to the above LCS. Green dotted lines demarcate the 200, 300 and 500 m isobaths.

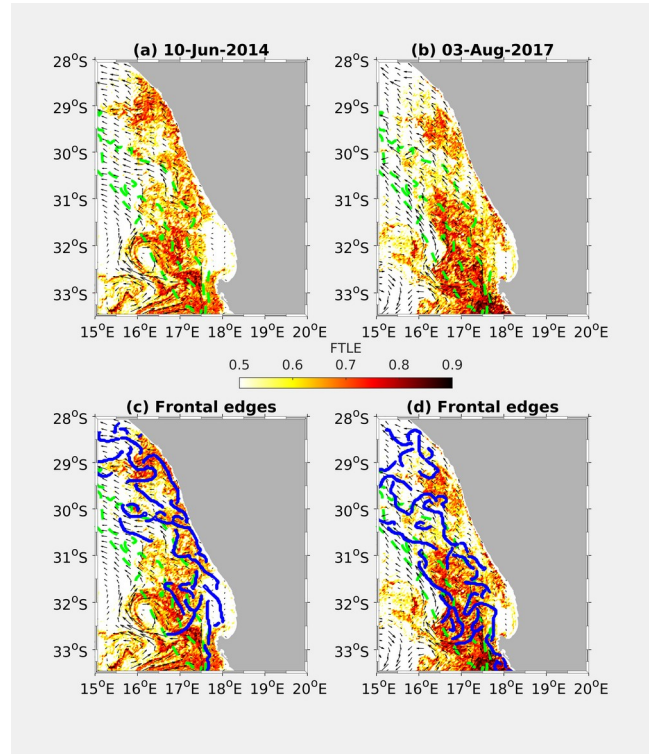


Figure 3.18: Snap shots of winter LCS, defined as ridges within the normalized FTLE field, overlaid with the surface velocity field after $T = 30$ days (a,b). In (c,d), the detected frontal edges are shown as blue contours relative to the above LCS. Green dotted lines demarcate the 200, 300 and 500 m isobaths.

From Fig. (3.18a ,b), the broad regions of high FTLE across the shelf elevate the cross-shore mixing of particles while a noticeable absence of high dispersion along the coast in St Helena Bay is likely due to the cyclonic circulation which isolates the region (refer to Sec. 3.3.3). The frontal edges (Fig. 3.18c, d) show some correspondence with the regions of high FTLE. In Fig. (3.18c), fronts border on the coastal side of the high FTLE within St Helena bay and along the coast further north while in Fig. (3.18d), the fronts frame the broad FTLE region between Cape Columbine and 31°S . The fact that high dispersion exists on the oceanic side of the frontal edges suggests that the fronts in winter are not playing a major dynamic role in restricting the offshore advection of particles. Therefore, alongshore fronts in summer will act to limit the offshore advection and losses of material, while in winter, the less intense fronts found further offshore will permit greater levels of cross-shore mixing and the potential losses of material.

3.3.4 Fronts and cross-shore circulation dynamics

From the previous section, the oceanographic properties of alongshore fronts in summer allow them to play a dynamic role in limiting the cross shore advection of particles. In winter, a homogenisation in cross-shore SST gradients and weak stratification of the water column promote the formation of less intense fronts which are found further offshore and do not play a major role for the shelf-bound retention of particles. To further explore the dynamics of retention along the SBUS shelf; the mean seasonal location of the UF and SBF are visualized relative to the major horizontal circulation

features (Fig. 3.19) and select cross-sections of vertical velocities along the shelf (Fig. 3.20). Additional cross-sections along the shelf showing meridional and zonal velocities are also referenced and available in the supplementary (Supp. 3.4S and Supp. 3.5S).

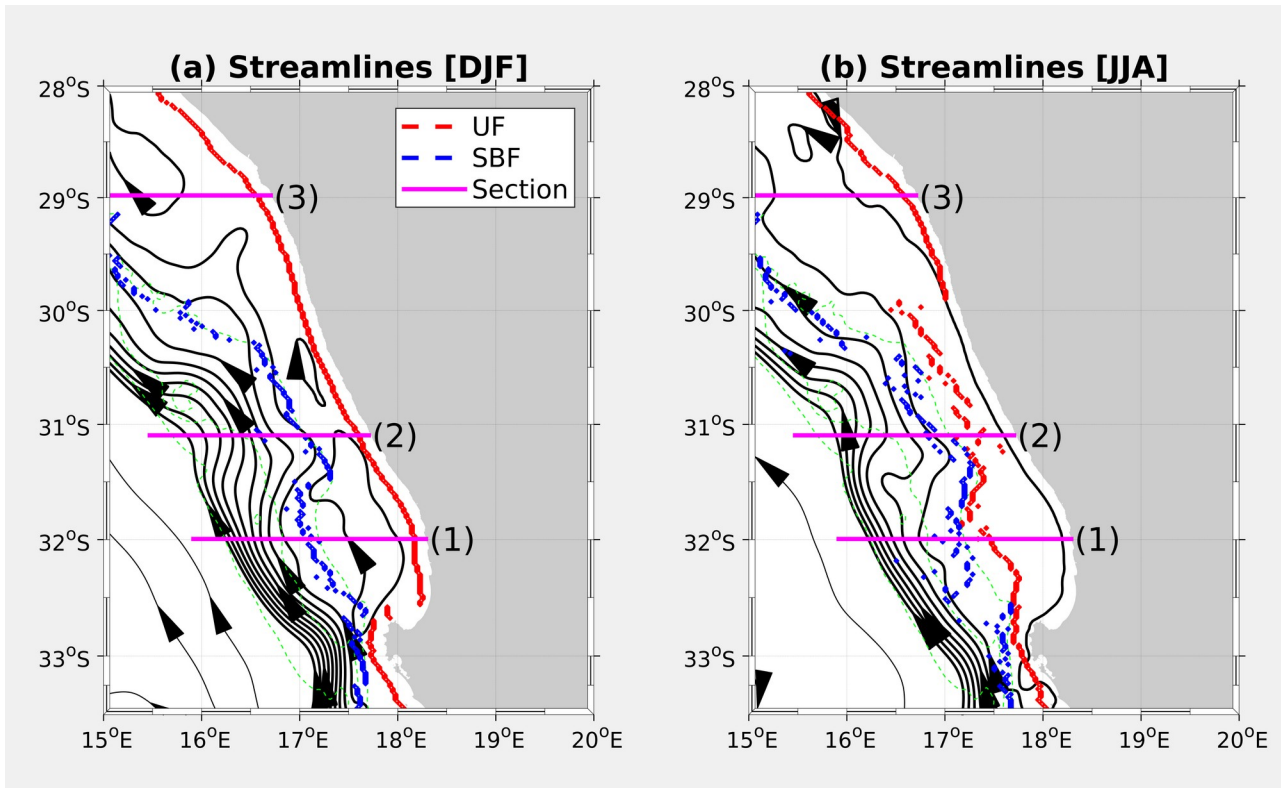


Figure 3.19: Streamlines of integrated currents from 0 – 300 m for summer (DJF) and winter (JJA) with selected isolines to show circulation structures and flow patterns along the shelf. The red and blue hashed lines indicate the position of the UF and SBF respectively. To calculate the respective frontal positions, the maximum probability within the upwelling and shelf-break region were assigned as the location of the UF and SBF. The three cross-section located within the (1) St Helena Bay, (2) Mid-shelf and (3) Namaqua regions are indicated.

In Fig. (3.19), the mean seasonal horizontal circulation patterns along the shelf are visualised with streamlines of the integrated currents from the surface to 300 m. In summer (Fig. 3.19a), the cyclonic circulation within St Helena Bay is evident as well as the prominent shelf-edge jet. The UF resides close to the coast while the SBF follows the shelf topography and meandering of the intense shelf-edge jet. The adopted methodology for the front identification could not resolve the detection of the CF but it is evident as the oceanic appendage of the cyclonic circulation within St Helena Bay (Lamont *et al.* 2015). Though not directly explored, the presence of the CF should also elevate retention within the region.

In winter (Fig. 3.19b), the cyclonic circulation within St Helena Bay has broadened such that the CF is not evident, a distinct poleward flow dominates along the coast and the intensity of the shelf-edge jet has weakened. An UF still forms along the Namaqua coast due to the persistence of southerly winds in the region. The near overlap in the position of the UF and SBF from north of Cape Columbine till 30°S suggests the absence of a distinct UF along the coast in winter, which is expected, owing to the absence of upwelling-favourable winds. In St Helena Bay, the SBF corresponds closely with the oceanic branch of the cyclonic circulation in the region and this overlap has previously been interrupted from observations as evidence of frontal convergence of the

CF with the SBF during winter (Lamont *et al.* 2015). The absence of the CF in St Helena Bay need not be tantamount to frontal convergence; instead, the shifting wind dynamics preclude upwelling off Cape Columbine and thus the formation of the CF.

To explore the relationship between the offshore positions of the UF and SBF relative to the cross-shore circulation dynamics along the shelf, three selected cross-sections of vertical velocities were investigated and are demarcated in Fig. (3.19). The first cross-section bisects St Helena Bay and crucially intersects with the three major alongshore fronts in the region during summer. The second is within the Mid-shelf region, where the shelf-width narrows resulting in a shoreward migration of the SBF. The third is associated with a broadening of the shelf within the Namaqua region and the associated offshore flow of the SBF as well as the persistence of an UF close to the coast during winter.

Fig. (3.20) reveals a clear connection between fronts and the vertical velocity fields during summer months along the SBUS. During winter, when these fronts are defined by weaker SST gradients and shift offshore, the velocities associated with them are much weaker. Furthermore, in both seasons, the MLD acts as a neat divider where vertical velocities are more intense below the MLD. The deepening of the MLD during winter limits the spatial extent of intense vertical velocities in the upper 50 m along the shelf. In summer, the UF corresponds with the oceanic edge of an upwelling cell along the coast for St Helena Bay and the Mid-shelf (Fig. 3.20a, b). In Namaqua (Fig. 3.20c), the UF seems to separate two upwelling cells close to the coast; with the coastal side of the front having negative velocities in the upper 40 m. The location of the UF is a bit anomalous in Namaqua (Fig. 3.20c) as one would expect the front to be on the oceanic edge of the positive velocities along the coast. Looking at the sections of meridional velocities along the shelf (Supp. 3.4S), the UF in Namaqua is found on the oceanic edge of an intense surface jet and corresponds well with the position of the 18 °C isotherm. Other frontal positions along the shelf are usually located on the coastal edges of the surface equatorward meridional jets and these regions have to be associated with maximums in SST gradient such that they are flagged by the Canny algorithm (Sec. 3.2.3). From Supp. (3.5S), onshore zonal velocities are present along the shelf bottom between the UF and coast, indicative of a recirculation pattern that may locally elevate retention in the region.

In St Helena Bay (Fig. 3.20a), the SBF is associated with the surface outcropping of the 18 °C isotherm as well as positive vertical velocities associated with shelf-edge upwelling during summer. From the patterns of vertical velocities, no obvious cross-shore recirculation system exists within St Helena Bay. The zonal velocities in Supp. (3.5S) support this as no onshore velocities are present along the shelf bottom close to the coast; however, the cyclonic circulation within the bay (Fig. 3.7) does add to the retention of particles in the region. The Mid-shelf region (Fig. 3.20b) has a defined cross-shore recirculation pattern. From Fig. (3.20b), the patterns of vertical velocity show upwelling along the coast and downwelling along the coastal edge of the SBF. From Supp. (3.4S), offshore zonal velocities are present in the upper 40 m and between the UF and the shore onshore velocities along the shelf bottom are present. This suggests that the cross-shore recirculation pattern in the Mid-shelf extends from the coast up to the SBF. Similar to St Helena Bay, intense vertical velocities along the oceanic edge of the SBF are linked to shelf-edge upwelling. Multiple upwelling cells are present along the broad Namaqua shelf and the SBF divides two such cells. The prominent offshore flow in the region (Supp. 3.5S) may not favour the accumulation of particles close to the coast; however, similar to the Mid-shelf region, onshore zonal velocities are present along the shelf bottom between the UF and coast. Conversely, no prominent downwelling is evident along the coastal edge of the SBF suggesting that a pronounced cross-shore recirculation pattern may not be

present along the Namaqua shelf. Regardless, the broad shelf and multiple upwelling cells are likely to enhance the retention of matter shoreward of the 200 m isobath; however, retention is not likely to be high in the nearshore regions as a consequence of the alongshore fronts being located further from the coast and the intense zonal velocities in the region.

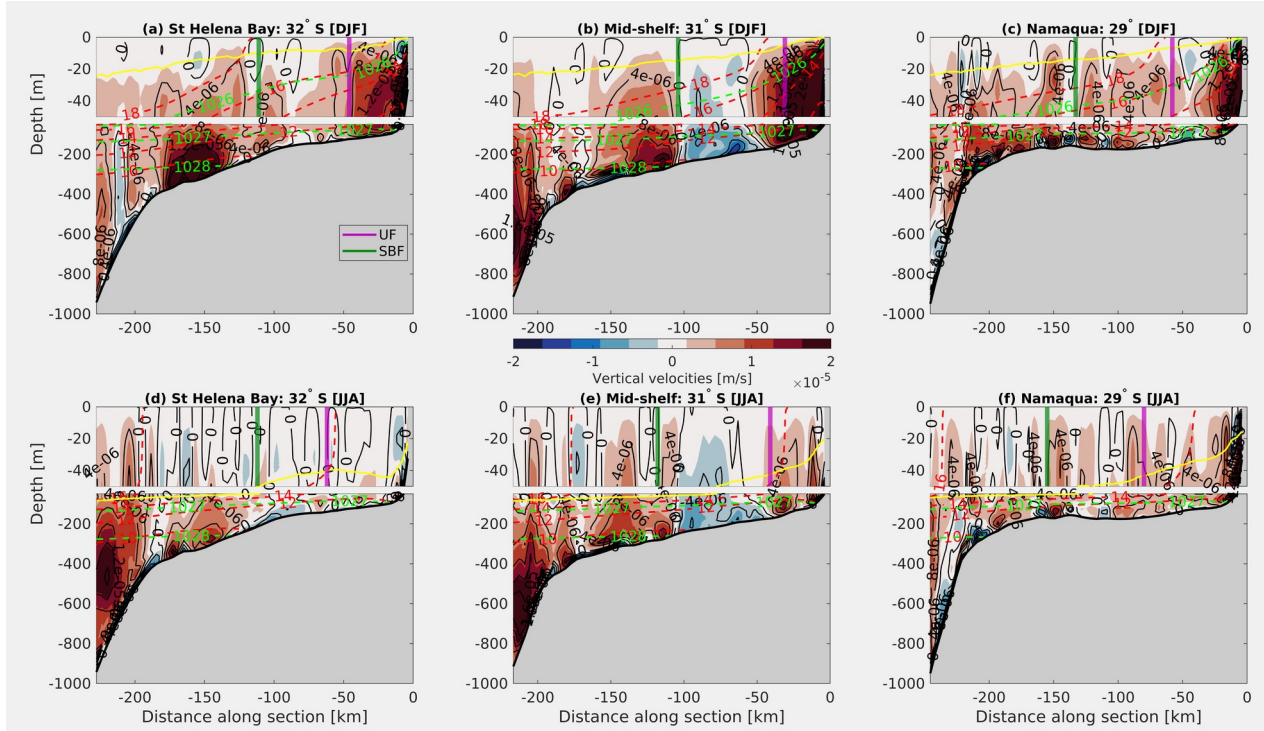


Figure 3.20: Averaged summer and winter sections of vertical velocities across the SBUS showing (a, d) St Helena Bay (b, e) Mid-shelf and (c, f) Namaqua. The yellow solid line demarcates the MLD while the hashed red and green lines shows the isotherms and isopycnals respectively. The vertical cyan line indicates the mean position of the UF while the dark green line is the position of the SBF.

In winter the deepening of the MLD and shifting wind patterns starkly impact the frontal environment as well as vertical, zonal and meridional velocities along the SBUS. In St Helena Bay (Fig. 3.20d), the absence of upwelling-favourable winds precludes upwelling along the coast. Although the computed UF corresponds with the position of the 14 °C isotherm, this front is not generated due to upwelling and is associated with a weak equatorward surface jet (Supp. 3.4S). The frontal feature does divide a poleward flow along the coast from an equatorward flow on the oceanic side of the front. The position of the SBF shows some relation to shelf-edge upwelling (Fig. 3.20d). Furthermore, the sparse positive vertical velocities between the UF and SBF in the upper 50 m is likely due to the absence of the CF (seen also in Fig. 3.19b). The weakening of the shelf jets and decreased stratification limit upwelling along the shelf-edge and this is a feature experienced across the SBUS.

In the Mid-shelf region (Fig. 3.20e), the cross-shore recirculation pattern is reduced in intensity relative to summer. Unlike in St Helena Bay, the UF and SBF are still defined by prominent vertical velocities and some onshore zonal flow is present between the UF and coast (Supp. 3.5S). The identified UF also divides a coastal poleward flow from an equatorward flow, similar to St Helena Bay. The persistence of a cross-shore recirculation pattern will likely elevate the coastal retention of particles, even in winter. The UF and SBF show the most significant offshore migration within Namaqua (Fig. 3.20f). The UF does not seem to be associated with a prominent isotherm, such as in

St Helena Bay or in the Mid-shelf. The front seems to be far removed from an obvious coastal upwelling cell that extends approximately 20 km offshore. The anomalous offshore location may be due to multiple fronts being detected within the upwelling region of the Namaqua shelf which skews the calculation of offshore position. Regardless, a genuine UF is likely to form along the coast, even during winter and the cell is associated with a prominent baroclinic jet (Supp. 3.4S). Furthermore, the zonal velocities still show some onshore velocities along the shelf bottom between the UF and the coast (Supp. 3.5S) suggesting that a cross-shore recirculation pattern is still sustained, albeit less intense. The weakening of the vertical velocities, offshore migration of the fronts and a weakening of the nearshore recirculation pattern is likely to limit the cross-shore cycling of particles but the absence of intense offshore flow could in fact elevate retention of particles in the nearshore during winter.

3.4 Discussion

3.4.1 Modelling the SBUS

An array of *in situ* observational and L4 satellite data products were used to comprehensively evaluate the model configuration for the SBUS. Several shortcomings were identified in the model's performance; notably were persistent temperature biases, both at the surface and with depth along the shelf as well as a significantly deeper winter MLD relative to observations. Despite these shortcomings, the major features such as the cyclonic circulation within St Helena Bay, the permanent CP jet along the Cape Peninsula as well as the pattern of surface equatorward jets and poleward undercurrents along the shelf were resolved for both summer and winter. Furthermore, the low frequency intra-seasonal and interannual variability for SST, intensity of mesoscale activity and some of the nearshore temperature variability along the shelf were well resolved.

The broad pattern of warm SST bias observed over the shelf is a ubiquitous feature to other EBUS (Ma, Xu & Wang 2019) and stems from a combination of model resolution as well as the resolution of the atmospheric forcing which influences the intensity of curl driven upwelling as well as upwelling along the coast (Capet, Marchesiello & McWilliams 2004). In summer (Fig. 3.5c), a cool nearshore bias was evident off Cape Columbine and St Helena Bay. The juxtaposition of this cool bias alongside the warm shelf bias lead to an enhancement of the cross-shore SST gradients resulting in a more intense and energetic CP jet in closer proximity to the coast off Cape Columbine and elevated frontal intensities within St Helena Bay. Consequently, the SST bias pattern along the shelf impacted the intensity and positions of the fronts and alongshore jets. Strub & James (2022) recently showed that the ERA5 reanalysis product had a tendency to have higher wind-stress within 10km of the coast relative to observations. This increase in wind-stress stemmed from the interpolation of the wind-stress fields to a regular rectilinear grid. Higher wind-stress along the coast would expectedly generate more upwelling and cooler SST patterns along the entire shelf in summer; however, the cool biases were localised to known region of intense upwelling and cross-shore SST gradients. Carr, Lamont & Krug (2021) showed L4 satellite products struggled to resolve regions of strong SST gradients within ~ 50 km off the shore due to the lack of input observations and reliance on interpolation schemes. Consequently, the near coastal SST patterns are likely to be better resolved within the model physics than the L4 data product and this warrants consideration when conducting a model evaluation for the SBUS.

The cool offshore SST bias in Fig. (3.5 and 3.6) could have been due to reduced Agulhas influx. Modelling work by Veitch & Penven (2017) showed how warm water introduced into the Benguela

system through Agulhas leakage has a warming effect that was greatest in the offshore regions. Further evidence for this was seen in the lower EKE in the open ocean regions in the model. This issue most probably arises due to the difficulties in resolving the boundary conditions as well as being a product of the physical forcing conditions themselves. Relative to the review conducted by Russo *et al.* (2022), the model improved upon many aspects of the GLORYS forcing data; notably was the reduction in the surface temperature biases. However, GLORYS inherently had a deep winter MLD (Russo *et al.* 2022) and this artefact was exasperated within the model. Consequently, a deeper MLD would impact the vertical mixing processes and stratification in the surface layer.

In exploring and appreciating the shortcoming of the simulation it does not heavily detract from the overall performance of the model in the SBUS. To improve the model configuration, regional specific wind atlases (Lennard *et al.* 2015), which are still in development, could be applied in future to reduce the warm shelf SST bias and improve the upwelling dynamics along the coast. Withstanding the temperature biases and deep winter MLD, a long-term interannual physical simulation of the SBUS has been presented and evaluated which may be used for further investigation of the physical dynamics of the region or for comparative modelling works.

3.4.2 *Physical oceanographic properties of hydrographic fronts*

The differing summer and winter frontal environments impact the respective seasonal cross-shore mixing patterns of material along the SBUS shelf (Fig. 3.14). In summer, intense oceanic fronts along the shelf are sustained by prominent cross-shore SST gradients. Divergence along the coast maintains intense and contiguous UFs parallel to the shore, while along the shelf-edge, warm waters associated with Agulhas leakage sustain the SBF as it follows the meandering shelf topography. These summer fronts show a good correspondence in position to prominent LCSs (Fig. 3.17). This suggests that fronts in summer are able to act as cohesive barriers to the offshore advection of passive tracers as well as dynamically divide the flow patterns along the shelf (Haller 2001; Shadden *et al.* 2005). The association of SST fronts and LCS has been previously observed by Mathur *et al.* (2019) in the Bay of Bengal as well as by McKee *et al.* (2023) who also noted enhanced vertical velocities along LCS. Consequently, the intense and cohesive fronts in summer will play a dynamically important role in limiting the cross-shore advection of particles and elevating the retention of matter along the shelf.

In winter, the coupled and non-linear effect of weakened stratification, a decrease in cross-shore SST gradients as well as a shift to predominantly northerly winds result in more numerous but less intense fronts forming along the shelf. These fronts are found further offshore and are statistically smaller in length than those in summer. The SBF still remains quasi-stationary; however, the UF shows a significant offshore migration between Cape Columbine and 30°S. The near overlap in the positions of the SBF and UF is tantamount to the absence of a clear UF in winter, which is expected. Further north, along the Namaqua shelf, the near perennial southerly winds in the region still sustain an UF in close proximity to the shore. In the SBUS, the filamentous and weak fronts in winter are not associated with any obvious LCS and as such do not play a major role in limiting the offshore advection of material.

Along with variability in the seasonal frontal patterns, the SBUS exhibits regional and seasonal differences in circulation patterns that will likely impact the shelf-bound retention of material along the shelf. From the various cross-shelf sections and frontal maps, the seasonal frontal locations corresponded well to the positions of the major alongshore surface equatorward jets and edges of

the various upwelling cells. Key circulation features along the shelf that will likely contribute to enhanced retention include the cyclonic circulation within St Helena Bay, cross-shore recirculation pattern within the Mid-shelf and multiple upwelling cells along Namaqua. The cyclonic circulation within St Helena Bay acts to isolate the region, which limits frontal activity and results in lower levels of particle dispersion. A seasonal comparison of the frontal and dispersion patterns reveal the cyclonic circulation as being a perennial feature, similar to modelling works by Penven *et al.* (2000). Another facet to the retention dynamics within the bay is the frontal arrangements. Oerder *et al.* (2018) showed that the position of fronts is controlled by the configuration of the alongshore jets. In summer, the presence of the CF as well as an enhanced cyclonic circulation within the bay likely elevate retention in the region. In winter, the absence of the UF and frontal convergence of the CF with SBF (Lamont *et al.* 2015) limit frontal activity close to the coast, negatively impacting retention.

For the SBUS, the seasonal frontal dynamics and recirculation features along the shelf should favour elevated coastal retention of particles in summer compared to winter. In summer, the more intense and continuous alongshore fronts in close proximity to the coast should complement the defined horizontal and cross-shore circulation patterns in fostering high retention. Conversely, the weak and filamentous fronts in winter, which are found further offshore, will not aid the shelf bound retention of particles. Furthermore, the weakening and absence of recirculation features will further impede coastal retention. Consequently, the high retention of particles during summer will foster elevated levels of biological activity and through ‘nutrient trapping’, hypoxic events will be more prevalent (Flynn *et al.* 2020). To better explore retention along the shelf, Lagrangian experiments are undertaken (Chapter 4) to understand how the seasonal frontal environments and shelf circulation patterns interact.

3.4.3 Wind and stratification on frontal variability

The physical oceanographic environments between summer and winter differ starkly to each other and so to do the behaviour of the fronts. From Fig. (3.14a, b), the patterns of frontal probabilities are similar to those of Wang *et al.* (2015) where frontal probabilities are greatest closest to the coast in summer whereas the region of highest frontal probability in winter is found further offshore. In summer, fronts are usually long and continuous alongshore features that are defined by strong SST gradients and found close to the coast. In winter, fronts are found further offshore and are more numerous and filamentous in nature, being defined by weaker SST gradients and shorter frontal lengths. The dichotomous seasonal behaviour of fronts is due to the non-linear interaction and oscillating seasonal dominance between alongshore wind and stratification.

The results suggest that the main driver of frontal activity and intensity in summer is alongshore wind and stratification in winter. The dominance of wind in driving upwelling intensity and variability in summer within the Benguela has been well established (Wang *et al.* 2015; Bonino *et al.* 2019; Bordbar, Mohrholz & Schmidt 2021) and thus our findings sit well within the current literature. Tipped to be competitive or complementary to the wind in affecting upwelling dynamics, the role of stratification is an important driver to consider for frontal variability. He & Mahadevan (2021) showed a scaling relationship of source depth waters being related to $(\tau/N)^{0.5}$ where τ is the surface wind-stress and N is stratification. The relationship suggests that deeper water will be brought to the surface during strong wind and low stratification events. Acknowledging the power of He & Mahadevan (2021)’s scaling relationship it is important to appreciate that the theoretical framework used a basic mixing scheme and no topography. Consequently, the complex

K-profile vertical mixing (Large *et al.* 1994) scheme employed in coastal models as well as diapycnal processes (Fearon *et al.* 2020) may add additional variability for fronts in the nearshore region.

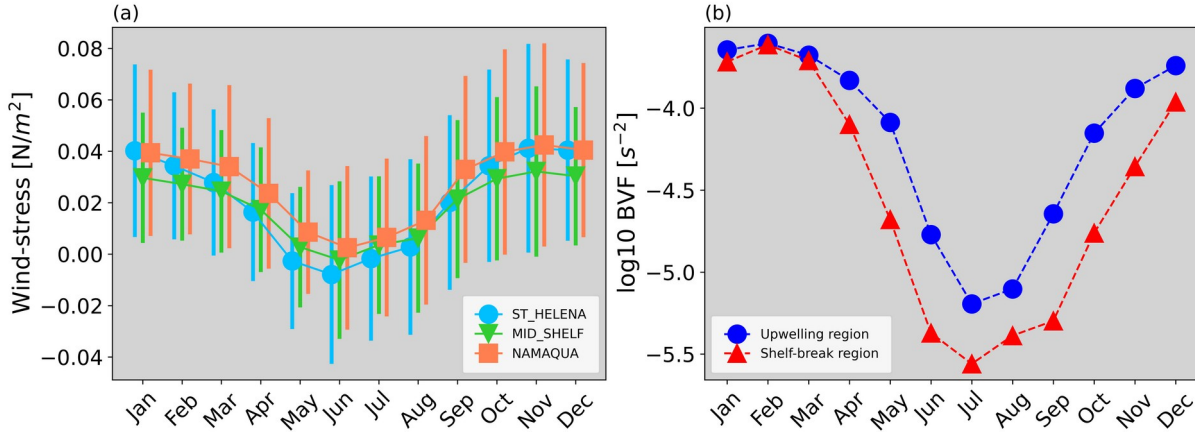


Figure 3.21: (a) Climatology of wind-component parallel to the coast with bars showing ± 1 monthly std. Coastal wind-stress field is averaged across 4 model grid-cells from the coast (~ 10 km) (b) Climatology of stratification (BVF at 30 m) along the SBUS shelf for the upwelling and shelf-break region.

In the SBUS, the seasonal cycle of the alongshore wind-stress and stratification are in phase with each other (Fig. 21). Strong southerly winds and high stratification are present in summer and weak northerly winds and low stratification prevail in winter. In summer, the alongshore wind is more dominant in driving frontal variability and counters the associated increase in stratification. In winter, the predominantly northerly winds are not upwelling-favourable and the low stratified well mixed waters limit the extent of the intense vertical velocities to being below the deep MLD. Consequently, the elevated offshore frontal probabilities in winter may arise as a consequence of the low stratification of the water column. Owing to the too deep winter MLD resolved in the model, the importance of stratification may be exasperated and as such a shallower MLD may amplify the importance of wind forcing in driving frontal variability in winter. From He & Mahadevan (2021)'s scaling relationship, low stratification would make the upper ocean more sensitive to weak mesoscale features that cause an uplift of the thermocline; generating frontal activity at the surface. If this were the case, it would explain why fronts are more numerous and smaller in length than in summer as they are generated by small instabilities.

For the SBUS, there was a good correspondence between elevated FTLE and frontal probability suggesting that frontal activity and mixing are positively related. Using Finite Size Lyapunov exponents (FSLE), which produce similar results to FTLE for long integration times (d'Ovidio *et al.* 2004; Hernández-Carrasco *et al.* 2011), Rossi *et al.* (2009) found similar seasonal trends in FSLE to our FTLE within the SBUS; although, their regions of interest were along the shelf-edge and open ocean. Nevertheless, Rossi *et al.* (2009) showed that the SBUS had a clear seasonal cycle for mixing, with mixing intensity being greatest in winter. Callies *et al.* (2015) showed that for deep winter MLD's, the lateral buoyancy gradients could store potential energy such that any slumping in the isopycnals could generate baroclinic instabilities and lead to enhanced frontal activity. Similar to He & Mahadevan (2021), Callies *et al.* (2015) showed that the well stratified waters in summer would rapidly dissipate small scale instabilities. Therefore, the low stratified winter column facilitates the formation of more fronts, although these generated fronts are significantly weaker than those in summer. Furthermore, a decrease in stratification and erosion of prominent alongshore

fronts could favour elevated cross-shore exchange (Rixen *et al.* 2021) and facilitate the flushing and re-oxygenation of the bottom waters of the shelf.

3.5 Conclusion

Using a numerical model of the SBUS in conjunction with a gradient-based edge detection algorithm, the seasonal frontal dynamics along the shelf region were investigated. In evaluating the model performance, it resolved the low-frequency interannual and intra-seasonal variability for SST and the major surface circulation features for the region. An artefact of the physical forcing conditions, the model resolved a very deep winter MLD and had positive temperatures biases along the shelf as well as cool nearshore biases off Cape Columbine and within St Helena Bay during summer. The bias pattern along the shelf impacted the seasonal positions and intensities of the various fronts and alongshore jets. The seasonal shift in the alongshore wind regimes, stratification and insolation affected the position, intensity and structure of fronts along the SBUS. In summer, fronts are usually long and continuous alongshore features which are defined by strong SST gradients and associated with prominent LCS. These fronts act as cohesive barriers and limit the cross-shore mixing of particles. In winter, fronts are generally found further offshore and are more filamentous and numerous compared to summer. Furthermore, they are defined by weaker SST frontal gradients and shorter frontal lengths. Fronts in winter are not associated with any obvious LCS and the high dispersion patterns across the shelf indicated that they do not play a dynamic role in restricting the offshore advection of particles. Along with a seasonal shift in the physical dynamics of the alongshore fronts, the arrangement of fronts along the shelf also changed.

In summer, the UF, formed by divergence along the coast, was found close to the shore. Along the shelf-edge, the SBF followed the meandering of the shelf-topography and was sustained by warm Agulhas leakage. Upwelling off Cape Columbine generates an upwelling plume and the associated CF. In winter, the shift to more northerly winds precludes upwelling and the formation of the CF off Cape Columbine; however, the SBF still remained near quasi-stationary. Furthermore, the UF showed an offshore migration and near overlap in position with the SBF between Cape Columbine and 30°S, indicating the absence of the UF during winter. Along the broad Namaqua shelf, the perennial southerly winds still sustained an UF in close proximity to the coast. In interrogating the seasonal circulation patterns along the shelf, the position of the fronts aligned well with the surface equatorward jets and edges of the various upwelling cells along the coast. Prominent circulation features that likely enhance retention along the coast included the perennial cyclonic circulation within St Helena Bay, cross-shore recirculation cells in the Mid-shelf region and multiple upwelling cells in Namaqua. Consequently, the alongshore fronts in summer complimented the major retentive cross-shore circulation structures along the shelf. In winter, the less intense circulation features and weaker alongshore fronts are unfavourable to the high retention of material.

Supplementary

3.1S. Statistical equations

$$r = \frac{\sum_{i=1}^N (O_i - \bar{O})(P_i - \bar{P})}{\sqrt{\sum_{i=1}^N (O_i - \bar{O})^2 \sum_{i=1}^N (P_i - \bar{P})^2}} \quad (3.i)$$

$$RMSD = \sqrt{\frac{\sum_{i=1}^N [(P_i - \bar{P}) - (O_i - \bar{O})]^2}{N}} \quad (3.ii)$$

$$Bias = \frac{\sum_{i=1}^N (P_i - O_i)}{N} = \bar{P} - \bar{O} \quad (3.iii)$$

Eq. (3.i) is the correlation coefficient, Eq. (3.ii) is the centred Root-mean square and Eq. (3.iii) is the bias. For the statistical equations, N is the number of observations, O_i is the i th of N observations and P_i is the i th of N predictions, where $\bar{O}$ and $\bar{P}$ are the respective observed and predicted means.

3.2S. Detection thresholds

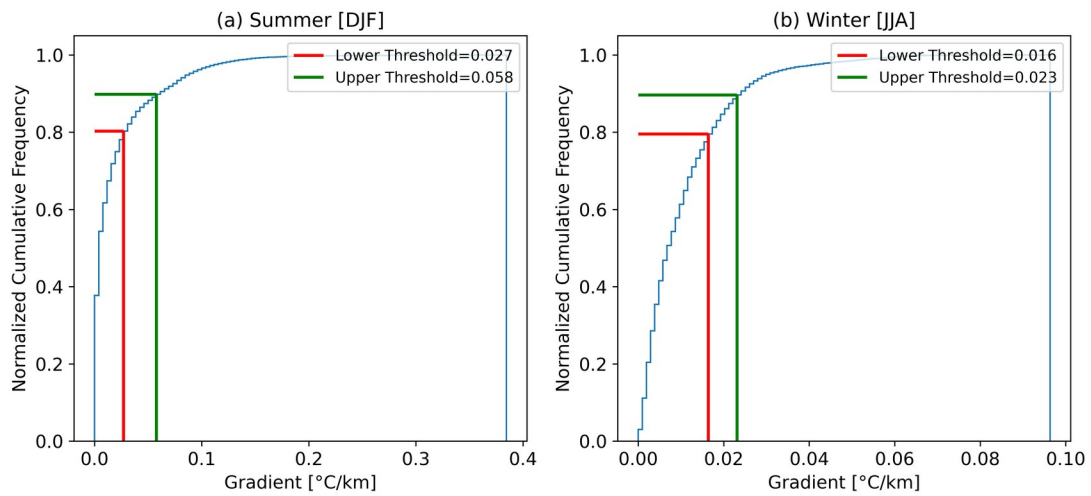


Figure 3.i: Seasonal thresholds for frontal detection in (a) summer and (b) winter

The detection thresholds for the Canny algorithm used the 80th and 90th percentiles for the lower and upper bounds. For each SST map, a cumulative frequency of the SST distribution was constructed such that each SST map was individually and independently evaluated. Consequently, the absolute detection criteria would change per SST map. Fig. (3.i) shows the seasonally averaged detection thresholds from the model along the SBUS shelf region. The detection criteria for fronts in summer were significantly greater than in winter. This arises because summer is defined by more intense SST gradients than winter on account of greater solar insolation as well as upwelling along the coast which set up intense cross-shore SST gradients along the coast. Changing the thresholds would inevitably affect the nature of the fronts detected. For example, if absolute thresholds defined from the annual mean SST were used then fewer fronts would be detected in winter and more fronts would be detected in summer. Therefore, the choice and use of thresholds warrant important consideration when interpreting the results.

3.3S SAMBA mooring bottom temperature time-series

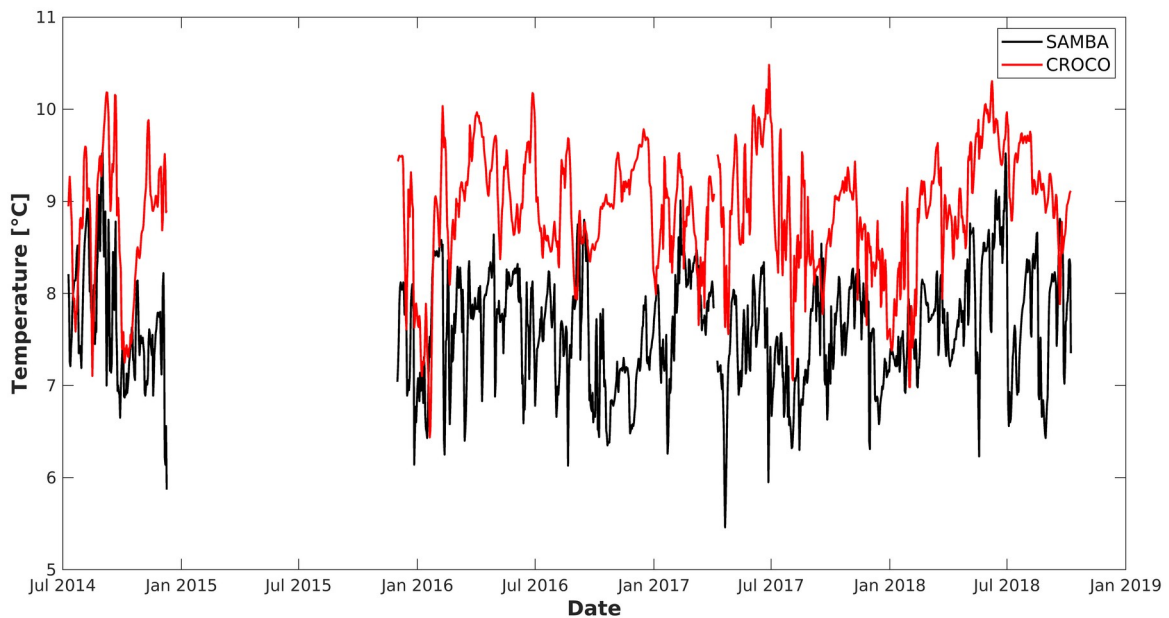


Figure 3.ii: Bottom temperature time-series of model (red) vs observations (black)

The model had a positive SST bias at depth along the SBUS shelf which impacted the depth of the MLD and the intensity of the poleward return flow. Relative to the SAMBA mooring at 300 m, the model was consistently warmer; showing considerable interannual variability. The biggest discrepancies between the model and observations generally occurred in summer, with the model resolving the variability best during winter; albeit with a positive bias.

3.4S Cross-sections of meridional velocities

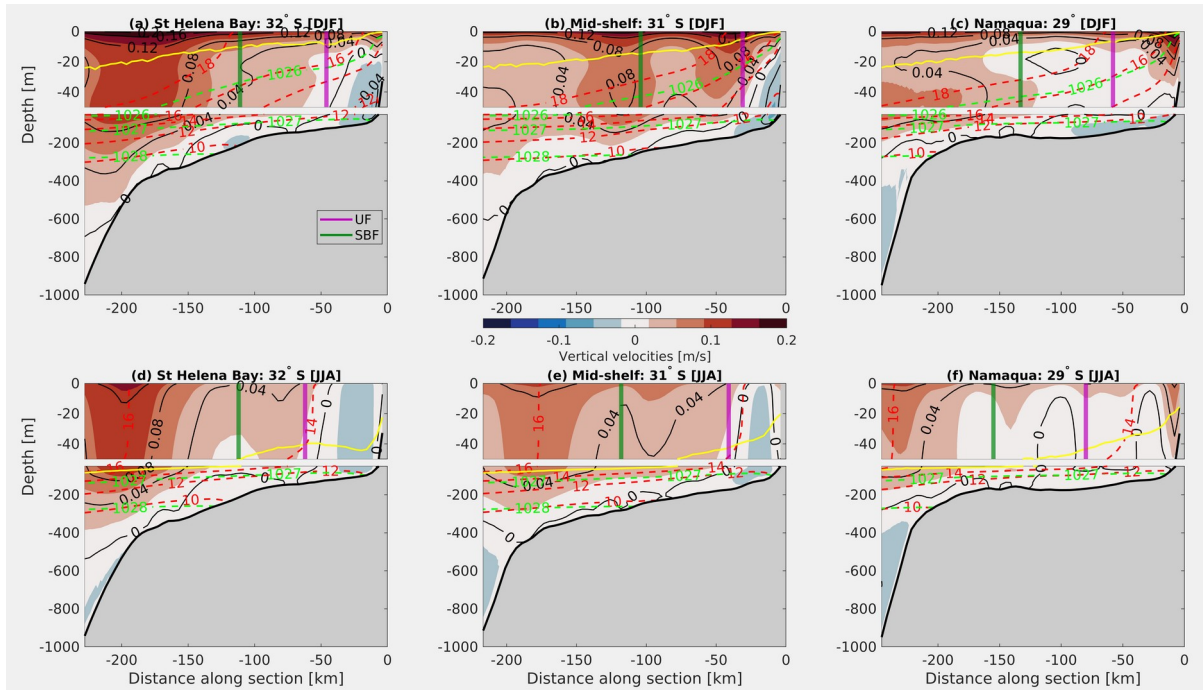


Figure 3.iii: Averaged summer and winter sections of meridional velocities across the SBUS showing (a, d) St Helena Bay (b, e) Mid-shelf and (c, f) Namaqua. The yellow solid line demarcates the MLD while the hashed red and green lines shows the isotherms and isopycnals respectively. The vertical cyan line indicates the mean position of the UF while the dark green line is the position of the SBF.

3.5S Cross-sections of zonal velocities

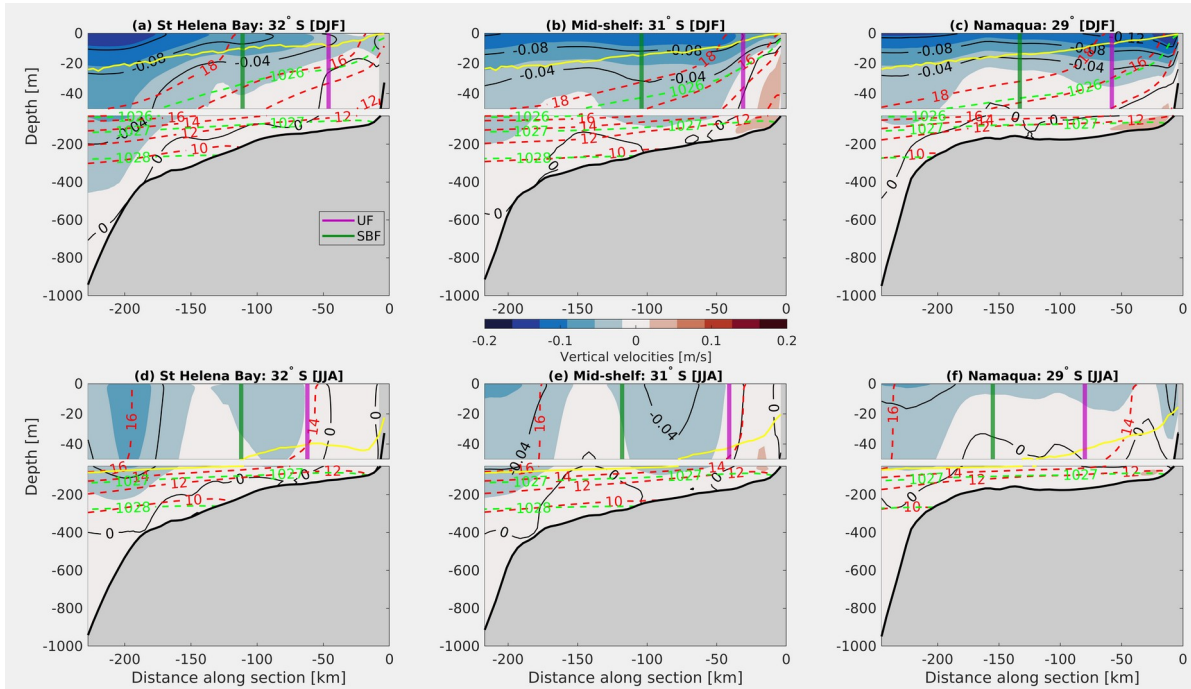


Figure 3.iv: Averaged summer and winter sections of zonal velocities across the SBUS showing (a, d) St Helena Bay (b, e) Mid-shelf and (c, f) Namaqua. The yellow solid line demarcates the MLD while the hashed red and green lines shows the isotherms and isopycnals respectively. The vertical cyan line indicates the mean position of the UF while the dark green line is the position of the SBF.

Chapter 4: Investigation of retention dynamics along the Southern Benguela Upwelling System: a modelling approach

*“All those moments will be lost in time, like tears in rain”
– (Blade Runner, 1982)*

Summary

Cross-shore hydrography plays a major role in elevating the shelf-bound retention of matter within Eastern Boundary Upwelling Systems (EBUS). The elevated retention of nutrients is crucial for enhancing biological productivity along coastal margins but a caveat is the increase in occurrence of hypoxic events. A physical ocean model in conjunction with a Lagrangian tracking code and an edge-detection algorithm is used to investigate some of the physical oceanographic drivers of retention along the Southern Benguela Upwelling System (SBUS). Relative to observational drifters, the model resolves a more westward propagation of floats along the shelf edge in summer and reduced northward transport of floats during winter. The intensity of the alongshore surface jets and poleward undercurrents are weaker within the model compared to observations and as such the computed residence times are likely to be overestimated. Floats in the upper 50 m are transported northward by the prominent equatorward surface jets while floats below 100 m are primarily transported south along the narrow shelf region between Cape Town and Cape Columbine by the poleward undercurrents. The average seasonal residence times for particles initiated in the upper 50 m are 69 ± 14 days for summer and 123 ± 35 days in winter. For particles between 100-250 m, the mean residence times are 57 ± 10 and 132 ± 38 days for summer and winter respectively. Interannual and seasonal variability in the alongshore wind dynamics impact the intensity of the offshore Ekman transport, poleward undercurrents as well as the offshore positions of the alongshore fronts which inevitably affects the residence times of particles along the SBUS shelf. Furthermore, residence times also show a somewhat cyclic pattern, where maxima in residence times occur at intervals of 4-5 years which is weakly correlated in the surface layers to the periodicity of the El Niño–Southern Oscillation.

4.1 Introduction

The four major Eastern Boundary Upwelling Systems (EBUS) of the global ocean constituting the California, Humboldt, Canary and Benguela upwelling systems collectively occupy ~1% of the ocean surface area but account for approximately 20% of all commercial fishery catches (García-Reyes *et al.* 2015). Strong alongshore equatorward winds force offshore Ekman transport which results in cold, nutrient-rich deep waters being upwelled into the surface layers and stimulating high levels of biological productivity. Though somewhat paradoxical, the juxtaposition of enhanced offshore transport and elevated biological productivity along coastal margins stems from the role

the cross-shore hydrography plays in elevating the shelf bound retention of matter (Roughan *et al.* 2006; Lamont *et al.* 2015). In this work, focus is given to exploring these dynamics within the Southern Benguela Upwelling System (SBUS).

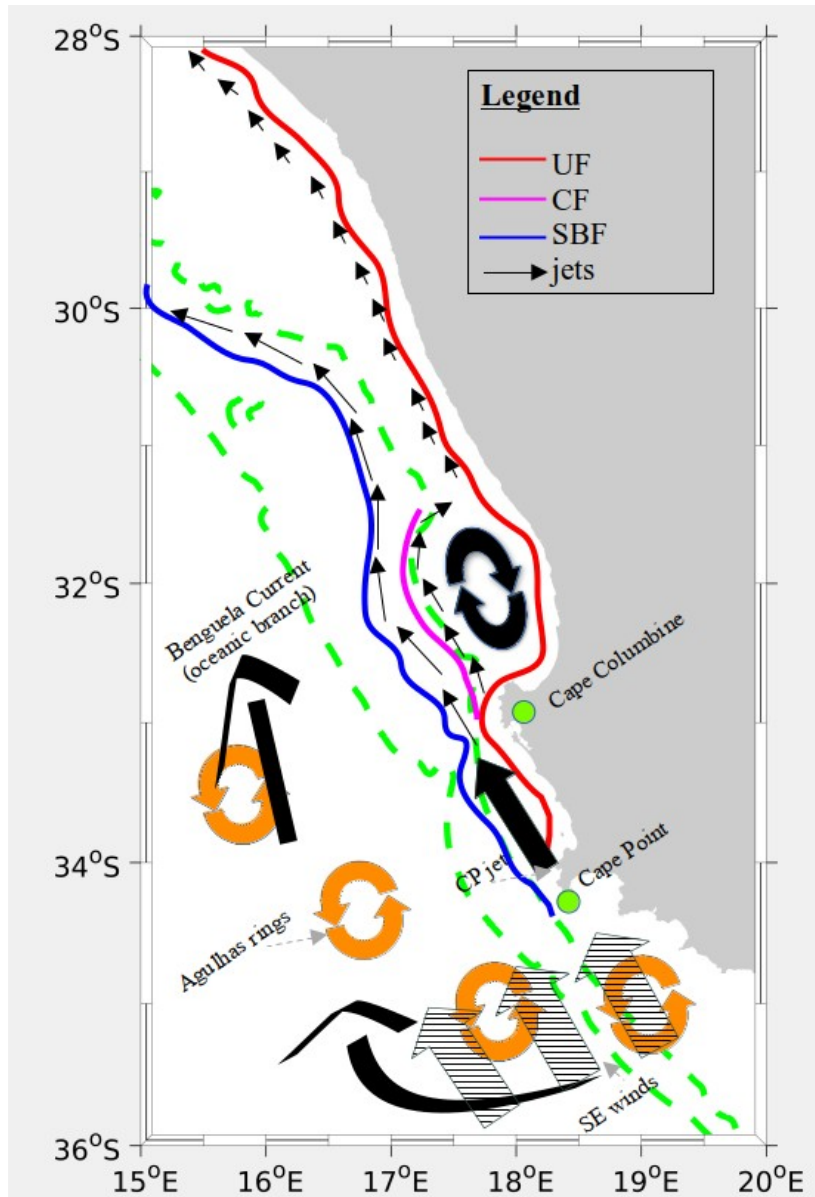


Figure 4.1: Schematic showing the major summer physical oceanographic features of the SBUS. Southerly winds stimulate upwelling along the shelf resulting in the formation of the Upwelling front (UF) and Columbine front (CF). The shift to northerly winds in winter precludes the formation of these alongshore fronts. The Shelf-break front (SBF) is a permanent hydrographic feature of the region. Green dotted lines show the position of the 200 and 500 m isobath.

The Benguela Upwelling System consists of a broad northward flowing current along the west coast of southern Africa. Typically, the system is delineated into two sub-regions divided by the Lüderitz upwelling cell at 27°S (Hutchings *et al.* 2009; Veitch, Penven & Shillington 2010). The northern region experiences perennial upwelling while the south (SBUS) is prone to seasonal upwelling during late spring and summer (Andrews & Hutchings 1980; Hutchings *et al.* 2009; Veitch, Penven

& Shillington 2009). The seasonal wind regimes for the SBUS (Fig. 4.1) are controlled by the South Atlantic Anticyclone which gives rise to upwelling-favourable south-easterly winds in late spring and summer and north-westerly winds for the shelf region south of 30°S in winter (Strub *et al.* 1998; Shillington *et al.* 2006). Superimposed on these seasonal wind patterns are mesoscale and synoptic scale variability (Shillington *et al.* 2006). Walker (1987) showed internannual changes in the position of the Inter Tropical Convergence Zone to impact the position of the South Atlantic Anticyclone. A shift in the meridional position of the anticyclone would affect upwelling intensities along the South East Atlantic. Consequently, the role of remote forcing and large-scale modes of variability such as El Niño–Southern Oscillation (ENSO) and the Antarctic Oscillation can influence the atmospheric circulations along the South East Atlantic and thus over the Benguela region (Walker 1987; Shillington *et al.* 2006; Rouault, Pohl & Penven 2010; Tim, Zorita & Hünicke 2015; Rouault & Tomety 2022).

Along with the seasonal wind variability, the oceanography of the SBUS (Fig. 4.1) is influenced by several alongshore fronts and jets (Armstrong *et al.* 1987; Boyd & Nelson 1998; Penven *et al.* 2001; Lamont *et al.* 2015; Veitch *et al.* 2018), two prominent capes (Penven *et al.* 2000; Veitch *et al.* 2009; Veitch *et al.* 2010; Strub *et al.* 2013), variable shelf bathymetry (Boyd 1987; Pfaff *et al.* 2022), heat inputs from the Agulhas Current (Strub *et al.* 1998; Blanke *et al.* 2009; Veitch & Penven 2017) and a poleward undercurrent along the shelf bottom (Nelson 1989). A strong equatorward jet exists between Cape Town and Cape Columbine (Boyd & Nelson 1998), herein referred to as the Cape Peninsula (CP) jet. Modelling work by Veitch *et al.* (2018) has shown the CP jet to be a permanent feature, flowing between the 200-500 m isobath and being more intense in summer than winter. The jet plays an important role in the northward transport of larvae and eggs from the Agulhas Bank to nursery grounds in St Helena Bay (Penven *et al.* 2001). North of Cape Columbine, the CP jets bifurcates (Veitch *et al.* 2018) and along with several other important hydrographic fronts, contributes to the shelf-bound retention of matter.

Armstrong *et al.* (1987) distinguished between two main fronts: a seasonal Upwelling front (UF) and a permanent quasi-stationary Shelf-break front (SBF) (Fig. 4.1). The UF is prominent in summer, and develops due to the upwelling of cold nutrient-rich South Atlantic Central Waters (SACW) along the shelf. Far from being a continuous alongshore feature, the episodic UF can develop filamentous appendages that can spread offshore (Brink 1987). Furthermore, the UF can be displaced cross-shore during sustained upwelling events (Roughan *et al.* 2006). A continuation of the CP jet, the SBF is a permanent oceanographic feature of the region and is sustained by the cross-shore temperature gradients between the cool shelf waters and warmer offshore waters that are supplied by Agulhas leakage (Blanke *et al.* 2009; Veitch & Penven 2017). From long-term cross-shelf hydrographic sampling stations, Lamont *et al.* (2015) showed another seasonal front associated with the Columbine Upwelling plume that develops off Cape Columbine during upwelling events (Penven *et al.* 2000). Termed the Columbine front (CF), the feature is partially fed by the inner branch of the CP jet as it bifurcates north of Cape Columbine (Penven *et al.* 2000; Veitch *et al.* 2018) and is enhanced by the upwelling plume formed off of the headland (Nelson 1981; Penven *et al.* 2000). The transition from summer to winter sees a shift in the wind regimes and a weakening in the stratification of the water column associated with winter mixing. This inevitably precludes the formation of the UF and CF, weakens the intensity of the SBF and sees the alongshore jets moving further offshore (Wang, Castelao & Yuan 2015; refer to Sec. 3.3.3).

Variations in the orientation of the coastline as well as the presence of capes and bays influences the patterns of the alongshore winds (Pitcher & Pillar 2010; Strub *et al.* 2013) such that upwelling

intensity along the SBUS coast is not contiguous but instead limited to a series of upwelling cells at the Cape Peninsula (34°S), Cape Columbine (33°S) and Namaqua (30°S) (Veitch *et al.* 2010). In the lee of the Cape Columbine headland, equatorward winds generate and modulate the intensity of a cyclonic circulation within St Helena Bay (Penven *et al.* 2000). The feature adds to the retentiveness of the region such that the bay is highly productive but prone to seasonal hypoxia and harmful dinoflagellate blooms (Pitcher *et al.* 2010; Jury 2014). Variability in the shelf width also exerts significant control on the offshore position of the alongshore fronts and jets (Pfaff *et al.* 2022). From Cape Point to Cape Columbine, the shelf is narrow and steep before broadening and shallowing north of Cape Columbine. Thereafter, the shelf narrows north of St Helena Bay and widens again at 30°S which results in an offshore veering of the surface equatorward currents (Veitch *et al.* 2009).

Multiple facets of the oceanography of the SBUS such as the cyclonic circulation within St Helena Bay (Penven *et al.* 2000) or the various alongshore fronts (Lamont *et al.* 2015; Flynn *et al.* 2020) have been shown/hypothesized to locally enhance the retention of particles along the shelf (Fig. 4.1). However, very few studies have quantitatively computed or explored interannual and interseasonal variability in the residence times of particles across the entire shelf domain (Lett *et al.* 2006; Miller *et al.* 2006; Pfaff *et al.* 2022). The aim of this study is to understand the dynamics of retention along the SBUS. The objective is to use a physical ocean model of the region in conjunction with a Lagrangian tracking code to compute interannual residence times throughout the water column from a series of virtual Lagrangian float experiments. The methodology begins by detailing the model configuration, Lagrangian experiments and residence time computation. Thereafter, the results showcase particle trajectories along the shelf from different depths, cross-shore dispersion patterns and interannual summer and winter residence times. The discussion then explores (1) the topics of Lagrangian modelling in the SBUS. (2) the role alongshore fronts play in summer and winter on the residence times of particles along the shelf. Focus is also given to the role of local surface equatorward and deep poleward currents. The discussion ends by briefly exploring the variability in the regional wind regimes as well as remote forcing mechanisms as possible drivers of interannual variability in residence times along the shelf.

4.2 Methodology

4.2.1 Model configuration

The physical ocean model CROCO (Coastal and Regional Ocean Community model) based on ROMS (Shchepetkin & McWilliams 2005) was used to study the circulation and frontal dynamics in the SBUS. Built upon ROMS_AGRIF (Debreu *et al.* 2012), CROCO is a free-surface, terrain-following coordinate model that is well suited for regional applications and it solves the incompressible primitive equations based on the Boussinesq and hydrostatic approximations. The code used is that of CROCO v1.1 version available at <http://www.croco-ocean.org>. The model configuration of Chapter (3) is used here. In summary, the model was configured as a single domain with a horizontal spatial resolution of 3 km consisting of 60 vertical sigma levels. The full model domain extends from 8-20°E and 36-28°S (Fig. 3.1). For the study, daily model outputs from January 2004 – December 2018 were used from a zoomed portion of the full domain (Fig. 3.1 and 4.2) in order to focus on the shelf region of the SBUS. Evaluated against a combination of *in situ* and satellite data, the model was able to resolve the low-frequency interannual and intra-seasonal variability in SST and surface circulation features in the SBUS (Sec. 3.3.1). An artefact of the ocean boundary physical forcing conditions, the model resolved a very deep winter mixed layer depth

(MLD) and had warm temperature biases along the shelf region which impacted the position and intensity of the alongshore jets (Sec. 3.3.2).

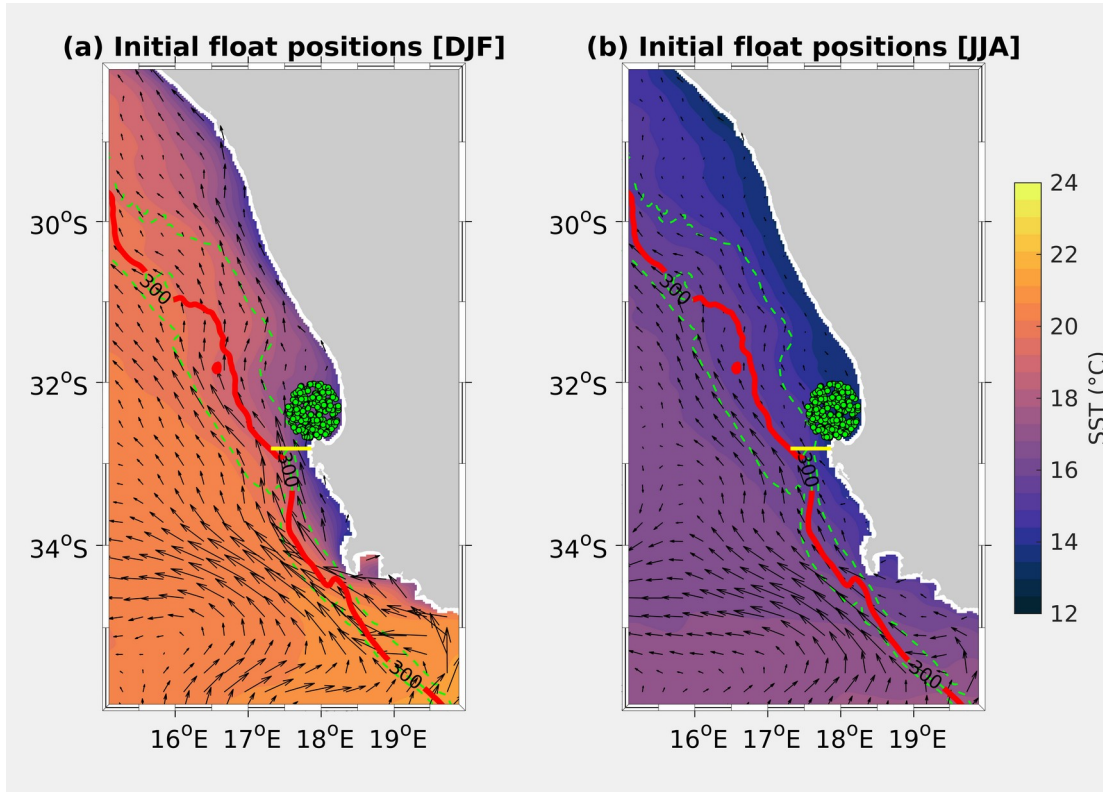


Figure 4.2: Seasonally averaged (a) summer and (winter) SST from 2004-2018 computed from daily CROCO outputs. Green dots are where the virtual Lagrangian floats are initiated for the residence time experiments. The yellow line demarcates the position of Cape Columbine and the red highlights 300 m isobath that delineates the maximum cross-shore extent of retained floats. Black arrows show the intensity of the seasonal surface geostrophic currents.

4.2.2 Lagrangian experiments

To assess the role alongshore hydrographic fronts play in restricting and/or enhancing the residence time of particles on the SBUS shelf, the ROMS offline tool (ROFF) was used (Capet, Marchesiello & McWilliams 2004; Carr *et al.* 2008) to seed and track virtual Lagrangian floats. ROFF uses a fourth-order Adams-Bashford-Moulton predictor corrector scheme to integrate Eq. (4.1)

$$\frac{d\mathbf{x}}{dt} = \mathbf{v}(\mathbf{x}, t) \quad (4.1)$$

over time given an initial condition $\mathbf{x}(t_0) = \mathbf{x}_0$ and 3D daily velocity fields from CROCO. ROFF is model specific for ROMS and has been employed in a number of studies encompassing: water mass pathways in the Tropical South Eastern Pacific (Espinoza-Morriberón *et al.* 2019), larval dispersion patterns along the Iberian Peninsula (Domingues *et al.* 2012) and investigating the fate of upwelled nitrate in the Peru-Chile Upwelling System (Hauschildt *et al.* 2021) to list a few. Three different Lagrangian experiments were run (Tab. 4.1): (1) comparing the model float trajectory paths to observations floats from the NOAA Global Drifter Programme (GDP); (2) residence time

experiments, with floats seeded in St Helena Bay and (3) a once-off backwards trajectory to investigate source waters along the shelf as well as the time-scale of transport for the poleward undercurrent. For all the experiments, ROFF was run with a time-step of 600 seconds and float positions were saved every 6-hours. The following sub-sections detail the experimental designs for (1) and (2) as they were the main focus of the study.

Table 4.1: List of Lagrangian experiments

Experiment	Duration [days]	Initiation date	Number of simulations	Number of floats	Location
Backwards	300	1 Jan 2008	1	1200	St Helena Bay
Residence times	90	1-31 Jan 2004-2018 1-31 Jul 2004-2018	930	1200	St Helena Bay
GDP_summer	11	8 Feb 2011	2	100	33.30 °S 16.80 °E
	16	25 Nov 2012			32.35 °S 16.68 °E
GDP_winter	54	2 Aug 2013	1	100	32.98 °S 17.75 °E

Comparing ROFF to observational data

In comparing the model float trajectories to the GDP, the objective was to not to evaluate ROFF but merely to briefly evaluate the Lagrangian dynamics within the model as a precursor to the residence time experiments. Model floats were evaluated against 6-hour spatially interpolated GDP data (Lumpkin & Centurioni 2019), which included the positions (longitude and latitude) as well as the potential temperatures of the floats. The interest herein was to evaluate the flow paths along the shelf region. Consequently, observational floats were only considered for the analysis if their trajectories were between the coast and the 500 m isobath. Furthermore, GDP floats were only considered if they were within the corresponding time-frame of the model and in order to evaluate the seasonal dynamics, only floats in summer (Dec, Jan, Feb) and winter (Jun, Jul, Aug) were used. Inevitably, the aforementioned criteria limited the number of observational floats (Tab. 4.1) available to compare against the model trajectories. Therefore, this did place a limitation on interpreting the performance of the model.

From Lumpkin *et al.* (2013), satellite tracked drifters of the GDP have a drogue centred at 15 m to better capture the near-surface currents. To replicate this dynamic, the GDP_summer and GDP_winter (Tab. 4.1) experiments were run with a fixed passive depth (refer to code of Carr *et al.* 2008) of also 15 m. Small variations in the initial position of floats or perturbations of oceanic conditions can significantly influence the trajectory path of a Lagrangian particle. To better account for the non-autonomous nature of the physical ocean model as well as the small number of observational floats, 100 virtual floats were deployed in each of the two model experiments (Tab. 4.1). These virtual floats were centred at the coordinates in Tab. (4.1) and initiated randomly within a 0.02° (22 km) radius. The time-frame over which the model experiments were run exceeded the duration over which the GDP floats were present in the model domain by usually a single day (Fig. 4.4). This was done to account for the 6-hour interpolation of the observational data and ensure the model floats were tracked for the corresponding length of time.

Residence time experiments

Virtual floats were initiated in St Helena Bay (Fig. 4.2), with 200 released at discrete depths from the surface (-1 m) to the bottom (-250 m) in 50 m intervals, totalling 1200 floats per release. Summer experiments were conducted in January and winter experiments in July. To calculate the residence times, a simple decay function (Eq. 4.2) similar to Piñones *et al.* (2011) was used.

$$f(t) = f_0 e^{-\lambda t} \quad (4.2)$$

In Eq. (4.2), $f(t)$ is the number of floats on the shelf after t time in days from initialisation while f_0 is the initial number of floats released at a particular depth and λ is the decay constant. The 300 m isobath was assumed to delineate the location of the shelf-break jet (Fig. 4.2) and floats were considered to be removed from the control volume if they moved south of Cape Columbine (32.83°S), north of 28°S and/or spent 1 day beyond the 300 m isobath. The residence time t was computed when $f(t)/f_0 = 1/e$ and using the toolbox of Moisy (2023), λ was empirically fitted. A goodness-of-fit for the fitted curve was evaluated using Eq. (4.3)

$$R^2 = \frac{\sum (\hat{y} - \bar{y})^2}{\sum (y - \bar{y})^2} \quad (4.3)$$

where $\hat{y}$ is the fitted value for every corresponding data value y and $\bar{y}$ is the mean. R^2 values close to 1 are indicative of a good fit between the data and empirically fitted curve.

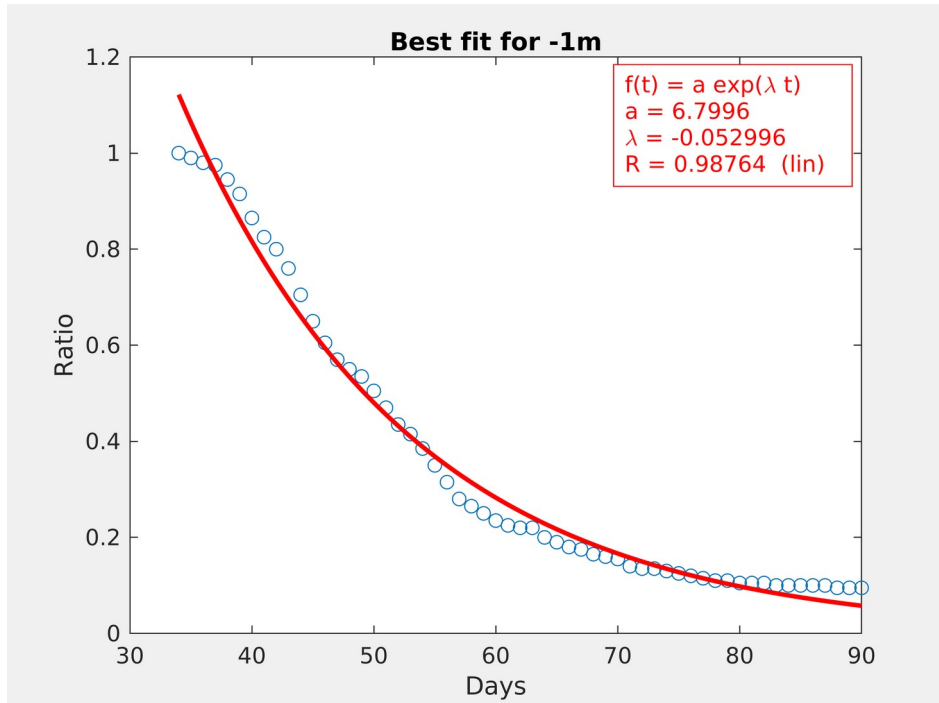


Figure 4.3: Example fit. The red line shows the fitted curve while the blue dots are the data points.

On analysis, floats did not immediately move off the shelf and would take between 10-15 days before the first ones did so. Consequently, to ensure an intercept as close to 1 as possible, Eq. (4.2) was reformulated into a piecewise function (Eq. 4.4).

$$\begin{cases} f(t) = 1 & t \leq c \\ f(t) = e^{-\lambda t} & t > c \end{cases} \quad (4.4)$$

In Eq. (4.4), c was defined as the time-period over which the number of floats/particles remained constant on the shelf after initialisation such that the residence times were computed using a simple addition with $c : t = c + 1/\lambda$. An example is shown in Fig. (4.3) for particles released at -1 m during summer. Using the piecewise function of Eq. (4.4) allowed the fitted curve to better capture the exponential decay and therefore more accurately compute the residence times. In conjunction with the residence times, the zonal, meridional and total dispersion (Eq. 4.5) of the floats were computed following the formula in LaCasce (2008).

$$D(t) = \frac{1}{2N(N-1)} \sum_{i \neq j} [\mathbf{x}_i(t) - \mathbf{x}_j(t)]^2 \quad (4.5)$$

In Eq. (4.5) (LaCasce 2008), $D(t)$ is the time evolution of dispersion (km^2), N is the total number of floats and $\mathbf{x}(t)$ is the displacement of a float from its initial position such that dispersion is computed as the sum over all particle pairs.

4.2.3 Front detection and diagnostics

The front detection method and various computed frontal diagnostics are similar to those from Chapter (3). The Canny algorithm (Canny 1986) was used to identify frontal features along the shelf region of the SBUS north of Cape Columbine, bounded by the coast and the 300 m isobath. Once the frontal edges were identified from daily SST fields, two diagnostics were collected from each daily edge map: the number of fronts and the mean offshore distance of the fronts. For each front detected on a particular day, the offshore distance of the centroid was computed parallel to the coastline and thereafter the mean offshore distance was simply the mean of all the detected fronts' distances (Eq. 3.5).

4.3 Results

4.3.1 Lagrangian particle evaluation

In evaluating the Lagrangian dynamics for summer and winter conditions along the SBUS shelf, Fig. (4.4) and Tab. (4.2) summarize the model's performance relative to the GDP data. In Fig. (4.4), model float trajectories are compared to observational buoy tracks along the shelf. All observational and model trajectories for summer (Fig. 4.4a) and winter (Fig. 4.4b) show a distinct north-westward propagation; consistent with the dominant surface equatorward jet patterns for the region. In summer, the two observational buoys have different initial tracking locations, with the 2011 buoy located along the edge of the 500 m isobath off Cape Columbine and the 2012 buoy being further north and on the oceanic side of the 300 m isobath (Fig. 4.4a). The buoys' trajectories correspond to the oceanic extremity of the shelf-edge jet. Both sets of summer model trajectories show a more pronounced westward propagation relative to the observational tracks resulting in model floats

being quickly transported offshore. Furthermore, the respective model clusters show minimal relative dispersion and variability in the trajectories of individual floats.

In winter (Fig. 4.4b), the single observational buoy consistently follows the 200 m isobath as it moves north-westward along the shelf whereas multiple divergent pathways are evident for the model float trajectories. A portion of the model's floats follow a defined nearshore jet, consistent with the observations; however, the intensity of this jet is weaker within the model. A second equatorward jet is evident along the 300 m isobath and further offshore a number of floats are entrained by an anti-cyclonic eddy that propagates the floats offshore. Unlike summer (Fig. 4.4a), where the individual model float trajectories are fairly homogeneous, multiple divergent pathways are present in winter. This likely stems from the jet features in summer being more cohesive in structure than those in winter. Consequently, a difficulty then when trying to compare model and observational Lagrangian data is accounting for small variations in initial location which can drastically alter the trajectory pathways of floats.

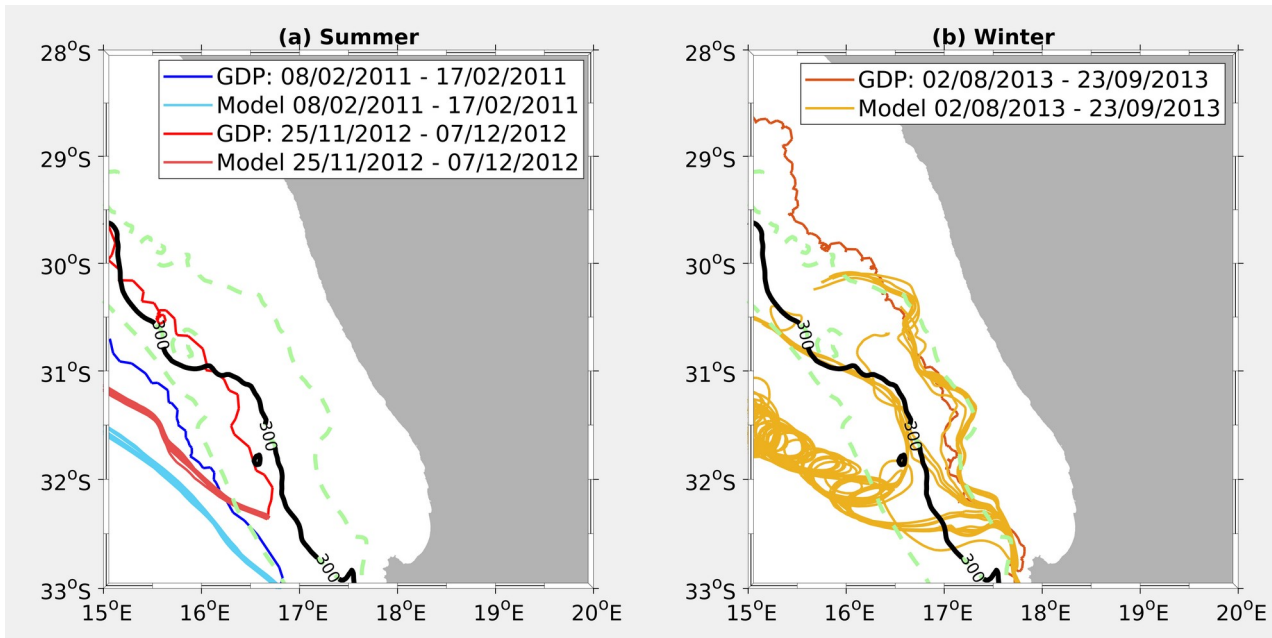


Figure 4.4: Summer (a) and winter (b) of model vs GDP float trajectories along the SBUS shelf. The 300 m isobath is shown as the solid black contour while the 200 and 500 m isobaths are dashed green.

From Tab. (4.2), the model has a positive temperature bias, which is more pronounced in summer than winter, relative to the observations. A comprehensive evaluation of the model configuration in Chapter 3, Sec. (3.3.1), showed the intensity and position of the alongshore jets over the shelf to be affected by the warm temperature bias in the model that resulted in the alongshore jets being less intense and found closer to the coast compared to observations. The near coastal regions were also shown to experience more intense currents due to a negative temperature bias along the coast.

In summer, the 2011 observational buoy deployed along the shelf edge experienced the greatest temperature bias relative to the model's floats but the intensity of the surface currents were similar for both (Tab. 4.2). The alongshore shelf-edge jet in the SBUS is maintained by the across-shore temperature gradients between the cooler shelf waters and warmer offshore waters that are supplied by Agulhas leakage. Along the shelf edge, though a temperature bias is present within the model, the gradient must be similar. In contrast, the 2012 observational summer buoy also had a positive

temperature bias relative to the model but the resolved velocities were significantly weaker for the model floats. Although the model's floats were not prominently on the shelf itself, an explanation for the more westward propagation of the model floats, seen in Fig. (4.4a), may be due to the shelf-edge jet being found closer to the coast within the model. Consequently, the model float trajectories are mainly being driven by the oceanic flow patterns of the Benguela Current and not the shelf-edge jet itself.

Table 4.2: Statistical metrics for model trajectories and observational buoys

Date	Summer				Winter	
	Model	Obs	Model	Obs	Model	Obs
	08/02/2011 - 08/02/2011 - 25/11/2012 - 17/02/2011	17/02/2011	07/12/2012	07/12/2012	02/08/2013 - 23/09/2013	02/08/2013 - 23/09/2013
Temperature [°C]	21.09	20.04	17.76	17.10	14.40	14.38
Temperature std [°C]	0.041	0.52	0.013	0.59	0.26	0.45
Velocity [m/s]	0.43	0.39	0.22	0.41	0.17	0.17
Velocity std [m/s]	0.018	0.21	0.012	0.19	0.088	0.085

In winter, the temperature bias is negligible and averaging the velocities across the multiple float trajectory pathways show the model as having near identical current intensities compared to the observational buoy (Tab. 4.2). From the nearshore track in Fig. (4.4b), the model floats did not travel the same distance as the observational buoy over the same release period. Inspecting the intensity of currents along the shelf, the offshore flow was > 0.2 m/s and between the coast and the 200 m isobath, currents were on average 0.12 m/s; making them 30% weaker than the observations.

From Fig. (4.4) and Tab. (4.2), the model does show a capacity to resolve the broad surface flow patterns of floats, especially in winter. In summer, floats show a more significant westward offshore propagation that will negatively impact the residence times of particles along the shelf edge. The float release experiments initialised in St Helena Bay to compute residence times are not likely to be affected by this westward bias. From Chapter (3), the model resolves weaker shelf based alongshore currents and jets that are in closer proximity to the coast relative to observations. This will artificially inflate the residence times computed for surface particles.

The presented model vs data comparison is only conducted for the surface ocean, as no observational floats could be used to validate flow patterns outside of the upper 15 m. In Sec. (3.3.1), the poleward flow in the model is shown to be less intense compared to observations (Fig. 3.10). A weaker poleward undercurrent in the model will also lead to elevated residence times being computed for particles released at depth. With these limitations in mind, the following section explores Lagrangian pathways at different depths along the SBUS shelf and Sec. (4.3.3) quantitatively computes residence times during summer and winter.

4.3.2 Lagrangian pathways along the SBUS shelf

As seen in the previous section (Fig. 4.4), the trajectories of model floats differ between summer and winter as floats in summer show similar flow paths to each other and rapid northward transport compared to the more dispersive and slower trajectories in winter. An ensemble of model float trajectories for summer and winter, initiated throughout the upper 250 m of the water column, reveal that particle trajectories exhibit interannual and intramonthly variability along the SBUS

shelf. Exploring this further, Fig. (4.5-4.8) show ‘typical’ examples of surface (-1 and -50 m) and deep (- 150 and – 250 m) Lagrangian float trajectories for summer and winter.

Surface trajectories

Fig. (4.5) and Fig. (4.6) showcase examples of ‘typical’ float trajectory patterns for summer and winter respectively within the upper 50 m along the shelf. In summer (Fig. 4.5), floats initiated at -1 m collectively maintain a near uniform trajectory pathway, remaining confined to the coastal side of the 300 m isobath, as they are rapidly transported northward. This uniform behaviour differs to particles released at -50 m (Fig. 4.5b) as floats show longer and more variable transit times along the shelf. The elevated transit times stem from a decrease in intensity of the equatorward jets with depth as well as enhanced retention within St Helena Bay due to the cyclonic circulation. Furthermore, floats released at -50 m exhibit more cross-shore movement, with some propagating offshore despite the fact that offshore zonal velocities are greater at the surface than at – 50 m. The alongshore fronts could be playing a greater role in the surface layer than for particles at -50 m in restricting the offshore advection of floats; however, as floats in Fig. (4.5a) traverse the length of the shelf in half the time as those in Fig. (4.5b) it is likely that floats initiated at -1 m do not show the same cross-shore displacement simply because they are too rapidly transported northward and removed from the model domain.

Contrasting Fig. (4.5c) and Fig. (4.5d), floats initiated at -50 m exhibit greater sensitivity to upwelling activity along the shelf than for particles at -1 m. A shallow summer MLD maintains a well stratified water column for the upper 30 m that acts to rapidly dissipates small mesoscale features and perturbations. Consequently, intense vertical velocities in Fig. (4.5c) are isolated to defined frontal features. In Fig. (4.5d), vertical velocities exhibit a typical cross-shore pattern where upwelling occurs close to the coast and further offshore towards the 300 m isobath, downwelling occurs. The cross-shore temperature dynamics in Fig. (4.5e) are typical for the surface layer, increasing towards the shelf edge. Complementing the vertical dynamics in Fig. (4.5d), Fig. (4.5f) shows cold ($< 14^{\circ}\text{C}$) and recently upwelled deep waters within St Helena Bay that gradually warm as they are displaced upwards towards the surface and transported northward.

In winter (Fig. 4.6), surface currents are significantly weaker and the oceanic environment is more dispersive than summer, resulting in longer residence times and more cross-shore displacement for floats on the shelf. Floats initiated at -1 m (Fig. 4.6a) show a more rapid northward transport than those at – 50 m (Fig. 4.6b); however, the transit times of floats along the shelf in Fig. (4.6a) are more variable compared to those in summer (Fig. 4.5a). In winter, narrow jets only locally enhance transport along portions of the shelf and the less cohesive alongshore fronts, which are found further offshore, permit greater cross-shore displacement of floats than those in summer. It is interesting that the float trajectory paths and time-scales in Fig. (4.6b) are comparable to those observed in summer (Fig. 4.5b).

Seasonally averaged cross-sections of meridional velocities along the shelf (Supp. 3.4S) show the intensity of the equatorward flow to be similar at – 50 m for summer and winter. Dynamically, 50 m falls below the mean summer MLD, where the intense equatorward currents are confined to the upper 30 m of the water column, but above the winter MLD which is approximately 60 m on average within the model (Fig. 3.9). The seasonal switch of particles from being below the MLD in summer to above in winter is likely responsible for the similar behaviour of the float trajectories. In Fig. (4.6c), vertical velocities are weak for particles initiated at -1 m and from Fig. (4.6d), intense

vertical velocities are confined to being along the coast as well as along the SBF. A reduction in upwelling intensity along the shelf is expected in winter due to the predominant alongshore winds becoming north-westerly. In Fig. (4.6e) and Fig. (4.6f) the across-shelf temperatures are rather homogeneous due to a reduction in insolation as well as a decrease in the stratification of the water column due to winter mixing. The absence of defined cross-shore temperature gradients within the surface layer (Fig. 4.6e) is primarily responsible for the weaker surface currents.

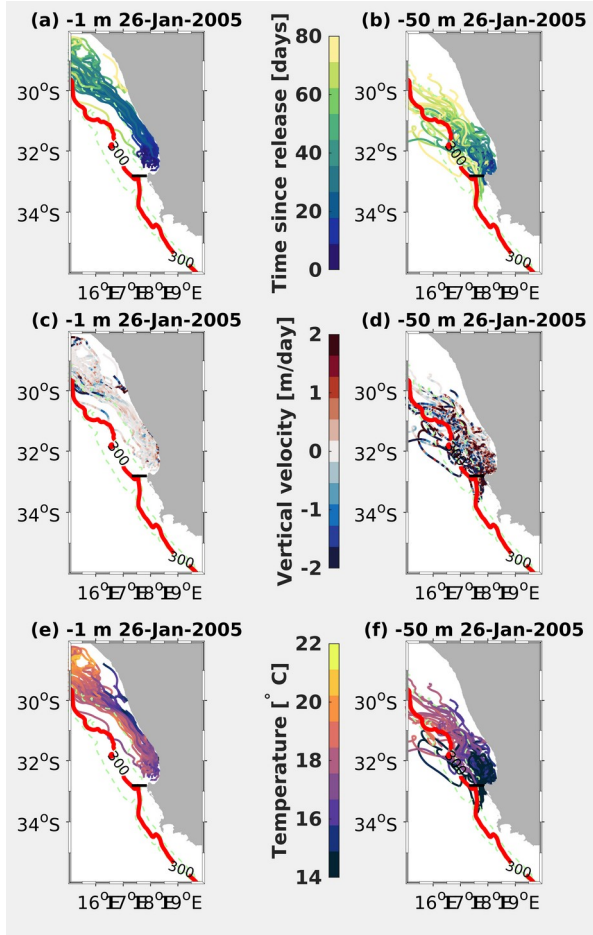


Figure 4.5: Example summer trajectories of model floats tracked for 90 days from release. Left column is particles initiated at the surface (-1 m) and right is for -50 m. Top row shows days since release of floats, middle row is vertical velocities and bottom is temperature.

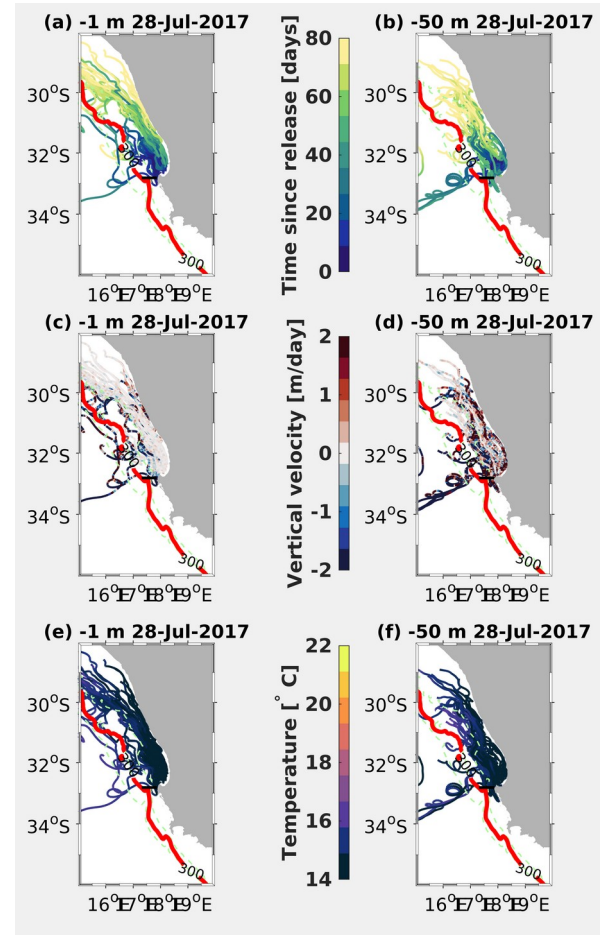


Figure 4.6: Example winter trajectories of model floats tracked for 90 days from release. Left column is particles initiated at the surface (-1 m) and right is for -50 m. Top row shows days since release of floats, middle row is vertical velocities and bottom is temperature.

Fig. (4.5) and Fig. (4.6) showcase typical flow pathways for material in the upper ocean for summer and winter respectively. It is important to appreciate that float trajectories can be influenced by a myriad of event scale and circulation features. In both seasons, an enhanced cyclonic circulation within St Helena Bay can elevate the retention time of particles throughout the upper water column. In summer, reversals in the nearshore currents during wind relaxation events can result in surface particles being transported around Cape Columbine and poleward along the narrow shelf. Furthermore, variability in the dispersion and cross-shore displacement of particles can be a function of the offshore position, width and intensity of the alongshore jets. Therefore, intra-seasonal variability in residence times for the surface layer along the shelf can be driven by

variability in the alongshore frontal patterns, intensities of the surface jets as well as variability in offshore Ekman transport.

Deep trajectories

Similar to the previous section, Fig. (4.7) and Fig. (4.8) showcase ‘typical’ trajectory patterns for summer and winter respectively for particles initiated at -150 m and -250 m (or shelf bottom) along the shelf. In summer, Fig. (4.7a) and Fig. (4.7b) show similar transit times for floats as they are transported poleward along the narrow shelf towards Cape Town. The similar times for particles initiated at -150 m and -250 m can be attributed to the broad poleward undercurrents present along the shelf. For shallower floats that move close to the shelf edge (Fig. 4.7a), they can be influenced by the oceanic branch of the Benguela Current. The deep penetrating turbulent regimes of mesoscale eddies can entrain these floats and transport them westward along the Agulhas eddy corridor. From an ensemble of studied trajectories, there was no example where deep initiated floats that were transported back north re-entered the shelf region north of Cape Columbine.

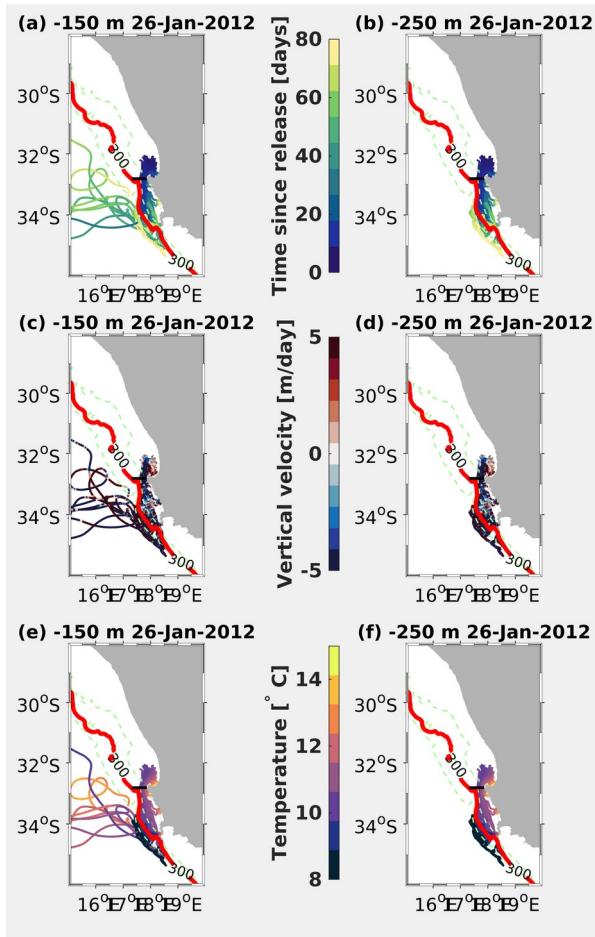


Figure 4.7: Example summer trajectories of model floats tracked for 90 days from release. Left column is particles initiated at -150 m and right is for -250 m. Top row shows days since release of floats, middle row is vertical velocities and bottom is temperature.

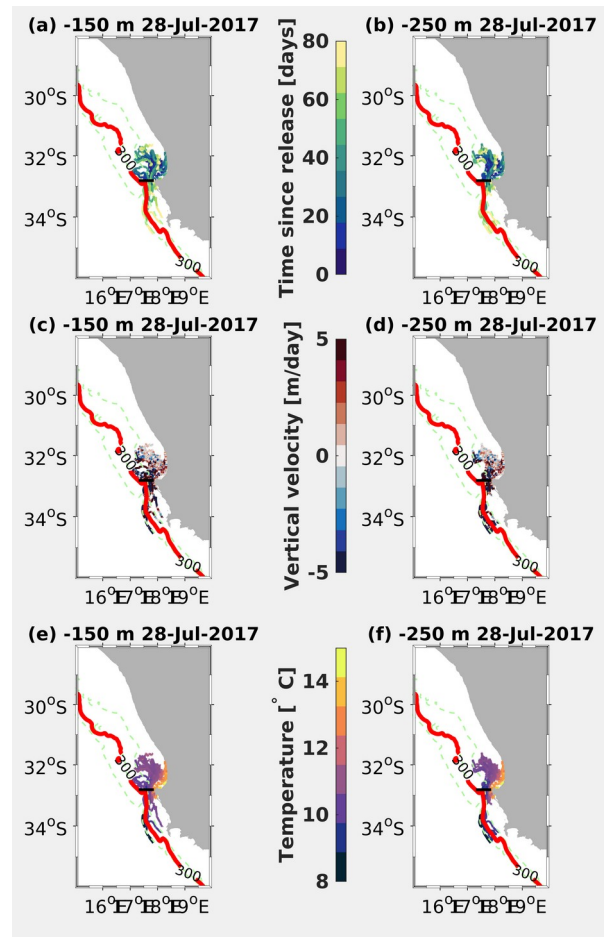


Figure 4.8: Example winter trajectories of model floats tracked for 90 days from release. Left column is particles initiated at (-150 m) and right is for -250 m. Top row shows days since release of floats, middle row is vertical velocities and bottom is temperature.

The cross-shore patterns of vertical velocities in Fig. (4.7c) and Fig. (4.7d) are similar, where positive vertical velocities are prevalent close to the coast while along the narrow shelf region, negative velocities are associated with the downwelling of particles along the intense shelf-edge jet. The confinement of particles along the narrow shelf region between Cape Town and Cape Columbine is likely due to the presence of the CP jet, which in large part restricts the offshore advection of particles as they are topographically steered by the poleward undercurrents. Evidence of subduction along the frontal interface can be seen in Fig. (4.7e) and Fig. (4.7f), where the temperature of the floats significantly drops on the oceanic side of the 300 m isobath. In Fig. (4.7e), floats that move offshore show a temperature increase, associated with the warmer oceanic waters and also being vertically displaced closer to the surface layer.

In winter (Fig. 4.8), the Lagrangian trajectories of floats are rather similar to those in summer. Floats in winter seem to be more pronouncedly influenced by the cyclonic circulation within St Helena Bay and a reduction in the intensity of the poleward undercurrents influences the extent of southward transport along the shelf. For particles initiated at -150 m (Fig. 4.8a) and -250 m (Fig. 4.8b), both show distinct cyclonic trajectories within St Helena Bay. Some floats can be permanently trapped within the bay, while the vast majority are transported along the narrow shelf south of Cape Columbine. As expected, weaker vertical velocities are observed in winter with strong negative velocities occurring along the 300 m isobath associated with the presence of the shelf-edge jet (Fig. 4.8c and d). Though not present in Fig. (4.8), particles initiated at -150 m can also become entrained and transported westward by mesoscale eddies; as seen in summer. From Fig. (4.8e) and Fig. (4.8f), the cyclonic transport of floats with St Helena Bay results in some floats being upwelled along the southern portion of the bay. This upwelling results in floats warming as they are moved into the upper layers of the water column. Furthermore, the drop in float temperatures along the 300 m isobath adds further evidence to the downwelling of floats along the CP jet.

Fig. (4.7) and Fig. (4.8) showcase typical Lagrangian float patterns for deep trajectories along the shelf. The float trajectories are primarily driven by the poleward undercurrents but the enhanced retention of particles due to the cyclonic circulation within St Helenan Bay is an important source of variability. Similar to the upper ocean, interannual and intramonthly variability is present within the lower 100 -250 m of the water column. Variability in the intensity of the poleward undercurrents as well as the prominence of the cyclonic circulation with St Helena Bay affect the trajectories of floats in both seasons as well as their residence times along the shelf.

4.3.3 Summer and winter residence times

The Lagrangian patterns of floats exhibit interannual and intramonthly variability for summer and winter within the upper 50 m and lower 100 - 250 m of the water column in the SBUS. Floats in the upper 50 m are transported northward by the prominent surface equatorward currents and jets while floats below 100 m are primarily transported south along the narrow shelf region between Cape Town and Cape Columbine by the poleward undercurrents. The spatial variability of the float trajectories can be influenced by several factors. A main one is variability in the alongshore spatial patterns of fronts on the shelf as well as their intensities. The intensity of the cyclonic circulation within St Helena Bay is another important feature that influences the retention of particles as well as the poleward undercurrents along the shelf-bottom. It is necessary then to understand how these various shelf based oceanographic features impact the residence times of floats along the shelf in summer and winter.

Dispersion patterns along the shelf

Dispersion is a measure of how particles spread out over a time interval from an initial point of release. Using Eq. (4.5), the absolute dispersion of floats with respect to their initial release location are computed and shown in Fig. (4.9). The total dispersion of floats is significantly greater in summer than winter (confirmed using a t-test at an alpha level of 0.05). In the upper 50 m (Fig. 4.9a) of the water column, the zonal component is dominant while the lower 100 – 250 m (Fig. 4.9b) shows the meridional component as being the controlling factor for dispersion¹. Integrating total dispersion over 90 days across the various discrete release depths for all the summer and winter Lagrangian experiments show the upper 50 m of the water column to be a factor of 10 times greater than the lower 200 m. The decrease in magnitude of particle dispersion with depth is unsurprising given the fact that the more intense meridional and zonal velocities are associated with the surface layer. Furthermore, plots of integrated dispersion for 2004-2018 (Supp. 4.1S) show interannual variability throughout the water column during summer and winter. Interannual variability in the frontal spatial patterns, intensity of the surface equatorward jets and poleward undercurrents as well as the magnitude of offshore Ekman transport can all impact the total dispersion of particles along the shelf.

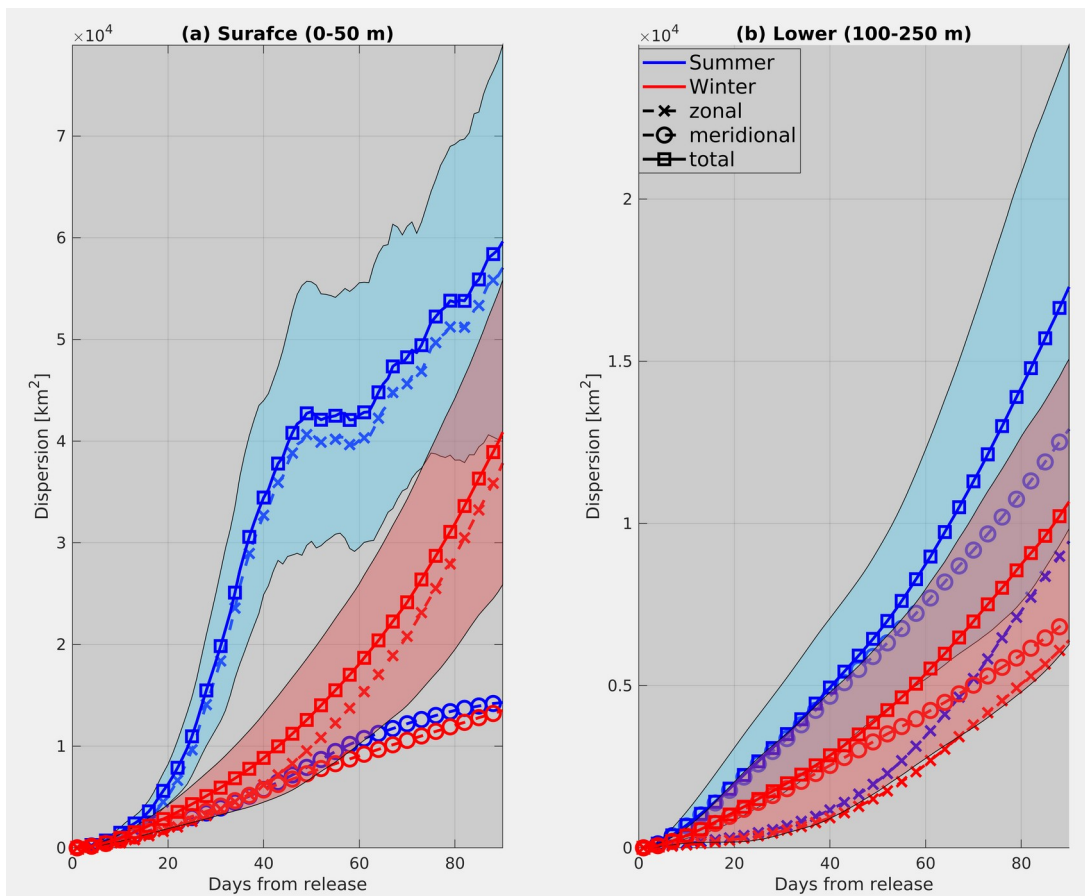


Figure 4.9: Seasonally averaged dispersion from 2004-2018 with the shading being $\pm$ one standard deviation of the total dispersion averaged over the upper 50 m and lower 150 m of the water column.

¹ The kink in Fig. (4.9a) for the summer zonal and total dispersion curves is not due to spurious physical dynamics but instead a mathematical feedback in the calculation that arises due to floats leaving the model domain.

For the upper 50 m (Fig. 4.9a), the total dispersion of floats in summer and winter is dominated by the zonal component. This at first contradicts the aforementioned transport regimes from Sec. (4.3.2) where the equatorward jets are identified as being the major features influencing the float trajectories. A closer inspection of the temporal evolution of particle dispersion (Fig. 4.9) over the course of the float trajectories reveals why this is so. The meridional and zonal components for summer and winter contribute near equally to the total dispersion for the first 20 and 40 days respectively after release. After such time, the zonal component becomes the major driver of dispersion along the shelf. From Fig. (4.5) and Fig. (4.6), particles show elevated cross-shore displacement upon moving north of 30°S as they then enter the broad Namaqua shelf and experience elevated westward propagation due to the offshore veering of the surface equatorward currents. Therefore, the patterns of dispersion show topographical controls along the shelf. The narrowing of the shelf north of St Helena Bay results in the intense SBF being found closer to the coast, restricting the cross-shore displacement of particles. The broadening of the shelf north of 30°S influences the positions and flow patterns of the alongshore jets, resulting in the zonal component dominating the dispersion of floats.

In contrast to the surface layer, Fig. (4.9b) shows the dispersion of particles between 100-250 m to be dominated by the meridional component in both summer and winter. Less particle dispersion in winter arises merely due to a reduction in intensity of the poleward undercurrents, which are the major oceanographic features responsible for the transport of floats along the shelf-bottom and lower water column. In Fig. (4.9b), all the dispersion curves show an expected exponential growth. The zonal components grows rapidly after 40-50 days for summer and winter. From the trajectories in Fig. (4.7) and Fig. (4.8), summer particles exhibit greater zonal dispersion as they move along the narrow shelf towards Cape Town while in winter, the time period at which the zonal component dominates, corresponds to when particles have almost completed a full cyclonic rotation within St Helena Bay. As most particles initiated in summer leave St Helena Bay and are transported poleward, there is more opportunity for floats to spread out along the narrow shelf or potentially move offshore which elevates the zonal dispersion. In winter, a portion of floats are permanently retained within St Helena Bay while others flow south and this divergence in trajectories likely elevates the zonal dispersion. Therefore, the dispersion patterns for deep initiated particles in both seasons are controlled by the intensity of the poleward undercurrents as well as the intensity of the CP jet as it limits the cross-shore advection of particles. Furthermore, the cyclonic circulation within St Helena Bay elevates float retention along the shelf, especially in winter.

Residence times

Tab. (4.3) shows the seasonally averaged residence times (Sec. 4.2.2) computed for summer and winter on the SBUS shelf using Eq. (4.4). As expected, the residence times for particles released in summer are significantly lower than those in winter. The more intense zonal and meridional flow patterns (Fig. 4.9) rapidly transport and remove floats from the control volume constituting the shelf environment north of Cape Columbine. Floats released at -50 m show similar residence times for both summer and winter (confirmed from a t-test assuming unequal variance at an alpha level of 0.05). From Sec. (4.3.2), the transport of floats in the upper 50 m is dominated by the surface equatorward jets and below 100 m by the poleward undercurrents. Within a control volume, it is not possible to directly compare the residence times of deep and surface initiated particles; however, a once off backwards trajectory (Supp. 4.2S) shows that it takes approximately 300 days for the poleward undercurrents to transport particles from the Namaqua shelf into St Helena Bay.

Table 4.3: Seasonally averaged residence times for summer and winter floats for discrete depths along the shelf

Summer			Winter	
Release depths [m]	Residence time [days]	R ²	Residence time [days]	R ²
-1	41.27 ± 4.69	0.807 ± 0.086	146.36 ± 59.98	0.889 ± 0.115
-50	97.11 ± 27.99	0.878 ± 0.088	99.73 ± 34.94	0.953 ± 0.061
-100	66.73 ± 17.88	0.987 ± 0.110	140.87 ± 104.94	0.946 ± 0.089
-150	52.40 ± 19.95	1.039 ± 0.116	124.90 ± 63.94	0.929 ± 0.089
-200	53.60 ± 22.17	1.029 ± 0.113	130.75 ± 64.51	0.923 ± 0.087
-250	53.65 ± 22.64	1.029 ± 0.113	130.78 ± 65.02	0.923 ± 0.087

The R² values in Tab. (4.3) show the coefficient of determination relative to the non-linear function of Eq. (4.2). Particles initiated at -1 m show the lowest R² scores and for deeper particles (> 100 m) the fits are very close to and slightly greater than 1. According to Kvålseth (1985), the definition of R² employed in Eq. (4.3) has a favourable quality of generally being non-negative but has a tendency to exceed 1. Nevertheless, the calculated R² values are greater than those of Piñones *et al.* (2011) and thus deemed acceptable for quantifying the residence times.

A two-sided t-test assuming unequal variance at an alpha level of 0.05 shows particles released at -1 m and -50 m in summer to have statistically distinct residence times; unlike in winter (Tab. 4.3). The shallow summer MLD limits the vertical extent of the intense equatorward jets to the upper 30 m which impacts the transport dynamics of floats at -50 m. In winter, the deepening of the MLD results in the -1 m and -50 m floats being within the same stratified layer which results in statistically similar residence times. It is this seasonal shift in the position of the MLD that likely generates the almost identical residence times of floats initiated at -50 m in summer and winter respectively. Conducting the same t-test for floats below 100 m, for summer and winter respectively, there is no statistically significant difference in the residence times of floats between -100 m and -250 m. The similar residence times are due to the broad poleward undercurrents along the shelf which dominate the transport of floats (Fig. 4.9). Furthermore, for deep initiated floats, the residence times in summer are three times smaller than in winter due to the elevated retention of floats within St Helena Bay (Fig. 4.8) as well as the less intense poleward undercurrents in winter.

Complementing Tab. (4.3), Fig. (4.10) shows interannual residence times for summer and winter for the various discrete release depths along the shelf. Far from being a constant value, the residence times for summer and winter show a somewhat cyclic pattern, where maxima in residence times occur at intervals of 4-5 years. Furthermore, this pattern is not restricted to a single depth but seems to be a transitive property of the whole water column. Clear examples of this are 2008 and 2016, where the residence times for summer throughout the water column are all above average. For winter, the same pattern holds for 2014 and 2017; however, floats at -1 m (Fig. 4.10a) do not show the same trend. As expected, the winter residence times throughout the water column are usually longer than those in summer. For the upper 50 m, it is interesting that in Fig. (4.10a) for 2004, 2007 and 2011 the winter residence times were similar, if not lower, than the corresponding summer residence times. In Fig. (4.10b), the summer and winter residence times are of a similar magnitude and the greatest residence times for a particular year can oscillate from being in summer to winter. From Fig. (4.10c-f), for particles below 100 m, the winter residence times can be three times greater compared to summer and there is no example where the winter residence times are shorter than those in summer. Therefore, there is a need to further investigate this interannual signal.

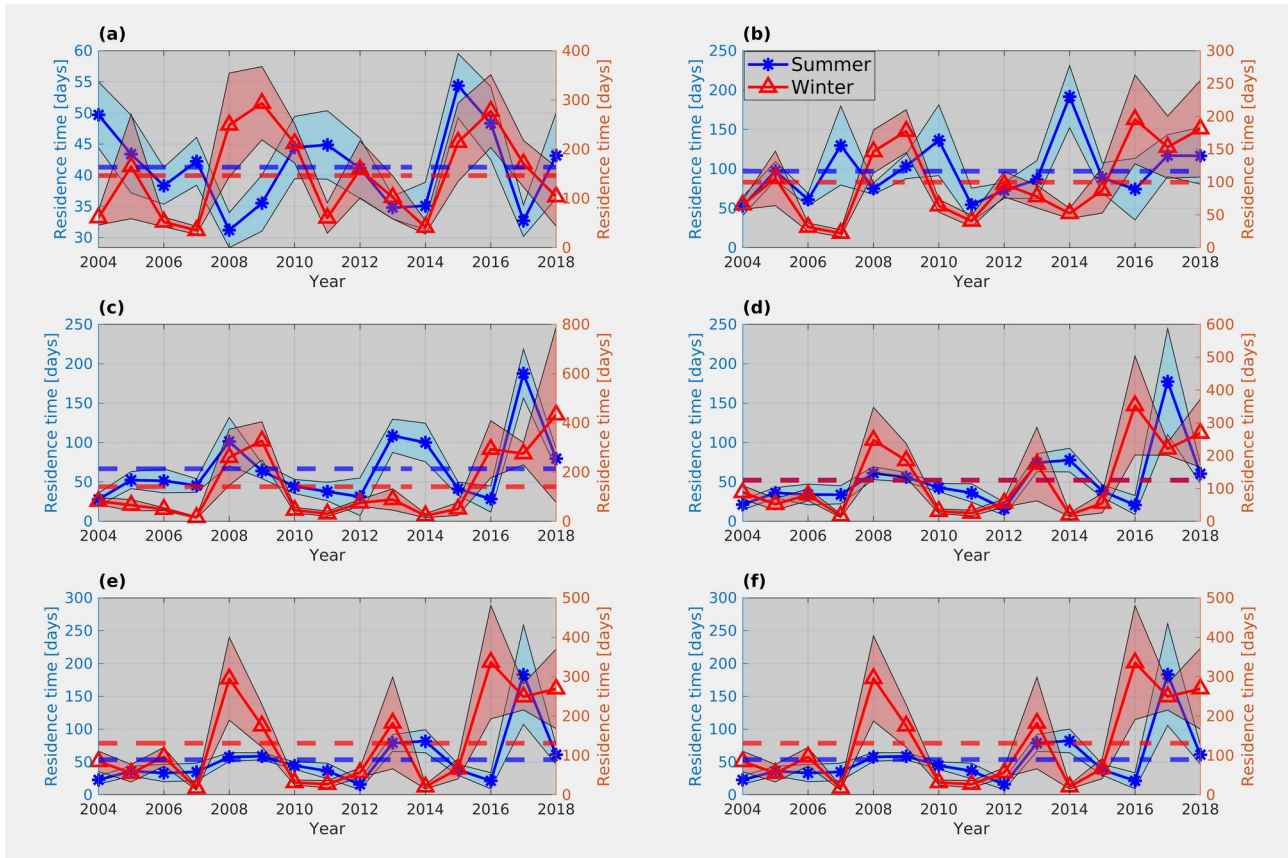


Figure 4.10: Seasonally averaged residence times for summer (blue) and winter (red) for 2004-2018. Plots (a-f) respectively correspond to the depths of release: -1, -50, -100, -150, -200, -250 m. Shaded region corresponds to $\pm$ one standard deviation and the horizontal blue and red dashed line represent the long-term seasonal mean retention. Summer retention (in days) is read from the left axis while it is read from the right for winter.

To further explore the interannual signal for the residence times (Fig. 4.10), a power spectral analysis of the alongshore wind field (not shown) was conducted. The spectral analysis showed a spike at 12 months and another spike corresponding to a period of approximately 5 years; which is typical of the periodicity of ENSO. To investigate a possible link between ENSO and the computed residence times, the Southern Oscillation Index (SOI) for 2004-2018 from NOAA (<https://www.noaa.gov/>) was used. For the SOI, negative/positive values coincide with El Niño/La Niña episodes. To compare the interannual summer and winter residence times (Fig. 4.10) with the SOI, the SOI is averaged over January-March and July-September to correspond to summer and winter periods respectively. Thereafter, the SOI is correlated against the summer and winter residence times from Fig. (4.10). In summer, a negative correlation of -0.46 ($p \approx 0.08$) is seen for residence times computed at -1 m and -0.34 for -50 m ($p \approx 0.22$). Throughout the rest of the water column in summer and for all depths in winter p-values > 0.3 such that there is no relation. Though not significant at the 95% level, the pronounced negative correlations for the upper 50 m in summer give some evidence for ENSO impacting the retention dynamics of the SBUS and if anything adds robustness to the approach used to compute the residence times because they show a correspondence to remote forcing signals.

In general, spikes in higher residence times correspond to periods of higher variability. Correlating the seasonal anomaly of residence times from Fig. (4.10) with the seasonal anomaly of integrated

dispersion (Supp. 4.1S) show an expected negative relationship during both seasons. Greater dispersion should result in smaller residence times as floats are either advected off the shelf or rapidly removed from the model domain via the intense surface equatorward jets. It is noteworthy that elevated summer retention times for a particular year do not precede the same behaviour in winter. Looking at the summer vs winter correlations for the different depths, correlations were close to 0 for the upper 50 m and ≈ 0.3 on average from 100 m and below; although they were not statistically significant ($p > 0.05$). Exploring this further, doing a cross-correlation of the summer and winter time-series show that from -1 m to -100 m, the maximum correlation always occurs at 0 lag; however, for -150 m and below, correlations peaked at a lag of +1.

Role of alongshore fronts and transport dynamics on residence times

From Fig. (4.10), it is clear that the residence times for floats show interannual variability and are significantly longer during winter. The negative correlation of residence times with dispersion, though expected, highlights how variability in the physical oceanography of the SBUS impacts the transport and mixing patterns for material on the shelf. From Fig. (4.9), floats in the upper 50 m show the greatest dispersion arising from the zonal component while for particles below 100 m, total dispersion is dominated by the meridional component. Therefore, it stands to reason that changes to the intensity of the offshore flow regimes in the upper ocean and/or variability in the intensity of the poleward undercurrents will have definite effects on the residence times of floats.

Alongshore fronts are not passive features and instead show significant variability in position, width and intensity (Sec. 3.3.3). In summer, alongshore winds drive upwelling along the shelf which impacts the spatial patterns of SST and steepens the cross-shore temperature gradients along the shelf which are crucial for frontogenesis. From Fig. (4.5-4.8), it is evident that the offshore position of a front plays a major role in controlling the cross-shore movement of floats. Interannual variability in the spatial patterns of fronts, especially during summer, can be linked to the variability in the alongshore winds. This relationship is advantageous because the alongshore winds can also be used to compute offshore Ekman transport. Ekman transport will serve as a proxy for the intensity of zonal dispersion along the shelf. With all this considered, Fig. (4.11) and Fig. (4.12) interrogate how the interannual variability in the offshore position of the alongshore fronts, Ekman transport and poleward undercurrents impact the residence times of floats within the upper 50 m and lower 200 m of the water column.

In Fig. (4.11), the monthly anomalies of Ekman transport, the mean offshore position of the alongshore fronts and intensity of the poleward undercurrents show interannual variability. For Ekman transport (Fig. 4.11a), negative anomalies correspond to periods of elevated offshore transport and positive anomalies indicate the opposite. From Fig. (4.11b), offshore Ekman transport is greatest in summer due to the prominent southerly winds whereas a shift to more northerly winds in winter results in little to no offshore transport. Northerly winds in winter may even drive the onshore movement of surface waters. The spatial patterns of alongshore fronts are significantly impacted by the variability in the wind regimes (see Chapter 5). In Fig. (4.11c), negative anomalies in distance indicate fronts being found closer to the shore while positive anomalies show fronts being found further offshore. In Fig. (4.11d), there is a striking similarity between the climatology of offshore Ekman transport and the offshore position of fronts.

The bottom panel (Fig. 4.11e), shows the variability in the intensity of the poleward undercurrents along the shelf. Negative anomalies correspond to a more intense poleward flow and a positive

anomaly is indicative of a weaker poleward flow. The climatology in Fig. (4.11f) reveals the poleward flow to be more intense in summer and progressively weakens throughout winter; even becoming slightly northward by late spring (November). The strength of the nearshore poleward flow is not likely linked to the intensity of the alongshore winds but instead to wind-stress curl through the Sverdrup relation.

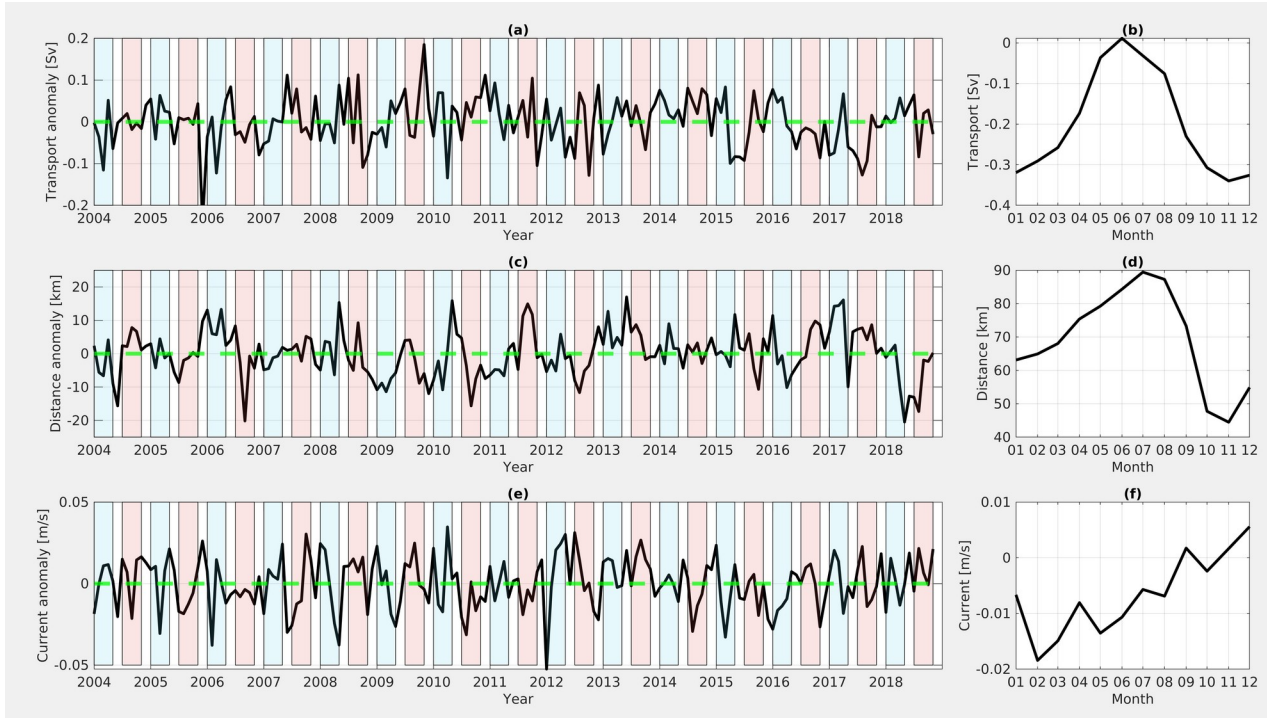


Figure 4.11: (a) Monthly anomalies computed over 2004-2018 for total offshore Ekman transport along the shelf (Sv), (c) offshore distance of fronts (km) and (e) meridional currents (m/s). Meridional current computed from transect along 32 °S. Adjacent to the respective interannual anomaly time-series are the respective climatologies (b, d and e). The alternating blue and red banding indicates the duration of the Lagrangian tracking experiments during summer and winter respectively. The dashed green line is showing the mean which is centred at 0.

Fig. (4.12) transforms the time-series of Fig. (4.11) into scatter plots to further assess the relationships between frontal spatial patterns, offshore transport and poleward undercurrents on residence times on the SBUS shelf. For each year, the summer anomalies of the respective variables in Fig. (4.10) and Fig. (4.11) are averaged over January and February; while for winter, the anomalies are averaged over July-September. The chosen periods over which the anomalies are averaged take into consideration the fact that the mean residence times for particles initiated in the upper 50 m are approximately 69 and 123 days in summer and winter respectively whereas they are 57 and 131 days for particles initiated between 100-250 m (Tab. 4.3). Consequently, the chosen anomaly periods better account for the differences in residence times between summer and winter.

From Fig. (4.12), it is evident that variability in the intensity of offshore Ekman transport and the poleward undercurrents have a substantial impact on the residence times of material along the SBUS shelf. Computing the correlations (Supp. 4.3S and 4.4S), particles released in summer at -1 m show a positive (0.51) relationship with offshore Ekman transport and none in winter whereas particles at -50 m show a smaller positive but not statistically significant correlation in summer and no significant correlation in winter. For particles released at 100-250 m, residence time is always

positive relative to the intensity of the poleward flow in both seasons, being on average 0.55 in summer and 0.68 in winter.

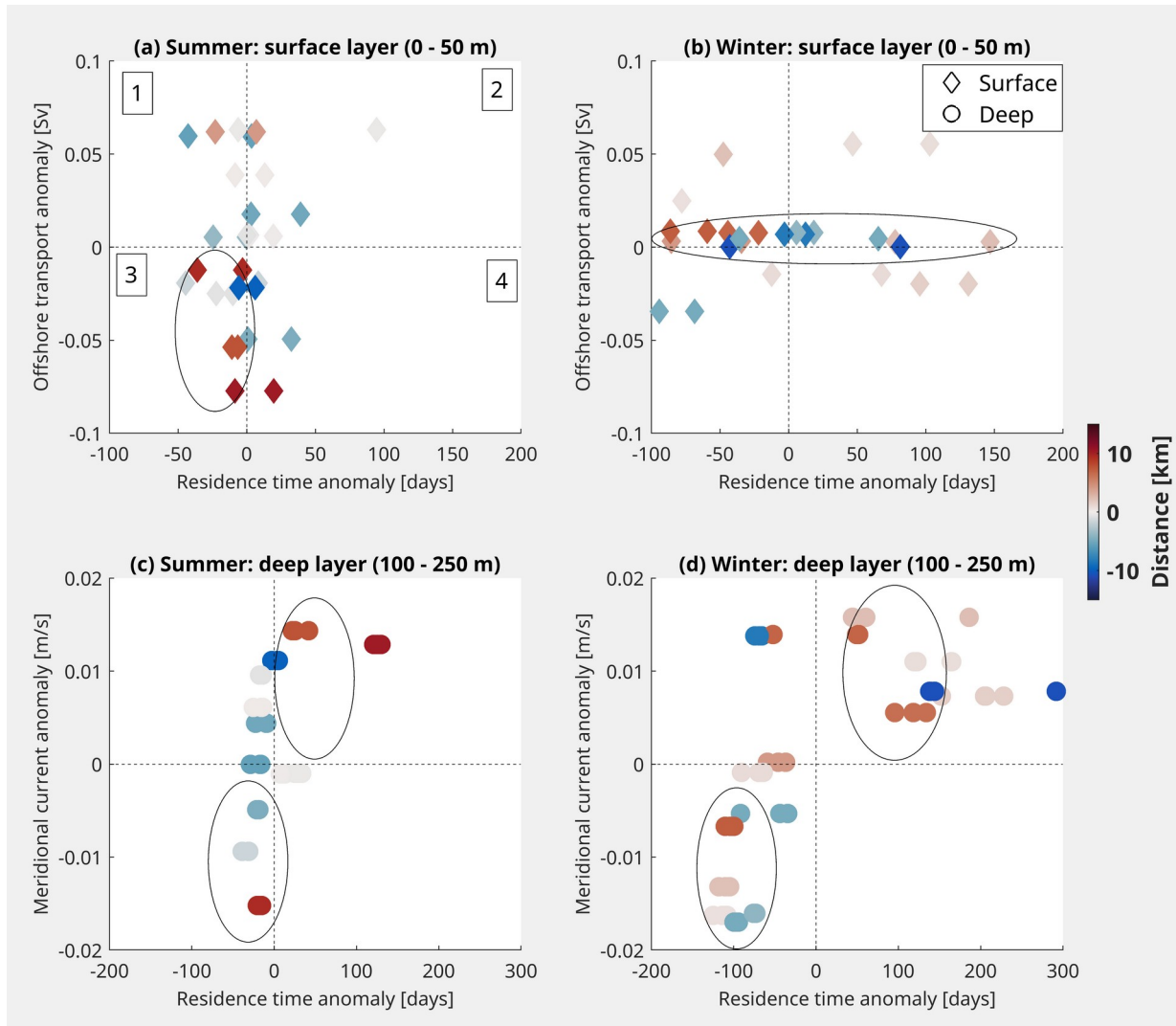


Figure 4.12: Scatter plots of retention anomalies for surface (0-50 m) and deep (100-250 m) vs the offshore Ekman transport (a, b) and meridional current anomalies (c, d). The scatter points are coloured relative to the distance anomaly of fronts and in (a) the numbers 1-4 represent a specific quadrant. Ellipses highlight regions of interest.

For the upper 50 m (Fig. 4.12a), negative retention anomalies in summer are mainly associated with periods of elevated offshore transport. This results in the alongshore fronts being found further offshore which negatively impacts their ability to retain particles. In quadrant two of Fig. (4.12a), longer residence times occur during periods of weaker offshore transport and this results in fronts being found closer to the shore which can assist in the retention of particles. Evidently, the frontal environment is influenced by the underlying transport and flow patterns along the shelf which explains why there are very few data points in quadrants one and four as this would correspond to periods where the relationship between offshore transport and residence times becomes negative. In winter (Fig. 4.12b), variability in the offshore Ekman transport is minimal owing to the weaker northerly winds. The absence of distinct offshore transport inevitably highlights the potential that fronts have as features to elevate the retention of particles along the shelf. Negative residence times are primarily associated with fronts being found further offshore which allows particles to

experience greater cross-shore displacement whereas positive anomalies in residence times occur when fronts are found closer to the coast. Though alongshore fronts are more intense in summer than winter, owing to the more defined cross-shore temperature gradients, it is interesting that fronts even in winter can elevate the residence times of particles along the shelf. Evidently, the offshore position of fronts does impact the residence times of material along the shelf. Though masked by the underlying offshore transport in summer (Fig. 4.12a), it is likely that fronts in summer are also playing an important role in elevating the residence times of floats.

For particles between -100 m and -250 m (Fig. 4.12c,d), the primary driver of variability in the residence times of particles in summer and winter is the intensity of the poleward undercurrents; however, it is interesting too that the relationship between the offshore distance of fronts relative to residence times is reversed compared to the upper 50 m (refer also to Supp. 4.3S and 4.4S). In summer (Fig. 4.12c), fronts found closer the coast are associated with lower residence times than fronts found further offshore and a similar observation can also be made during winter (Fig. 4.12d). This contradicts the narrative of fronts in closer proximity to the coast elevating residence times; although, the correlations of frontal distance vs residence times were not significant at an alpha level of 0.05 (except for -100 m). As the seed region for the Lagrangian experiments was within St Helena Bay the positive relationship between frontal distance and residence times suggests that if the SBF is further adrift it will entrap and retain more particles before they are transported poleward and around Cape Columbine. Even if the result is a statistical abnormality, Fig. (4.7) and Fig. (4.8) show that alongshore fronts can still influence the dispersion and transport dynamics of deep floats which does impact their residence times on the SBUS shelf.

4.4 Discussion

4.4.1 Lagrangian modelling in the SBUS

Using a physical ocean model configuration of the SBUS, a Lagrangian tracking code and an edge-detection algorithm the seasonal and interannual variability in residence times, transport regimes and dispersion patterns on the shelf region have been investigated. Residence times were calculated using a simple exponential decay function from float release experiments that were initiated throughout the upper 250 m of the water column during summer and winter periods from 2004-2018. In general, floats released in the upper 50 m were transported northward by the prominent surface equatorward currents and below 100 m the broad poleward undercurrents dominated the southward transport of floats. Similar to Lett *et al.* (2006) and Miller *et al.* (2006), it was found that the residence times were significantly longer in winter than summer due to the seasonal shift in the intensity of the jets and circulation features. From Tab. (4.3), the average seasonal residence times for particles initiated in the upper 50 m were 69 ± 14 days for summer and 123 ± 35 days in winter. For particles between 100-250 m, the mean residence times were 57 ± 10 and 132 ± 38 days in summer and winter respectively. Sensitivity tests revealed that floats tracked for up to 120 days or shifting the cross-shore extent of the 500 m isobath to increase the size of the considered control volume produced qualitatively similar results. Though the results are comparable and in agreement with previous works it is important to acknowledge that the magnitudes of the computed residence times can be influenced by inherent biases present in the physical ocean model (Sec. 3.3.1), the choice of Lagrangian tracking code itself and the method used to compute the residence times.

Compared to the trajectories of the GDP floats along the SBUS shelf, the model floats showed a more westward propagation in summer along the shelf edge and reduced northward transport along

the shelf in winter. In Sec. (3.3.1), the model was shown have a positive surface and subsurface temperature biases along the shelf which impacted the intensities of the surface equatorward jets and poleward undercurrents. This temperature bias resulted in the alongshore fronts being found closer to the coast and the intensity of the equatorward jets and poleward undercurrents being weaker compared to observations (Fig. 3.10). Therefore, the computed residence times herein are likely to be overestimates of the real system.

A host of community based and model specific Lagrangian tools exist (Fredj *et al.* 2016; Van Sebille *et al.* 2018) which fundamentally all solve Eq. (4.1) and thus the choice of code can be somewhat arbitrary. Some Lagrangian tools include basic stochastic processes such as random walks to resolve subgrid-scale mixing processes and others can be run online. Although ROFF has no subgrid-scale parameterisations (Capet *et al.* 2004; Carr *et al.* 2008), authors such as Mason, Colas & Pelegri (2012) have used ROFF to great success to study upwelling pathways in the Canary Upwelling System and the authors noted that the absence of diffusion processes did not affect the mean trajectory of particles provided there were enough of them. This sentiment is echoed by Van Sebille *et al.* (2018) who also mentions that stochastic processes may not be necessary if a velocity field is of a high enough resolution. The resolution of the model configuration can resolve sub-mesoscale features and thus the absence of stochastic processes within ROFF should not influence the computed Lagrangian pathways. The Lagrangian experiments were run offline (Sec. 4.2.2) for greater computational efficiency and flexibility (refer to (Van Sebille *et al.* 2018; Wagner *et al.* 2019); however, CROCO does have the capability to run Lagrangian experiments online (Debreu *et al.* 2012). Owing to the temporal scales of interest; the daily outputs from the model were more than sufficient to compute the residence times along the shelf and it is not likely that running the same experiments in an online fashion would yield significantly different results.

The concept of residence time is rather simple whereby the objective is to calculate the length of time a particle takes to move outside or beyond a specified control volume. Inherently, this broad definition allows residence times to be computed and defined using a variety of methods. In this work, an exponential decay model, similar to Piñones *et al.* (2011), was used and adopted into a piecewise function to better compute the residence times of floats on the SBUS shelf. Most naturally occurring phenomena are non-linear (Iqbal 2021) and curve-fitting an exponential function yielded R^2 values close to 1 for particles released below 100 m and 0.8-0.9 for particles above 50 m. The lower R^2 scores for the surface layer are likely due to the greater variability experienced by particles. Pfaff *et al.* (2022) computed residence times as the time taken for particles initiated at the coast to move 25 km offshore. Using Finite Size Lyapunov Exponents, which calculates the time for two fluid parcels at an initial separation distance to reach a set final distance (d'Ovidio *et al.* 2004), Rossi *et al.* (2009) computed the horizontal mixing times in different seasons for the various EBUS. Other techniques to compute residence times include using transfer probabilities of mass (Hewitt & Martin 1973) or passive age tracers (Penven *et al.* 2000).

An extension of residence times are retention times. Although the two terms can sometimes be used interchangeably, retention or retention times are predicated by a hypothesis. Authors such as Miller *et al.* (2006) considered particles to be retained if after 14 days they were within 50 km of their initial location. It is unfortunately not possible to quantitatively compare our results with others due to the differing numerical definitions of residence times used by different groups of authors. This inevitably highlights how the definition of residence time can be research specific.

4.4.2 Alongshore frontal variability and its influence on residence times

Combining the results of the Lagrangian experiments (Sec. 4.3.3) with the findings from Sec. (3.3.3), Fig. (4.13) presents an idealised cross-section of summer and winter oceanographic conditions along the SBUS shelf that impact the cross-shore dispersion and retention dynamics of particles and fluid parcels. In summer (Fig. 4.13a), southerly winds drive upwelling along the shelf generating prominent cross-shore temperature gradients. The seasonal UF is prominently defined and found close to the coast in summer whereas the intense perennial SBF is maintained by the cross-shore temperature gradients between the cool shelf waters and warmer offshore waters which are supplied by Agulhas leakage. From Chapter (3), these summer frontal features are associated with defined Lagrangian Coherent Structures (LCS) (refer to Shadden, Dabiri, & Marsden 2006), indicating that the alongshore fronts in summer dynamically divide the flow patterns along the SBUS shelf and are able to act as cohesive barriers to the offshore advection of material.

Using surface drifters deployed in summer in the Northern Benguela, Dräger-Dietel *et al.* (2018) and North, Dräger-Dietel & Griesel (2023) showed that drifters deployed near a filament will accumulate on the convergence side of the feature and experience reduced relative dispersion compared to drifters deployed in more open oceanic conditions. In addition, drifters were shown to follow the flow path of the filament; in essence being steered by the jet feature. From Sec. (4.3.2), our results are in keep with the work of Dräger-Dietel *et al.* (2018) and North *et al.* (2023) as the alongshore fronts were shown to limit the cross-shore movement of particles as well as exert a definite steering effect on particles at the surface (Fig. 4.5a). This likely contributed to the low residence times of the floats. The influence of the alongshore fronts in steering the trajectories of particles was not as pronounced at -50 m; however, in both summer and winter, evidence of subduction along the CP jet for particles deployed below 100 m gives credence to the idea that alongshore front/jet features are able to control the flow paths and advection of particles well beyond just the surface layer. Bettencourt, López & Hernández-García (2012) showed that LCS can have curtain like structures that permeate the water column for baroclinic features like mesoscale eddies. Extending this idea, the intense baroclinic nature of the CP jet is then likely responsible for limiting the cross-shore advection of deep particles as they are transported south along the shelf by the poleward undercurrents.

In winter (Fig. 4.13b), the unfavourable upwelling northerly winds accompanied by reduced stratification of the water column and a deeper MLD impact the spatial patterns of fronts along the SBUS. The UF is all but absent (Fig. 3.19); with a weak and filamentous jet feature being somewhat present between the coast and the 200 m isobath (Supp 3.4S). Further offshore, the SBF is defined by weaker SST gradients due to a homogenisation of the shelf temperatures in the absence of active upwelling which negatively impacts the intensity of the associated jet. The less intense surface equatorward jets and poleward undercurrents are the primary reason for the elevated winter residence times; however, for particles released at -50 m, there was no statistically significant difference relative to summer (Tab. 4.3). An explanation for the similar residence time is likely due to the deepening of the winter MLD as strong stratification in summer limits the extent of the equatorward jets to above 30 m while in winter, particles lie within the same mixed layer and experience similar meridional velocities at 50 m to those in summer.

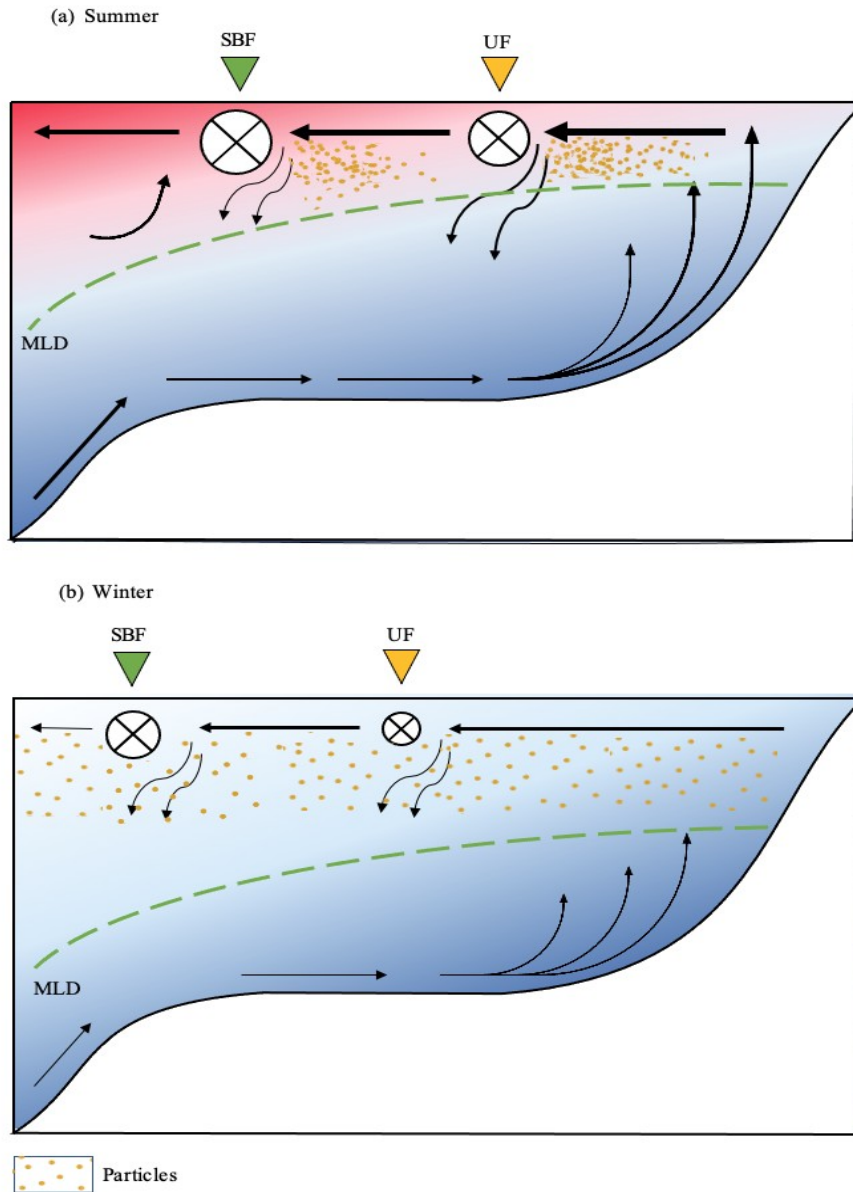


Figure 4.13: Idealised cross-section of (a) summer and (b) winter oceanographic conditions along the SBUS. Prominent cross-shore temperature gradients are present in summer whereas cross-shore temperatures are nearly homogeneous in winter. Equatorward jets (shown as circles) exist in the surface layer and their size is representative of their intensity. Black lines show the direction of transport along the shelf. Central waters upwell along the shelf edge, outcropping at the surface during summer. At the surface, horizontal arrows indicate the offshore Ekman transport, which is more intense in summer than winter. The yellow and green triangles indicate the offshore position of the upwelling (UF) and shelf-break (SBF) fronts respectively. No true UF is present in winter due to the near absence of upwelling-favourable winds. Instead, a weak equatorward jet exists between the coast and the 200 m isobath. The dotted green line represents the MLD. The winter MLD is significantly deeper than in summer. The summer frontal environment is more favourable for the accumulation and retention of particles; whereas in winter, the less cohesive alongshore fronts that are found further offshore do not act as effective barriers to the offshore advection of particles.

4.4.3 Drivers of interannual variability

Residence times show interannual variability throughout the water column (Fig. 4.10). For the upper 50 m, variability in the offshore position of the alongshore fronts as well as the intensity of offshore Ekman transport impact the residence times of particles as does variability in the intensity of the poleward undercurrents for particles initiated below 100 m. In general, negative/positive anomalies in Ekman transport correspond to fronts being found further/closer to the coast which results in lower/higher residence times for floats in the surface layer whereas positive/negative anomalies in the poleward undercurrent would expectedly be associated with longer/shorter residence times for deep initiated floats. These broad dynamical responses do hold for some years along the shelf (Fig. 4.11). The main driver of interannual variability is the alongshore wind regimes. The alongshore winds influence the shelf SST patterns by driving upwelling and this in-turn impacts the spatial patterns of fronts on the shelf. The alongshore winds also drive offshore Ekman transport and through wind-stress curl impact the intensity and flow patterns of the poleward undercurrents.

Large-scale modes of variability such as ENSO can impact the atmospheric circulations along southern Africa. In the Benguela, Rouault *et al.* (2010) showed that during summer, La Niña periods were associated with stronger upwelling-favourable winds whereas during El Niño events the southeasterly winds were significantly weaker. Relative to the SOI, a negative correlation of -0.46 ($p \approx 0.08$) was seen for residence times computed at -1 m and -0.34 for -50 m ($p \approx 0.22$) in summer. Throughout the rest of the water column in summer and for all depths in winter p -values > 0.3 such that there was no relation. Though not significant at the 95% level, the pronounced negative correlations for the upper 50 m in summer give some evidence for ENSO impacting the retention dynamics of the SBUS and if anything adds robustness to the approach used to compute the residence times because they show a correspondence to remote forcing signals. Therefore, as shown by Rouault *et al.* (2010) as well as Rouault & Tomety (2022), ENSO can influence the alongshore wind dynamics along the SBUS and in-turn this can impact the residence times of particles along the shelf. Consequently, residence times along the shelf during summer are likely to be longer during El Niño periods and lower during La Niña phases.

From Fig. (4.11), the intensity of the alongshore winds directly drives offshore Ekman transport which impacts the offshore position of the alongshore fronts. This finding is similar to that of Pitcher *et al.* (2010) who showed the offshore migration of the UF to be associated with prominent offshore Ekman transport during southerly wind events in summer. No correlation was seen between the intensity of the alongshore wind and the intensity of the poleward undercurrent along the shelf. This is somewhat expected as the poleward undercurrents are impacted by alongshore wind-stress, not intensity (Sverdrup relation). While not explored in this work, a lot of academic attention has focussed on the role of wind intensity and wind-stress curl as drivers of variability for the deep poleward undercurrents along the shelf edges in EBUS (Marchesiello, McWilliams & Shchepetkin 2003; Veitch *et al.* 2009; Vergara *et al.* 2016; Desbioles *et al.* 2016; Vélez-Belchí *et al.* 2021). Desbioles *et al.* (2016) linked the sensitivity of the deep poleward undercurrent along the SBUS shelf edge to variability in the wind-stress curl but cautioned against a Sverdrup relation. Using a physical ocean model in combination with varying spatial resolution wind products, Small *et al.* (2015) showed low resolution winds favour the development of a Sverdrup balance whereas higher resolution wind products elevate the intensity of winds close to the coast which narrows the band of intense wind-stress curl over the shelf region and weakens the Sverdrup relation. Although the Sverdrup relation may not explain all the dynamics of the poleward undercurrents on the SBUS

shelf, it is likely that the deep poleward flows are driven in some part by the alongshore wind regimes and interaction with the coastal boundary layer. Therefore, variability in the wind dynamics along the SBUS have a profound influence on both the equatorward and poleward flow patterns and this in turn impacts the spatial patterns of fronts on the SBUS shelf.

4.5 Conclusion

The flow paths, dispersion patterns and residence times of particles on the shelf region of the SBUS have been explored using a numerical model configuration of the region as well as a series of virtual Lagrangian float experiments. Particles in the upper 50 m of the water column are transported northwards along the shelf by the prominent surface equatorward jets. Once surface particles reach the broad Namaqua shelf, particles experience greater zonal dispersion on account of the offshore veering of the surface jets north of 30°S. Below 100 m, the transport of material is dominated by the deep poleward undercurrents along the shelf bottom which transports material south towards St Helena Bay as well as along the narrow shelf region between Cape Town and Cape Columbine.

Residence times of particles on the shelf are significantly longer in winter than summer throughout the water column; except at -50 m as the deepening of the winter MLD ensures similar meridional velocities to those in summer. Interannual variability in the alongshore wind dynamics impacts the intensity of the offshore Ekman transport, poleward undercurrents as well as the offshore positions of the alongshore fronts on the shelf through a modifications of the shelf SST patterns. It appears that fronts play a major role in restricting the offshore advection of material as well as rapidly transporting and steering particles along the shelf. The spatial patterns of fronts as well as their physical oceanographic traits are just an expression of the flow field spectrum that finally determines the Lagrangian drifter dispersion and residence times of particles on the shelf.

Natural interannual modes of variability linked to ENSO were found to impact the residence times of particles on the SBUS shelf. Although not statistically significant, a prominent negative correlation was found between ENSO and residence times in the upper 50 m during summer. Consequently, residence times along the shelf during summer are likely to be longer during El Niño periods and lower during La Niña phases. Another important source of variability not considered in this work is the influence of the Agulhas Current. Veitch & Penven (2017) demonstrated the influence the Agulhas Current has on the SBUS through the input of heat and salt into the offshore region. How changes in the intensity of the Agulhas Current impact residence times on the shelf region of the SBUS would be an interesting future study.

Supplementary

4.1S Integrated dispersion

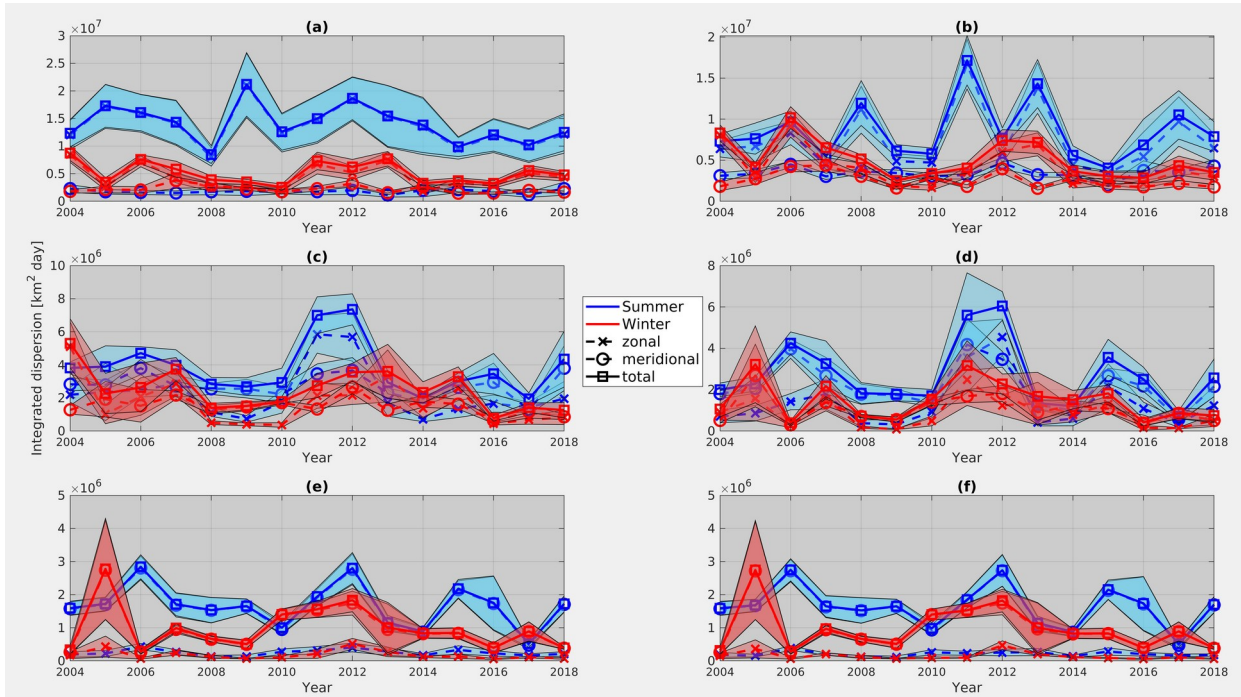


Figure 4.i: Integrated dispersion ($\text{km}^2 \cdot \text{day}$) for summer (blue) and winter (red) from the surface (a) to bottom/250 m (f) in 50 m intervals. The zonal, meridional and total dispersion curves are shown with the shading being ± 1 standard deviation.

From Fig. (4.i), the magnitude of integrated dispersion decreases with depth such that the upper 50 m of the water column are a factor of 10 times greater than the lower 200 m. Similar to Fig. (4.9), total dispersion is dominated by the zonal component for the upper 50 m and the meridional component for particles below 100 m. As expected, integrated dispersion is not constant and shows significant interannual variability throughout the water column.

4.2S Backward trajectory

The backward trajectory in Fig. (4.ii) shows the main sources of bottom waters within St Helena Bay to be a combination of shelf waters as well as SACW that upwell onto the shelf. Particles take approximately 300 days to be transported from the Namaqua shelf into St Helena Bay; significantly slower than the time taken for particles to traverse the length of the shelf by the surface equatorward jets. The interconnectedness of the bottom waters along the shelf highlight how the poleward undercurrents play an important role in the transport of biogeochemical tracers. Evidently, the poleward transport of oxygen poor/nutrient-rich waters can amplify hypoxia within St Helena Bay.

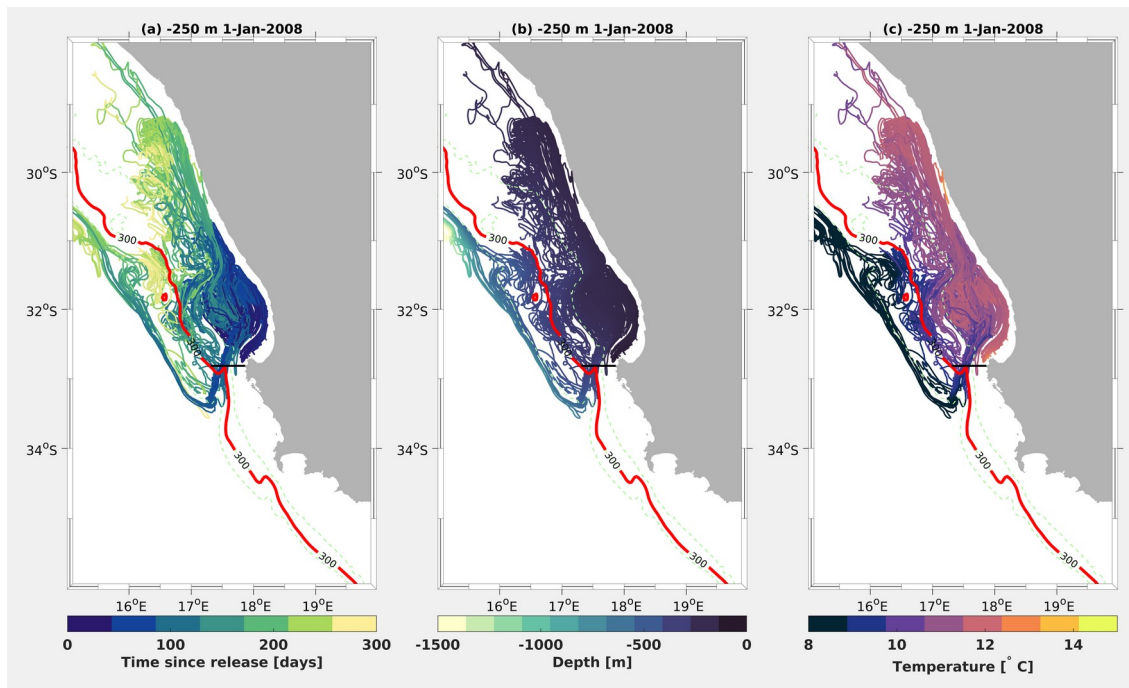


Figure 4.ii: Backwards trajectory of bottom particles tracked for 300 days. (a) is the age of the floats, (b) is the depth and (c) is the temperature.

4.3S Summer correlations

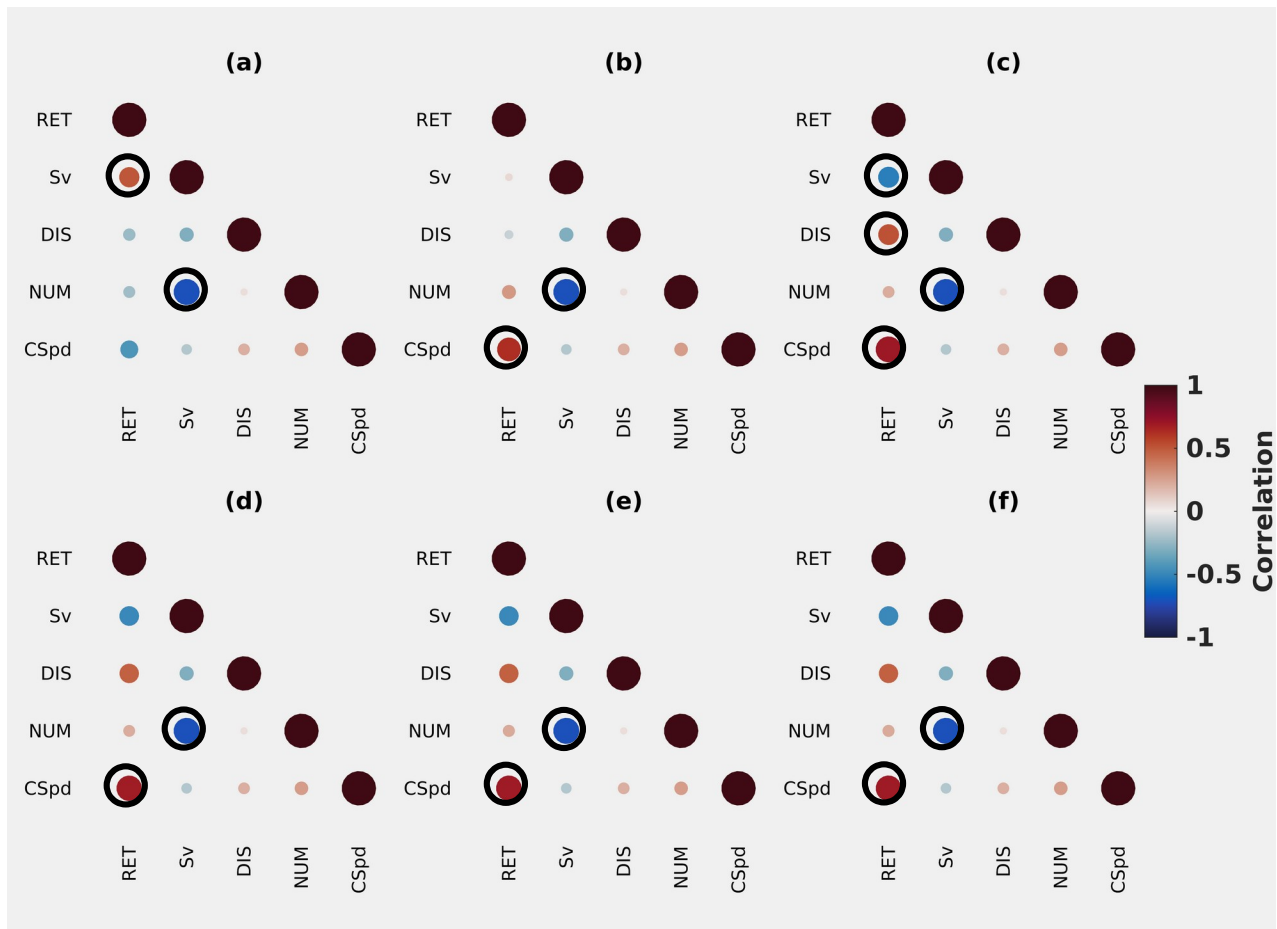


Figure 4.iii: Correlations of frontal indices with offshore Ekman transport, residence times and poleward current speed from the surface (a) to -250 m (f) in 50 m increments. Circled relations are significant at the 95 % level.

Key:

RET: residence time [days]

Sv: offshore Ekman transport [Sv]

DIS: distance of fronts offshore [km]

NUM: number of detected fronts

Cspd: Intensity of meridional component of the poleward undercurrent [m/s]

4.4S Winter correlations

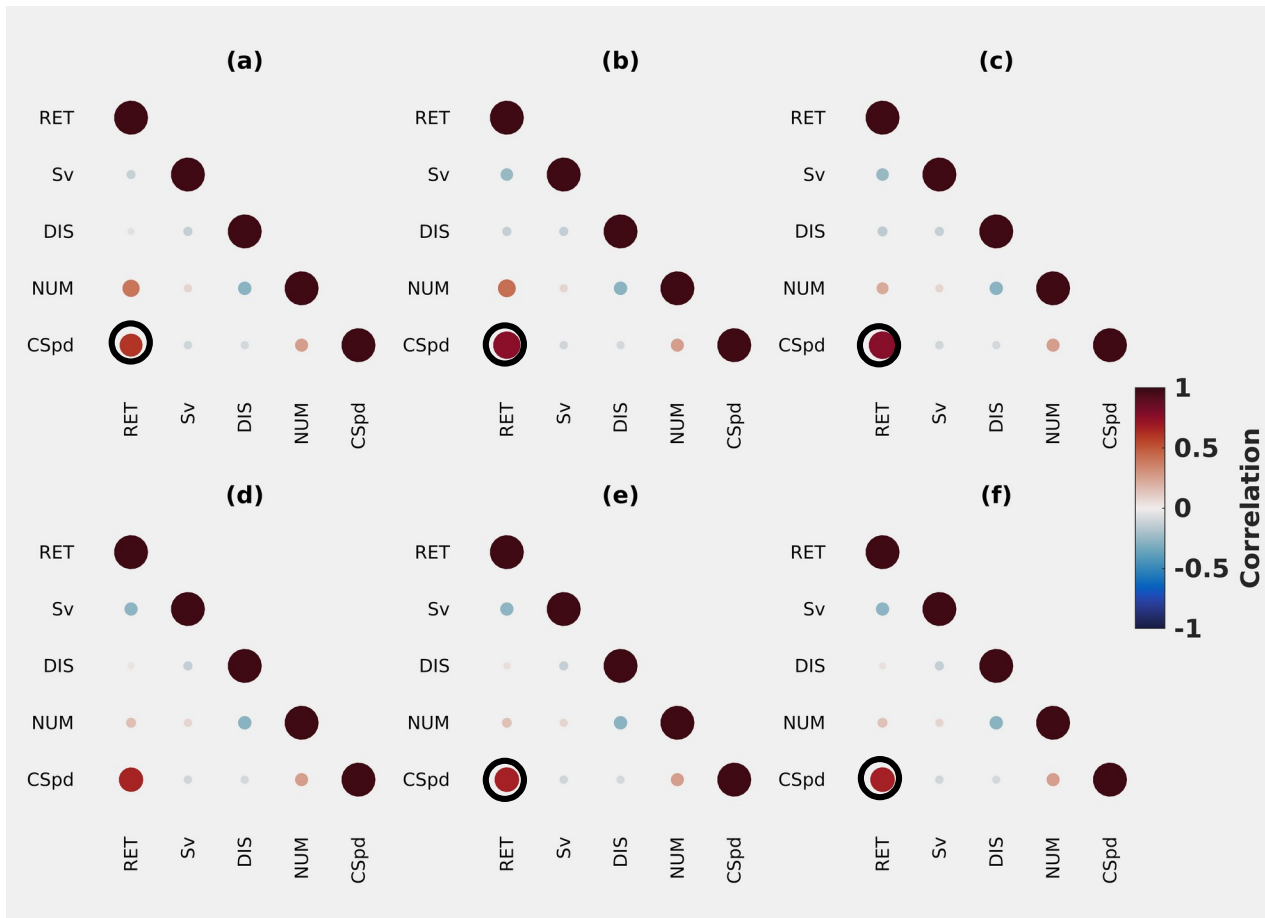


Figure 4.iv: Correlations of frontal indices with offshore Ekman transport, residence times and poleward current speed from the surface (a) to -250 m (f) in 50 m increments. Circled relations are significant at the 95 % level.

Key:

RET: residence time [days]

Sv: offshore Ekman transport [Sv]

DIS: distance of fronts offshore [km]

NUM: number of detected fronts

Cspd: Intensity of meridional component of the poleward undercurrent [m/s]

Chapter 5: The role of variable alongshore summer wind profiles on nearshore retention dynamics along the Southern Benguela Upwelling System

“But my dreams they aren’t as empty as my conscience seems to be. I have hours, only lonely.”
Limp Bizkit – Behind Blue Eyes

Summary

The phenology of alongshore upwelling-favourable winds near coastal margins play a major role in driving variability in the nearshore current profiles, offshore positions of the alongshore fronts as well as significantly influencing the productivity of coastal systems through modulating the fluxes of nutrients. Within Eastern Boundary Upwelling Systems (EBUS); alongshore, equatorward winds drive offshore Ekman transport; resulting in the upwelling of cold, nutrient-rich, deep waters into the euphotic zone, sustaining high levels of biological productivity. Using a physical ocean model in conjunction with a Lagrangian tracking code, the influence of variable summer alongshore wind profiles on the oceanography and retention dynamics within the nearshore environment of the Southern Benguela Upwelling System (SBUS) are investigated. Two types of summer, seven-day, southerly wind profiles are studied: parabolic and sustained events. The two distinct alongshore wind profiles generate unique responses in the nearshore current velocities as well as the offshore positions of the upwelling front. This generates significantly different retention regimes along the coast. During parabolic wind-events, nearshore surface currents show an oscillation between a net poleward and equatorward flow in response to the wind phasing while under sustained wind-events, net transport is always equatorward. Lagrangian float experiments reveal cross-shore and alongshore displacement of particles along the SBUS coast to be a factor of two times greater under sustained events relative to parabolic. Furthermore, the development of a poleward current following wind relaxation is more prominent for the shelf region south of 30°S as well as where the shelf-width is narrow. As nearshore retention is greater during parabolic wind-events, harmful algal blooms are likely to be more frequent and the presence of a poleward flow following wind relaxation will act to transport blooms south and into St Helena Bay which may enhance hypoxia in the region. Future projections of change wanting to investigate near coastal responses to variability in wind-forcing should not only consider variability in the intensity of the alongshore winds but also whether shifts will occur in their synoptic cycles.

5.1 Introduction

The dynamics of alongshore winds near coastal margins play a major role in driving variability in the nearshore current profiles (Gan & Allen 2002a, b; Gan & Allen 2005; Fawcett, Pitcher & Shillington 2008), cross-shore hydrography (Chao 1987; Send, Beardsley & Winant 1987; Austin & Lentz 2002; Roughan *et al.* 2006) as well as significantly influencing the productivity of coastal systems through modulating the fluxes of nutrients (Albert *et al.* 2010; Pitcher *et al.* 2014). Within Eastern Boundary Upwelling Systems (EBUS); alongshore, equatorward winds drive offshore Ekman transport, resulting in the upwelling of cold, nutrient-rich, deep waters into the euphotic zone; sustaining high levels of biological productivity. Focus herein is given to the Southern Benguela Upwelling System (SBUS) and it is necessary to contextualize how spatial and temporal scales of variability in the alongshore winds influence the oceanography of the region.

The Benguela Upwelling System (BUS) is one of the four major EBUS in the global ocean. The BUS is typically delineated into two distinct regions, divided by the Lüderitz upwelling cell at 27°S (Hutchings *et al.* 2009; Veitch, Penven & Shillington 2009). The northern Benguela experiences perennial upwelling whereas upwelling in the SBUS is highly variable and seasonal; linked to the position of the South Atlantic Anticyclone (SAA) (Fennel 1999; Shillington *et al.* 2006; Lamont *et al.* 2018). The SAA gives rise to upwelling-favourable south-easterly winds during austral summer (Dec, Jan, Feb) while a seasonal northward shift in the position of the atmospheric circulation in winter (Jun, Jul, Aug) results in northerly winds dominating the shelf region south of 30°S (Hutchings *et al.* 2009). Large-scale modes of variability such as El Niño-Southern Oscillation (ENSO) impact the meridional position of the SAA, resulting in stronger/weaker upwelling-favourable southerly winds in summer during La Niña/ El Niña years (Rouault, Pohl & Penven 2010; Rouault & Tomety 2022). Using 43 years of ERA5 reanalysis, Bordbar, Mohrholz & Schmidt (2023) attributed variability in the upwelling regime along the Benguela to variability in the dynamics of the SAA but cautioned against it being the sole driver. Superimposed on the large-scale seasonal changes in the alongshore winds is considerable mesoscale and synoptic scale variability which can be driven by variable coastal topography, nearshore Sea Surface Temperature (SST) patterns and the influence of land-sea diurnal winds (Shillington *et al.* 2006; Bordbar *et al.* 2023).

Upwelling intensity along the SBUS's coast is not contiguous and instead limited to a number of enhanced upwelling cells at the Cape Peninsula (34°S), Cape Columbine (33°S) and Namaqua (30°S) (Veitch *et al.* 2009). These upwelling cells arise due to the nature of the alongshore winds as well as the variability of the coastal geometry (Burls & Reason 2008; Veitch, Penven & Shillington 2010; Strub *et al.* 2013). A numerical modelling study by Veitch *et al.* (2010) demonstrated how variations in coastline topography have the effect of enhancing upwelling in the lee of prominent capes. Another study conducted by Penven *et al.* (2000) showed how the Cape Columbine headland induces a cyclonic wind-stress curl, resulting in the formation of a prominent upwelling plume off the cape in summer and the development of a perennial cyclonic eddy within St Helena Bay. Modelling work shows the size of the cyclonic circulation within the bay to be related to the intensity of the wind (Penven *et al.* 2000) and the feature plays a major role in enhancing the coastal retention of matter. Elevated retention results in St Helena Bay being highly productive but also prone to seasonally recurrent hypoxia driven by the nearshore decay of phytoplankton blooms (Pitcher *et al.* 2014; Burger *et al.* 2020).

Coastal topography plays a major role in shaping the mesoscale structures of the alongshore winds which in-turn drive upwelling along the coast. Low-level winds tend to be weaker nearer to the

coast as a result of friction with the land surface. This drop-off in wind intensity is associated with a cyclonic wind-stress curl which favours Ekman pumping and in combination with nearshore divergence results in cool SSTs along the coast (Capet, Marchesiello & McWilliams 2004; Astudillo *et al.* 2019). In the Benguela, modelling work by Burls & Reason (2008) showed cool SSTs result in a downward sensible heat flux that shallows the atmospheric inversion layer; making coastal winds more susceptible to topographical steering. Similar results were found by Boé *et al.* (2011) in the California Upwelling System; however, they stressed that coastal topography is likely to play a greater role in influencing the mesoscale wind variability than SST. Nevertheless, an ocean-atmosphere feedback exists between topographically controlled upwelling-favourable winds and the downward sensible heat flux driven by the resultant cool SSTs (Buris & Reason 2008).

In summer, the phasing of upwelling-favourable and relaxation/reversals in the wind operate at both the synoptic and inertial scales (Davies & Xing 2003; Fawcett *et al.* 2008). Southerly winds generate an equatorward surface current whereas during periods of relaxation/reversals, a nearshore counter-current can develop. A mechanism for this phenomenon was numerically demonstrated by Gan & Allen (2002a, b) who attributed the poleward flow to the development of a negative pressure gradient that is set-up by the interaction of the wind-induced coastal current with alongshore variations in the coastline and bottom topography. Observational work by Fawcett *et al.* (2008) along the southern Namaqua shelf showed sub-inertial currents reversed at periods of 2-5 days in response to the synoptic winds. Consequently, periods of enhanced relaxation can elevate the coastal retention of matter due to the juxtaposition of a nearshore poleward current and an offshore equatorward flow which effectively generates a dynamic barrier. Similar observational dynamics on the role of wind reversals in enhancing biological productivity through elevated retention have also been documented by Roughan *et al.* (2006) in the California Upwelling System. Following a favourable upwelling event, Pitcher & Nelson (2006) suggested that periods of prolonged wind relaxation thereafter may enhance Harmful Algal Blooms (HABs) within St Helena Bay as a consequence of the southward transport of warm, stratified, bloom filled waters along the shelf into the bay itself.

Although transport in the SBUS is dominated by sub-inertial scale currents, inertial oscillations are superimposed on the mean northward or southward flow (Fawcett *et al.* 2008; Lucas *et al.* 2014; Fearon *et al.* 2022). Along coastal margins, the land-sea breeze is a ubiquitous feature and has a diurnal cycle. Owing to the latitudinal position of the SBUS, this diurnal cycle is in phase with the inertial frequency such that there is a resonance between the diurnal land-sea breeze and inertial oceanic currents (Fearon *et al.* 2022). This resonance results in a similar 180° phase shift in the nearshore currents as well as enhanced diapycnal mixing close to the coast (Davis & Xing 2003; Lucas *et al.* 2014). Modelling work by Fearon *et al.* (2022) suggests that diurnal-inertial resonance along the SBUS may reduce the offshore advection of matter from sub-inertial flows, leading to elevated retention of nutrients along the coastal band. Furthermore, enhanced mixing at the sub-inertial and inertial scale can not only impact phytoplankton distribution in the water column (Lucas *et al.* 2014) but also community composition (Fawcett *et al.* 2007).

The spatial variability of upwelling-favourable alongshore winds is an important avenue of research owing to their role in driving coastal upwelling and impacting the near coastal SST patterns. From Capet *et al.* (2004), the difficulty for numerical models in resolving the nearshore wind dynamics stems from a combination of model spatial resolution as well as the resolution of the atmospheric forcing products themselves. Consequently, a number of numerical modelling studies have focussed on evaluating and contrasting various atmospheric reanalysis products of differing spatial

resolutions in an attempt to better resolve nearshore oceanographic environments (Blanke *et al.* 2005; Desbiolles *et al.* 2016; Bonino *et al.* 2019; Taboada *et al.* 2019). In addition to numerical modelling studies, observational works such as Pitcher *et al.* (2010) in the SBUS and Roughan *et al.* (2006) in the California Upwelling System have explored biogeochemical and hydrographic variability relating to the phases of upwelling-favourable and reversals in the alongshore winds.

Observational works such as Fawcett *et al.* (2008) are invaluable for describing the nearshore current dynamics and their responsiveness to the synoptic alongshore wind regimes. However, observational studies are limited in their temporal and spatial coverage. Consequently, a large ensemble of various wind-events and their spatial as well as temporal impacts on the alongshore current dynamics for the whole of the SBUS shelf is desirable. Therefore, this study aims to understand how the phasing of the synoptic alongshore upwelling-favourable winds influence the oceanography and retention dynamics along the SBUS coast. The objective is to use a full physical ocean model configuration of the SBUS in conjunction with a Lagrangian tracking code to assess the impact of varying summer wind profiles on the alongshore fronts, nearshore current profiles and Lagrangian transport dynamics. The methodology section details the numerical model configuration, wind-event identification as well as the Lagrangian code and computed diagnostics. The results are divided into two parts: (1) oceanographic response and (2) Lagrangian dynamics. The first section describes the responses of the nearshore currents and alongshore fronts to the varying wind profiles while the second details the displacement dynamics of virtual Lagrangian floats along the entire shelf region north of Cape Columbine. The discussion then explores the role of different wind profiles on the oceanography of the SBUS as well as commenting on the importance of wind characteristics in the context of projections of future changes relating to the SBUS.

5.2 Methodology

5.2.1 Model configuration

The physical ocean model CROCO (Coastal and Regional Ocean Community model) based on ROMS (Shchepetkin & McWilliams 2005) was used to study the oceanographic environment of the SBUS. Built upon ROMS_AGRIF (Debreu *et al.* 2012), CROCO is a free-surface, terrain-following coordinate model that is well suited for regional applications and it solves the incompressible primitive equations based on the Boussinesq and hydrostatic approximations. The code used is that of CROCO v1.1 version available at <http://www.croco-ocean.org>. The model configuration is that from Chapter (3) and is detailed in Sec. (3.2.1). In summary the model was configured as a single domain with a horizontal spatial resolution of 3 km and consisted of 60 vertical sigma levels. The oceanic boundary conditions were provided by the 1/12° GLORYS physical reanalysis product and the atmospheric forcing utilised the ERA5 3-hourly product which is the fifth generation of the European Centre for Medium-Range Weather Forecasts (ECMWF) reanalysis and has a spatial resolution of 0.25° x 0.25° for atmospheric variables. The K-profile parameterisation was used for the vertical mixing regime. Model outputs were saved daily and studied from January 2004 – December 2018. As done in Chapter (3) and Chapter (4), a zoomed portion of the full model domain (Fig. 5.1a) is used to better study and visualize the shelf and nearshore regions of the SBUS.

Evaluated against a combination of *in situ* and satellite data, the model was able to resolve the low-frequency interannual and intra-seasonal variability in SST and surface circulation features in the

SBUS (Sec. 3.3.1). An artefact of the ocean boundary physical forcing conditions, the model resolved a very deep winter mixed layer depth (MLD) and had warm temperature biases along the shelf region which impacted the position and intensity of the alongshore jets (Sec. 3.3.2). In summer, cool SST biases along the coast were localised to known regions of intense upwelling and cross-shore SST gradients such as within St Helena Bay, Cape Columbine and Cape Point (Fig. 3.5). Relative to two shallow water moorings, located north of Cape Columbine at 20 and 70 m depth respectively (refer to Pitcher *et al.* 2014), the model did not capture the same degree of variability as seen in the observations (Fig. 3.11). This was likely due to the low spatial resolution of the forcing winds along the coast which were not sufficient to capture the high degree of variability associated with the nearshore region such as diapycnal processes as well the effects of land-sea breezes (Fearon *et al.* 2020, 2022).

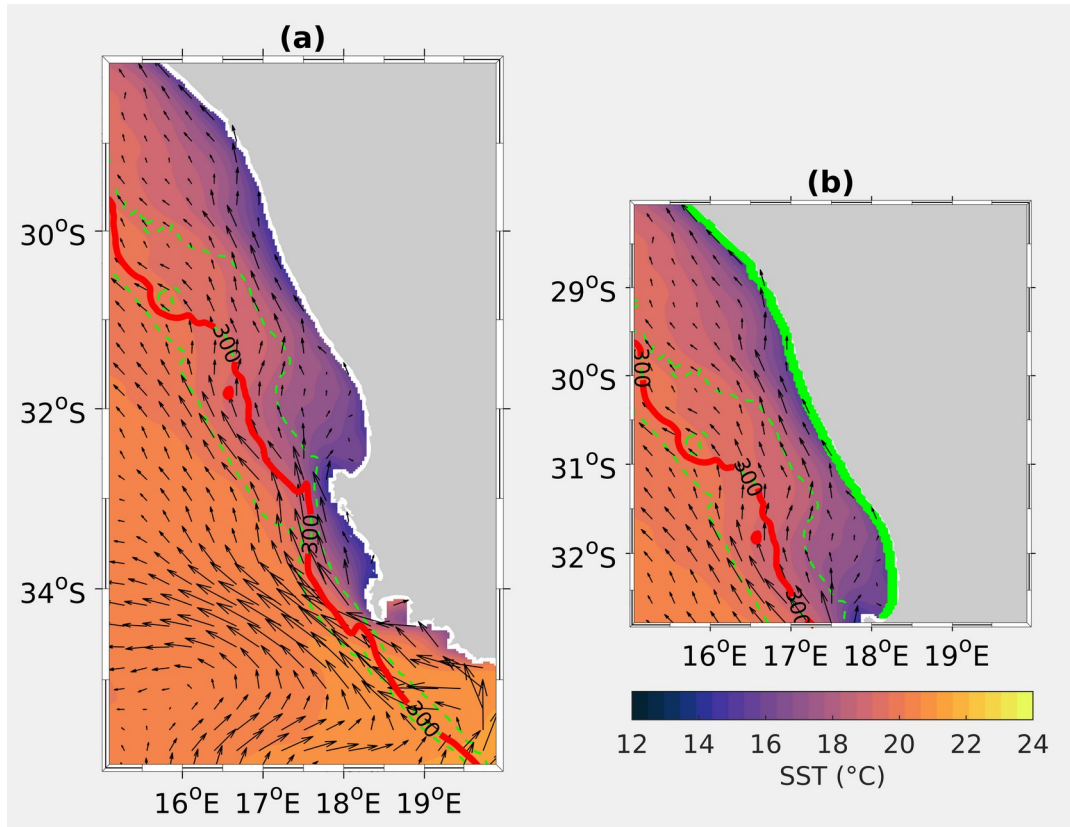


Figure 5.1: Mean summer (Dec, Jan, Feb) SST computed over 2004-2018. Black arrows show the surface geostrophic currents. The green dotted lines correspond to the 200 and 500 m isobath and the red line indicates the 300 m isobath. Panel (a) shows the zoomed domain of the model configuration. In (b), the green band along the coast corresponds to the release locations of virtual Lagrangian floats.

Although works such as Astudillo *et al.* (2019) have used coupled ocean-atmosphere configurations to investigate nearshore oceanic dynamics, using only atmospheric reanalysis as forcing still allows the atmospheric and oceanic responses along coastal margins to be resolved and studied (Bonino *et al.* 2019). García-Reyes *et al.* (2018) used NCEP2 reanalysis to investigate modes of upwelling-favourable wind variability along the Benguela while Desbiolles *et al.* (2016) used ECMWF products that form a part of the ERA5 reanalysis (Hersbach *et al.* 2020) to fill the coastal winds not captured by the 0.25° QuickSCAT observations in order to account for wind drop-off close to the coast along the SBUS. A study by Strub & James (2022) showed elevated wind-stress along coastal

margins within EBUS to not stem from too high winds speeds or incorrect model physics within the ERA5 product and instead was related to the interpolation of the wind-stress fields to rectilinear grids. The implications of regridding and interpolation of ERA5 within CROCO were not explored; however, the model configuration and use of ERA5 forcing was deemed sufficient to study the nearshore dynamics of the SBUS in response to variable alongshore wind profiles at the synoptic scale.

5.2.2 Identification of wind-events

The seasonal cycle of alongshore winds within the SBUS sees upwelling-favourable southerly winds being most intense during the summer months while in winter there is a shift in the wind regime to weak northerly winds (Fig. 5.2a). Upwelling is also not continuous throughout the summer season and instead limited to episodic events that operate on a synoptic cycle of 2-8 days. Furthermore, upwelling intensity is not contiguous along the shelf region; instead being more intense at discrete locations along the coast. Therefore, these temporal and spatial dynamics in the alongshore wind field required consideration when identifying desired wind-events.

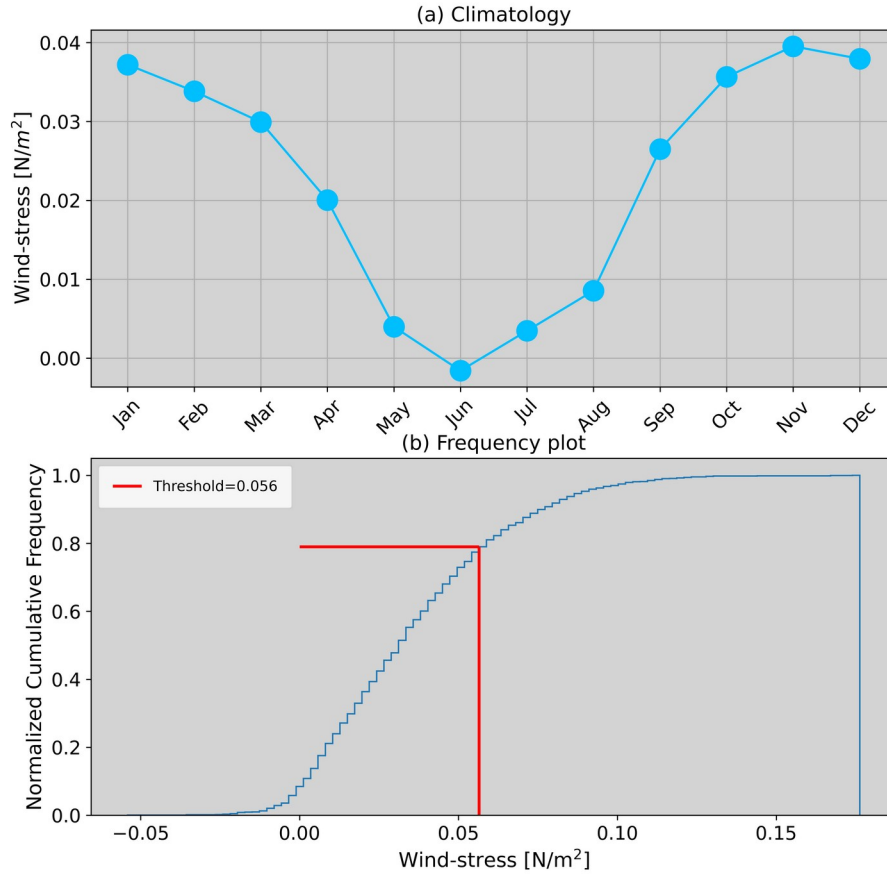


Figure 5.2: (a) Monthly climatology of alongshore winds spatially averaged for the shelf region north of Cape Columbine. (b) Normalized cumulative frequency of spatially averaged wind during summer (Dec, Jan, Feb). Red line corresponds to the 80th percentile of the data, having a value of 0.056 N/m².

A first step was to define a threshold of detection in order to identify prominent and statistically intense summer wind-events. The model wind-stress field was rotated to be parallel to the shoreline

and then averaged across 4 grid-cells (approximately 10 km) from the coast for the shelf region extending north of Cape Columbine. Thereafter, a spatial average was taken and a global histogram of the mean alongshore winds constructed (Fig. 5.2b). The dominance of southerly winds in Fig. (5.2b) is rather stark as northerly winds only account for < 10% of events. For simplicity, the 80th percentile was used to delineate strong southerly wind-events and this corresponded to a threshold value of 0.056 N/m².

To explore the role of wind relaxation/reversals contributing to retention along the shelf, two seven-day wind profiles were investigated (1) a parabolic and (2) sustained profile (Fig. 5.3). The parabolic profile exhibited an intensification of a wind-event over the first two days, with the subsequent three consecutive days being more intense and above the threshold value, followed by a defined relaxation phase. The second wind profile considered sustained southerly winds between 5-7 days in duration that were above the threshold value and during which there were no consecutive days below the threshold intensity. The objective was to investigate the nearshore oceanic responses to these differing wind profiles; with a focus on how they influenced the nearshore current profiles, frontal positions and transport dynamics along the coast.

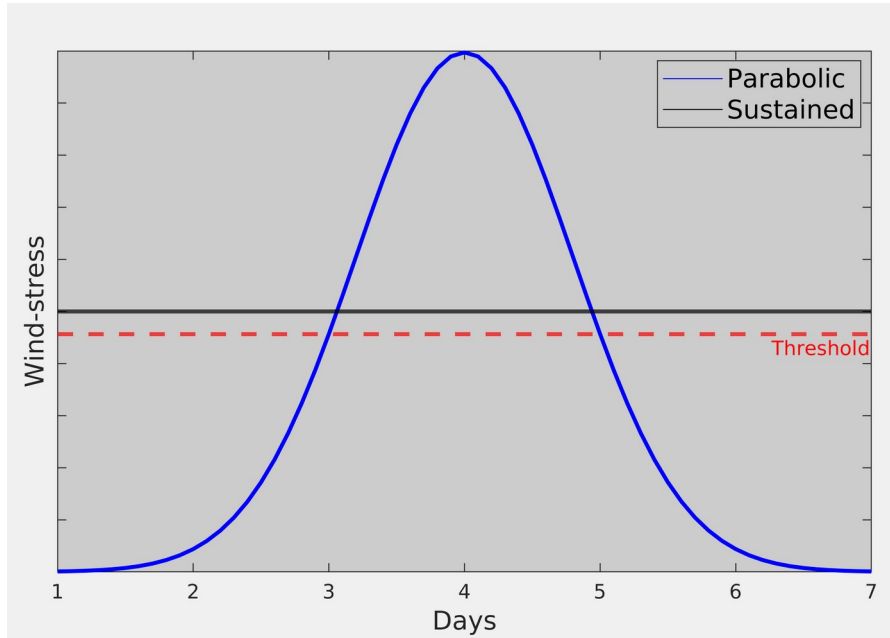


Figure 5.3: Idealised wind profiles for investigation. Blue line corresponds to parabolic wind-events while the black is for sustained. The red-dotted line shows the detection threshold used for the wind-event identification.

Using the spatially averaged winds for the shelf region north of Cape Columbine, the data was converted to a boolean array (a) such that a value of 1 corresponded to a day where the wind on average across the entire spatial domain was greater than the threshold value while a 0 was assigned to all days below the threshold. Thereafter, a secondary boolean sequence (b) corresponding to one of the desired wind profiles was convoluted over the temporal array (Eq. 5.1).

$$(a*b)_n = \sum_{k=-\infty}^{n=\infty} a_k b_{n-k} \quad (5.1)$$

A one-dimensional linear convolution was used to identify the indices matching the desired wind-event sequences. A small post-processing was thereafter applied to account for duplicate indices to ensure that a single event was uniquely identified.

Between January 2004 and December 2018; 19 seven-day parabolic and 5 sustained summer wind-events were identified. The contrast in the number of events detected between the two wind profiles stems from the fact that the parabolic wind-event is more in phase with the synoptic cycle observed in the SBUS. Sensitivity tests revealed that adjusting the threshold value for detection or changing the number of cross-shore grid-cells over which the coastal winds were averaged impacted the number of events detected for both wind profiles. Preliminary work also attempted to sub-divide the shelf-region into smaller spatial domains over which to identify wind-events. Doing so achieved the desired goal but added extra complexities due to duplicate indices as well as some events being region specific. This would inherently require each sub-region to be studied separately and was consequently deemed beyond the scope of this work. In lieu of sub-dividing the shelf region, the simpler option of just spatially averaging across an entire domain was adopted; however, a future study could use refined spatial domains to better study the intra-regional differences arising from variable wind profiles.

5.2.3 Lagrangian code and diagnostics

The ROMS offline tool (ROFF) was used (Capet *et al.* 2004; Carr *et al.* 2008) to seed and track virtual Lagrangian floats in order to investigate the alongshore transport dynamics of matter along the coast during parabolic and sustained wind-events. ROFF uses a fourth-order Adams-Bashford-Moulton predictor corrector scheme to integrate Eq. (5.2)

$$\frac{d\mathbf{x}}{dt} = \mathbf{v}(\mathbf{x}, t) \quad (5.2)$$

over time given an initial condition $\mathbf{x}(t_0) = \mathbf{x}_0$ and 3D daily velocity fields from CROCO. ROFF is a Lagrangian model specifically designed for ROMS and does not include any subgrid-scale parameterisations. However, the absence of diffusion processes should not significantly influence particle trajectories as the model configuration is of a high enough spatial resolution. For the entire nearshore region north of Cape Columbine (Fig. 5.1b), 3546 floats were initiated at several discrete depth levels [-1, -5, -10, -20, -30, -50 m] for the 2nd to the 4th model grid-cells from the coast. The initialisation corresponded to an offshore distance of 3-10 km approximately, which was similar somewhat to the Lagrangian experiments undertaken by Pfaff *et al.* (2022) who also explored nearshore retention dynamics along the SBUS. The first grid-cell was not used to avoid spurious interactions with the coastal boundary layer on initialisation. Float experiments were run for a period of seven-days, corresponding to the duration of the wind-events. An integration step of 600 seconds was used and outputs were saved every 6-hours. A total of 10 Lagrangian float experiments were run; all 5 of the sustained wind-events and 5 parabolic events, which were preferentially chosen for having prominently defined relaxation phases in addition to obvious poleward flows.

For the study, two main diagnostics were computed: displacement of Lagrangian floats along the coast and the offshore position and gradient of alongshore fronts located between the coast and 200 m isobath. To investigate the transport dynamics in the nearshore, a simple displacement metric was designed such that the initial ($\mathbf{x}(t_{ini})$) and final ($\mathbf{x}(t_{end})$) positions' of the floats after seven-days were used to compute the cross-shore and alongshore displacements as well as the total

distance travelled by the floats ($\Delta x = x(t_{end}) - x(t_{ini})$). As done in Chapters (3) and (4), frontal features on the SBUS shelf were identified from the daily SST fields using the gradient-based Canny algorithm (Canny 1986). The Canny algorithm, in short, uses a two-dimensional Gaussian filter to create a gradient map. Thereafter, a process of hysteresis is applied using a lower and upper defined threshold value to identify significant edges in the gradient map which are then flagged as frontal edges. A full description of the implementation is available in Sec. (3.2.3). Once the frontal edges were identified, the mean offshore position of the detected fronts parallel to the shoreline (Eq. 3.5) as well as the SST gradients of the features were computed.

5.3 Results

5.3.1 Oceanographic response to variable upwelling winds

Using a physical ocean model configuration of the SBUS, the oceanic responses in the nearshore region are interrogated for a series of sustained and parabolic summer wind profiles (Fig. 5.3). The section begins by exploring the alongshore current dynamics in response to the phasing of the upwelling-favourable winds (Fig. 5.4, Supp. 5.1S) before detailing the alongshore frontal responses (Fig. 5.5, Fig. 5.6). Thereafter, an example of a parabolic (Fig. 5.7) and sustained (Fig. 5.8) wind profile are showcased in order to highlight the differing upwelling regimes and frontal patterns that arise from variability in the respective alongshore synoptic winds.

The velocity of an alongshore wind profile plays a significant role in shaping the intensity and direction of nearshore currents. For parabolic wind profiles (Fig. 5.4), the phasing of upwelling and relaxation/reversal in the alongshore winds has a pronounced effect on the transport dynamics in the nearshore region along the SBUS. An expected and clear pattern emerges in Fig. (5.4a), where there is an oscillation between an equatorward and poleward flow in response to the phasing of the alongshore winds (Fig. 5.4b); however, there is significant inter-event variability relating to the intensities of the current velocities along the coast. A poleward current is usually present and near contiguous along the coast south of 30°S during the initial two days over which the winds are weak, but intensifying. The perennial nature of southerly winds north of 30°S likely limits the formation of a poleward flow, even during weak southerly wind periods and relaxation phases. An obvious shift to an equatorward flow occurs across the entire coastal domain as the winds intensify to above the threshold value. During wind relaxation, a poleward flow in part does develop along the shelf; however, it is spatially variable and more pronounced off of Cape Point and Cape Columbine. The narrowing of the shelf-width near prominent capes as well as being centres for upwelling generate intense alongshore fronts which are defined by prominent cross-shore SST gradients. These intense fronts strengthen the equatorward currents off of the capes.

For sustained wind profiles (Supp. 5.1S), the uniformity in wind intensity generates a near continuous equatorward flow along the entire shelf that persists for the duration of the wind-event. Similar to the parabolic profiles, the most intense nearshore current velocities are localized off of the prominent capes. Although sustained southerly winds will generate intense upwelling along the coast (refer to Fig. 5.8), the absence of a defined relaxation phase precludes the formation of a nearshore counter-current which will consequently limit the nearshore retention of recently upwelled nutrients. Instead, particles will be quickly transported north along the shelf and away from the coastal domain. A similar dynamic is likely to be present for the coastal region north of 30°S, even during parabolic wind-events. Therefore, it is evident that different wind profiles bring about unique responses in the nearshore current velocities.

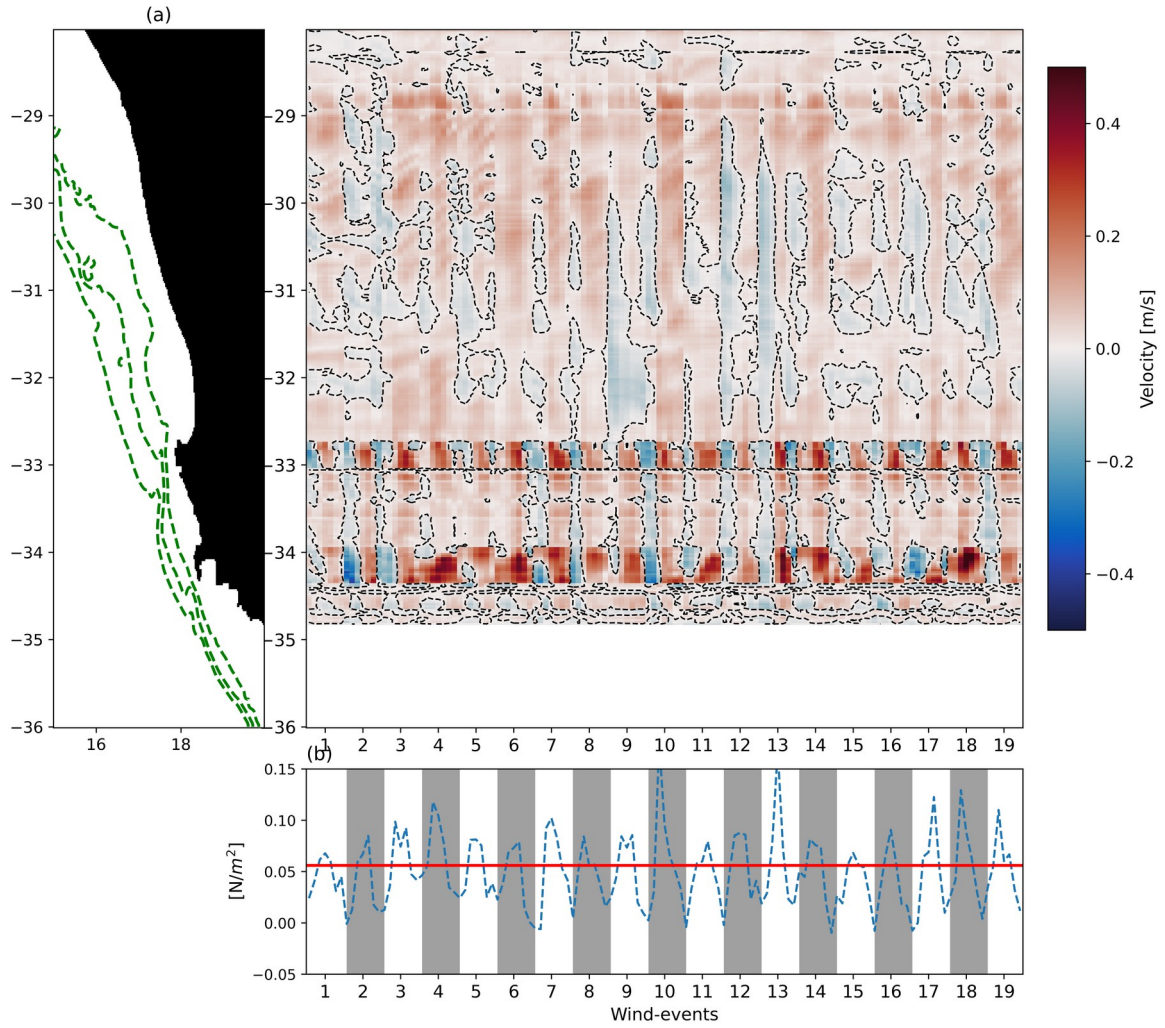


Figure 5.4: (a) Hovmöller plot of the vertically averaged alongshore current between the surface and 50 m for the various discrete parabolic summer wind-events. The green-dotted lines show the 200, 300 and 500 m isobaths. Important to note that the events are not consecutive. (b) The various alongshore wind profile, with the red-line denoting the threshold value. The grey and white banding in (b) just assists in readability.

The phasing of coastal winds not only affects the alongshore current velocities (Fig. 5.4, Supp. 5.1S) but also impacts the offshore position of alongshore fronts (Fig. 5.5, Fig. 5.6). Southerly winds stimulate upwelling along the coast which results in the formation of an Upwelling Front (UF) developing at the interface of recently upwelled and cold deep waters and warmer surface waters. As upwelling persists, these frontal features can be displaced offshore by Ekman transport and thus their spatial evolution is of interest under varying wind profiles.

In Fig. (5.5), the time-evolution of the SST gradient (Fig. 5.5a) and offshore position (Fig. 5.5b) of the UF is shown for the 19 identified parabolic wind-events. From Fig. (5.5), the large interquartile ranges (IQR) for SST gradient and offshore position during the first two days is indicative of the high degree of variability present in the initial frontal states, possibly due to the presence of small filaments. As the winds begin to intensify, the frontal gradients (Fig. 5.5a) rise sharply in response to the formation of a defined UF which is exemplified in the narrowing of the IQRs from day three and onwards. In Fig. (5.5b), a minimum in offshore distance by day four is likely associated with

the formation of a defined UF close to the coast and over the course of the parabolic wind-events, the UF shows an offshore displacement in response to Ekman transport. Given the large IQR, the change in offshore position of fronts close to the coast does not seem significant; however, they are intensified by the onset of strong upwelling-favourable winds. More intense fronts will restrict the offshore advection of material. Furthermore, it is interesting to note that the onset of relaxation does not result in a collapse or retreat in the UF. Instead, the feature still remains defined by strong SST gradients and continues to move slightly offshore. The UF is likely to persist for several days after the winds subside but as the feature moves further from the coast, mesoscale mixing will erode the intense SST gradients and cause the front to disappear.

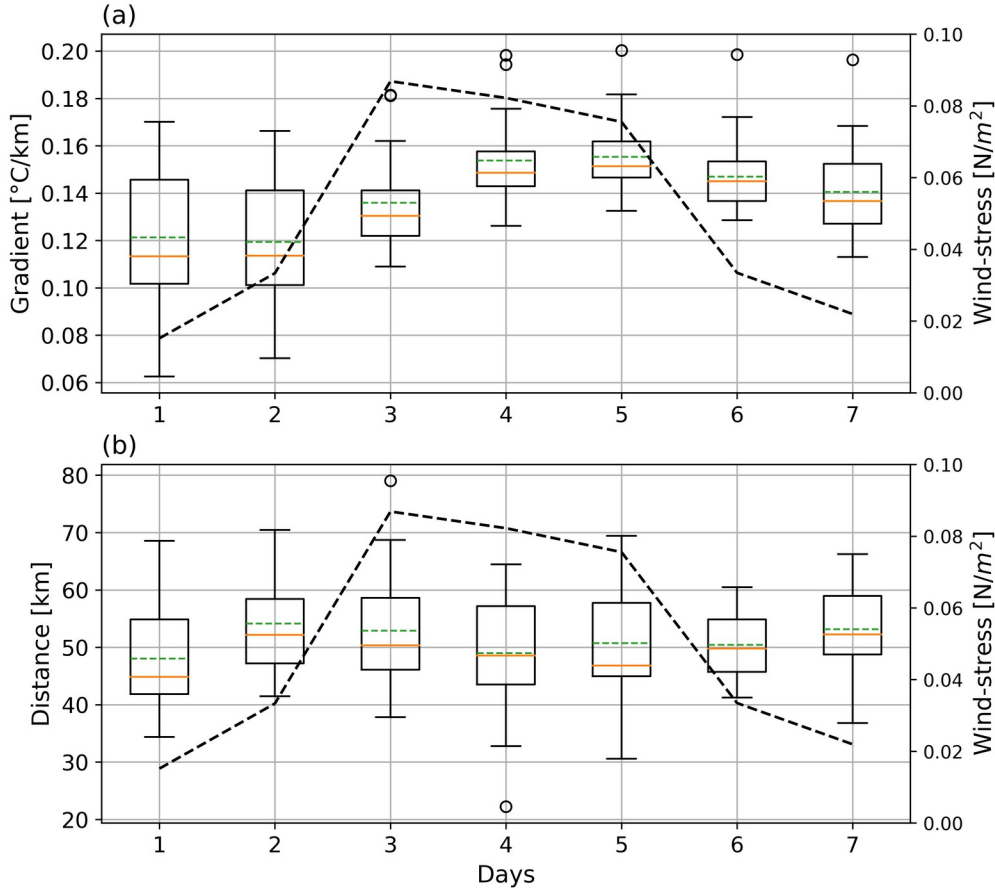


Figure 5.5: Box plots showing the time evolution of the frontal SST gradients (a) and displacement (b) for the parabolic wind profiles. Black dotted line shows the mean wind profile. The green and orange dotted lines show the daily mean and median values while statistical outliers are flagged as dots.

Fig. (5.6) shows identical frontal metrics for the 5 sustained wind profiles. The average wind profile for the sustained events has a somewhat parabolic curve; however, for all seven-days, the wind intensity is above the threshold value. Similar to the parabolic profiles (Fig. 5.5), there is greater variability in the gradient field for the first two days (Fig. 5.6a). It is unlikely that under these intense winds that an UF may not already be present and thus the sudden reduction in the IQR of the gradient field after day three may in fact correspond to the generation of a new UF along the coast. This is somewhat supported by the sudden increase in the IQR of offshore distance around the similar time (Fig. 5.6b). The offshore migration of the UF shows a similar pattern over the course of the wind-events to those in Fig. (5.5b). Taking the mean offshore position of the UF at day

seven, the parabolic wind profiles had similar offshore frontal positions to the sustained wind-events at 53 ± 7 km and 56 ± 7 km respectively. Again, the large IQR for offshore position from day four and onward suggests a high degree of variability such that the cross-shore displacement of the UF is not significant over the course of the wind-events. Similar to the parabolic wind-events, it is the intensification of the UF during these sustained wind-events that is more important for retention processes in the nearshore. Referring to Fig. (5.7) and Fig. (5.8), it is evident that the offshore position of the UF is more variable during sustained wind-events, especially along the broad Namaqua shelf. From a dynamics perspective, sustained events will generate greater offshore Ekman transport resulting in the UF being displaced further offshore; however, sustained events also stimulate more upwelling along the coastal margins and elevate the number of fronts present along the shelf. Therefore, while the mean position of the UF is similar for the two wind profiles, sustained wind-events are likely to be associated with more frontal activity.

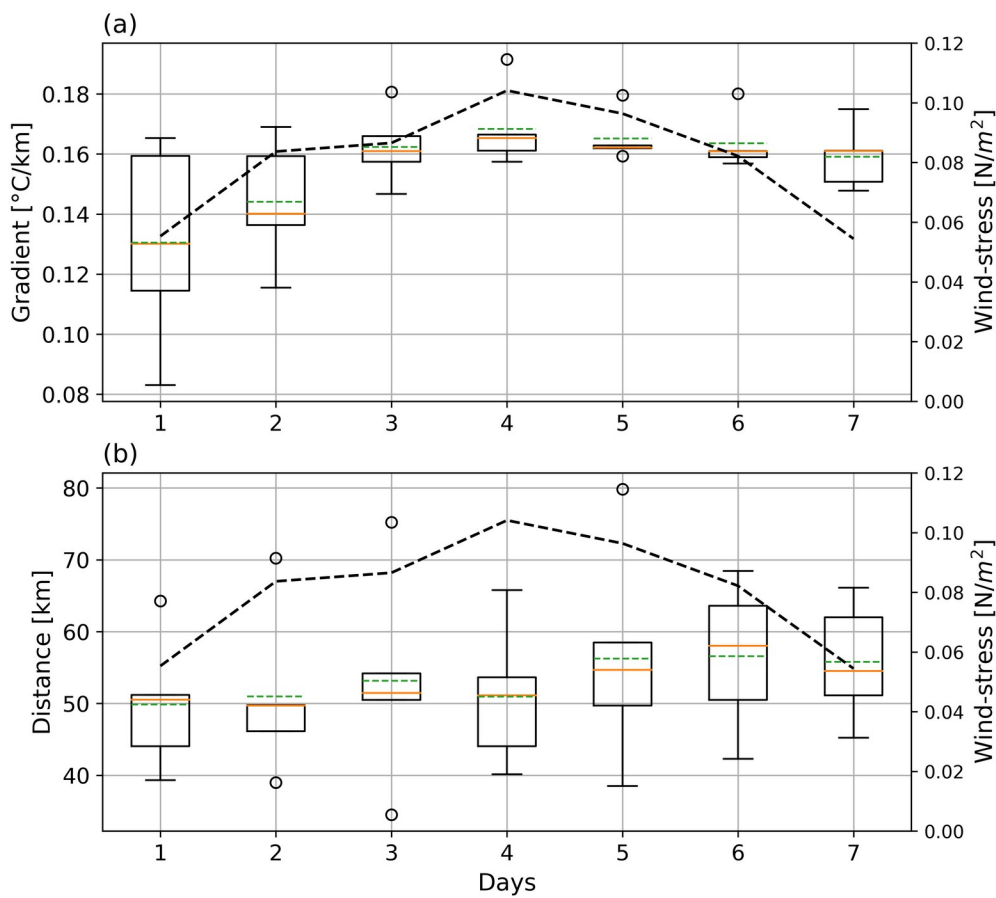


Figure 5.6: Box plots showing the time evolution of the frontal SST gradients (a) and displacement (b) for the sustained wind profiles. Black dotted line shows the mean wind profile. The green and orange dotted lines show the daily mean and median values while statistical outliers are flagged as dots.

It is not too surprising that the mean SST gradients are more intense for the sustained profiles than for the parabolic events. Sustained southerly winds generate more intense upwelling along the coast. The formation of intense cross-shore density gradients may add to the cross-shore retentiveness of nutrients but associated with more intense alongshore fronts are alongshore jets, which will act to transport nutrients away from their upwelling point of origin. The small sample size of sustained events does warrant some caution with this interpretation; however, the

implications of synoptic scale wind phenology impacting alongshore frontal intensities does seem apparent. Consequently, the interplay between more intense fronts limiting cross-shore dispersion but also accelerating the alongshore transport of matter is a unique dynamic to consider.

Example of a parabolic wind profile

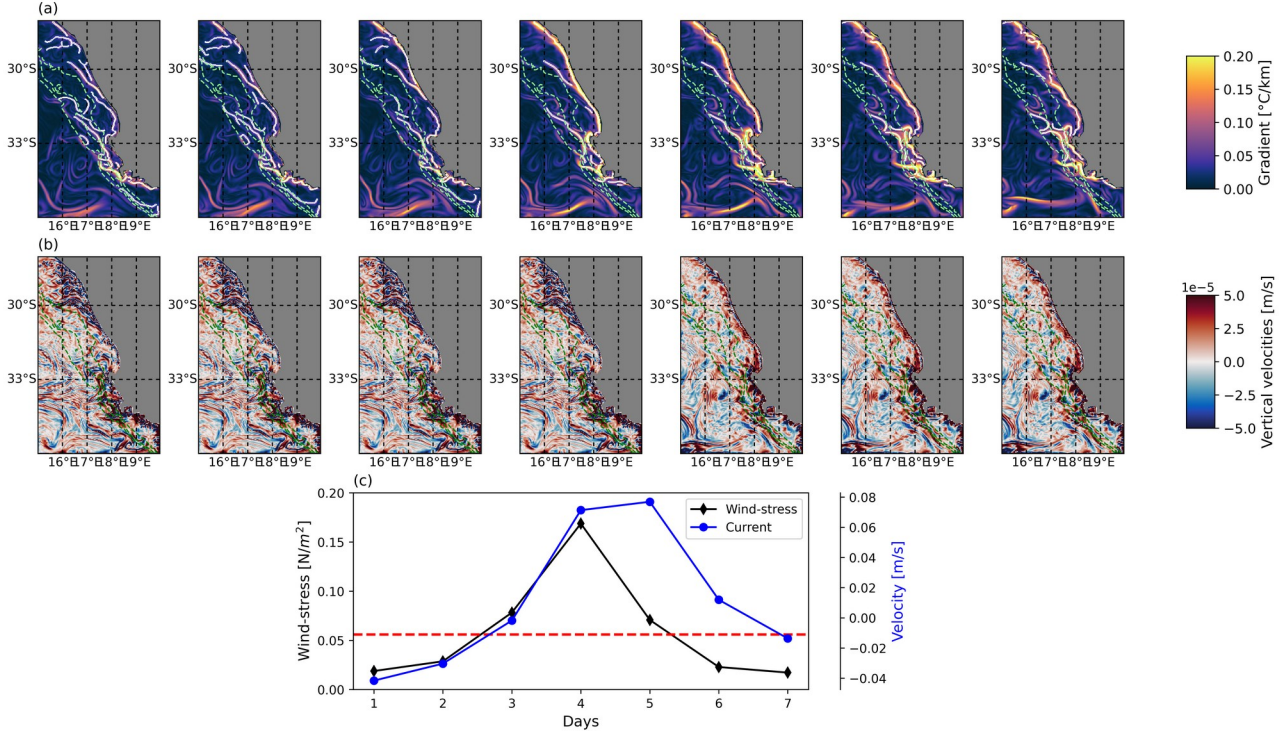


Figure 5.7: Parabolic seven-day summer wind-event example. (Top row) The evolution of the SST gradient over the course of the wind-event. Green dotted lines denote the 200, 300 and 500 m isobaths while the solid white lines indicate detected frontal edges from the Canny algorithm. (Second row) The event evolution of vertical velocities; red is indicative of upwelling and blue is downwelling. In panel (c), the spatial mean alongshore wind-stress and nearshore current velocities (averaged between the surface and 50 m) are shown. Red dotted-line is the wind threshold value.

The wind profile in Fig. (5.7) corresponds to event 13 in Fig. (5.4). A prominent parabolic summer wind-event drives upwelling along the coast and following a relaxation period results in the formation of a net alongshore poleward flow. As noted in Fig. (5.5), a high degree of spatial variability is present during the initial two days as the alongshore fronts are rather filamentous in nature (Fig. 5.7 top-row) and a patchwork of positive and negative vertical velocities are present along the coast and broad Namaqua shelf (Fig. 5.7 bottom-row). As the alongshore winds begin to intensify, a prominent cross-shore SST gradient develops along the coastal margin in response to nearshore upwelling, which is also marked by a near continuous UF. In conjunction with the formation of an UF, the nearshore currents switch to an equatorward flow and intense upwelling along the coast is evident by day 5. Over the course of the wind-event, the UF is displaced offshore.

The mean alongshore current velocity shows a direct relation to the wind forcing; however, a one day lag is evident between the peak in the wind-stress and alongshore current velocity (Fig. 5.7c). As the winds begin to subside, the alongshore current intensities weaken and a net poleward flow is established by day 7. At this time, the UF is still prominently defined, even during the relaxation phase in the wind. The presence of an intense and cohesive UF will likely add to the cross-shore

retention of nutrients while the current reversal during the wind relaxation phase will limit the northward displacement of material.

Example of a sustained wind profile

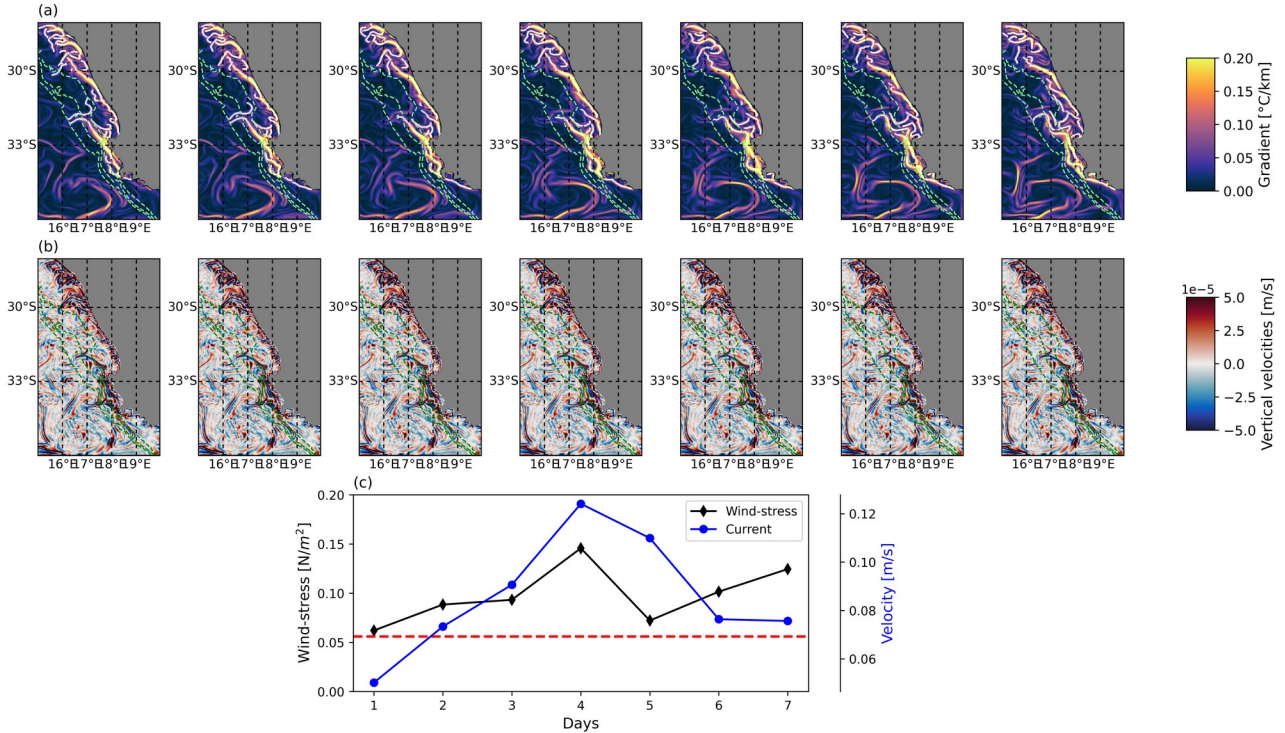


Figure 5.8: Sustained seven-day summer wind-event example. (Top row) The evolution of the SST gradient over the course of the wind-event. Green dotted lines denote the 200, 300 and 500 m isobaths while the solid white lines indicate detected frontal edges from the Canny algorithm. (Second row) The event evolution of vertical velocities; red is indicative of upwelling and blue is downwelling. In panel (c), the spatial mean alongshore wind-stress and nearshore current velocities (averaged between the surface and 50 m) are shown. Red dotted-line is the wind threshold value.

In Fig. (5.8), the oceanographic response in the SST gradients, vertical velocities and mean alongshore currents are shown for a sustained wind profile (event 1 in Supp. 5.1S). The seven-day window focusses on the oceanographic state of the region when the winds are above the threshold value but the prior quiescent phase would have been similar to the initial two days seen in Fig. (5.7). An UF is present along the coast as well as prominent upwelling in the nearshore. The UF is not as continuous relative to Fig. (5.7) and shows some variability in intensity and cohesiveness within St Helena Bay. Furthermore, multiple alongshore fronts are present along the Namaqua coast which may correspond to the formation of multiple UFs in the region. The extensive spatial variability in the alongshore wind-field is likely to explain these dynamics but sustained upwelling does elevate the number of frontal features in the region. Off of Cape Columbine, the shelf-break front, which is associated with the oceanic edge of the upwelling plume, is also clearly visible and defined by a strong SST gradient.

The mean alongshore current velocity shows a close coupling with the wind intensity and an equatorward flow prevails along the coast. Compared to Fig. (5.7), the more complex and numerous alongshore fronts in Fig. (5.8) are likely to result in greater cross-shore losses of material. Although more fronts may be present along the shelf, they are not as cohesive in alongshore structure and

some are even cross-shore orientated. The dominance of a sustained equatorward flow, less cohesive alongshore fronts as well as intense offshore Ekman transport will not favour the nearshore retention of material. This is in contrast to the parabolic wind-event in Fig. (5.7) as the more cohesive alongshore UF in conjunction with reduced offshore Ekman transport and less intense equatorward jets will elevate the nearshore retention of matter.

5.3.2 Lagrangian dynamics and transport along the coast

The previous section highlighted the role alongshore wind phenology has on affecting the nearshore current regimes as well as the offshore position of the UF. To explore the impact of variable wind profiles on the transport dynamics in the nearshore region Tab. (5.1) and Fig. (5.9) show the alongshore and cross-shore displacement of Lagrangian floats for the shelf region north of Cape Columbine during sustained and parabolic wind-events.

Tab. (5.1) summarizes the spatial patterns of Fig. (5.9) and evidently there is a statistically significant difference in the nearshore transport regimes under variable wind profiles. Southerly winds stimulate offshore Ekman transport which is epitomized in the pronounced cross-shore displacement of floats in the upper 20 m for both wind profiles as well as for floats between 30-50 m during sustained events. As seen in Fig. (5.6), intense upwelling generated by sustained southerly winds enhances the intensity of alongshore frontal/jet features leading to significantly greater alongshore transport relative to parabolic events.

Table 5.1: Spatial mean and standard deviation for the temporally averaged parabolic and sustained wind-events. A two-sample *t*-test not assuming equal variance and using an alpha level of 0.05 gives $p < 0.05$ across the various relaxation vs sustained indices.

	Parabolic	Sustained	Parabolic	Sustained
	[1 - 20 m]	[1 - 20 m]	[30 - 50 m]	[30 - 50 m]
Cross-shore [km]	53.0 $\pm$ 22.4	92.4 $\pm$ 26.9	-0.1 $\pm$ 15.1	34.5 $\pm$ 22.7
Alongshore [km]	17.1 $\pm$ 9.3	38.2 $\pm$ 20.1	-4.7 $\pm$ 10.9	13.9 $\pm$ 9.8
Total [km]	61.8 $\pm$ 18.3	104.1 $\pm$ 25.6	24.9 $\pm$ 8.8	45.3 $\pm$ 19.2

A stark contrast exists between the degree of displacement calculated for the upper 20 m and that between 30-50 m as a net equatorward flow is evident for the upper 50 m during sustained wind-events whereas parabolic events have an equatorward surface current atop a poleward flow. Under sustained wind-events, the pronounced southerly winds result in strong upwelling and mixing within the upper 50 m of the water column resulting in a net equatorward current between the surface and 50 m which becomes less intense with depth. During parabolic wind-events, there is an oscillation in the velocity of the mean alongshore surface currents in response to the phasing of the wind (Fig. 5.4a). Consequently, during the time in which wind intensities are above the threshold value, upwelling will similarly enhance mixing in the upper 50 m, generating an equatorward flow throughout the water column. However, during wind relaxation, there is the re-establishment of a poleward flow along the coast, which is more pronounced below 20 m. The weakening in the intensity of alongshore upwelling during wind relaxation results in a re-stratification of the water column and causes a slumping in the isopycnals close to the coast below 30 m which results in the formation of a poleward flow while the prevalence of a near-surface equatorward flow is sustained by weak southerly winds.

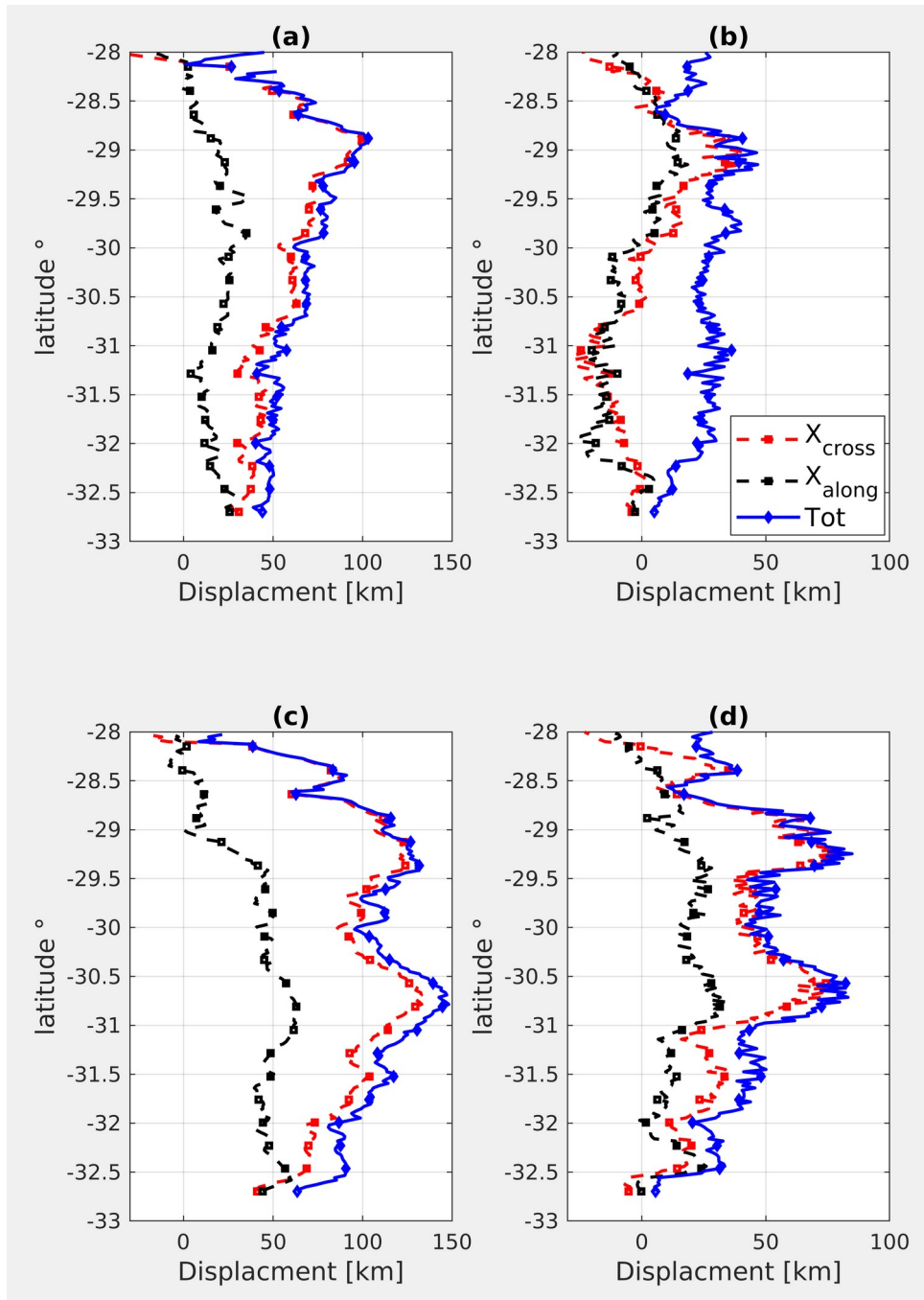


Figure 5.9: Binned depth averages for the float displacement metrics during parabolic (a, b) and sustained (c, d) wind-events for the upper 20 m (a, c) and lower 30-50 m (b, d). For parabolic vs sustained displacements there are statistical difference ($p < 0.05$).

Fig. (5.9) shows the temporally averaged displacement of particles during the seven-day sustained and parabolic wind-events for the shelf region north of Cape Columbine. In Fig. (5.9a, b), showing the parabolic wind-events, there is an expected decrease in the distance floats are transported with increasing depth but of interest is evidence of a net poleward transport of floats after initialisation for particles between 30-50 m (Fig. 5.9b); especially for the shelf region south of 30°S . From Tab. (5.1), the cross-shelf transport of floats between 30-50 m is near zero due to a switch in the

direction of net transport that occurs along the Namaqua shelf. The broadening of the shelf region north of 30°S may be responsible for inhibiting the formation of a poleward undercurrent. Applying the work of Gan & Allen (2002a), alongshore equatorward geostrophic currents are sustained by the upwelling-favourable winds. During wind relaxation, the geostrophic balance near the coast breaks down such that a counter-current develops due to the presence of a positive alongshore pressure gradient along the coast. Consequently, the offshore veering of the surface equatorward jets impedes the formation of a meridional pressure gradient along the coast such that a poleward flow cannot develop during wind relaxation. Upon further inspection, a poleward flow is also evident north of 28.5°S which corresponds to a narrowing of the shelf width. Reduced offshore transport is seen south of Namaqua and likely related to the spatial dynamics of the alongshore wind field. The wind shadow within St Helena Bay negatively impacts the intensity of the current velocities and the presence of the cyclonic circulation within the bay also contributes to a poleward flow in the region.

For sustained wind-events (Fig. 5.9c, d), a fairly homogeneous response is evident throughout the upper 50 m of the water column as floats are transported equatorward and offshore. Owing to the more sustained southerly winds, particles experience significantly greater displacement under sustained wind-events than during parabolic. Two prominent bulges in the meridional profile of total transport are evident at 31°S and north of 30°S. These regions correspond to the location of the Namaqua upwelling cell and where the shelf width broadens respectively. Similar to the parabolic profiles, both Fig. (5.9c and d) show decreased displacement within St Helena Bay as well as a drop in displacement north of 28.5°S. Critical to enhancing biological productivity as well as the formation of HABs is the retention of nutrients within the near coastal margins. Evidently, the prevalence of offshore transport under sustained wind-events is likely to generate high levels of biological activity but not be conducive for the formation of HABs along the coast. In contrast, the reduced displacement of Lagrangian floats during parabolic events as well as the presence of a net poleward flow south of 30°S will be critical in enhancing coastal retention as well as the transport of blooms and nutrients into St Helena Bay which will elevate biological productivity in the region and increase the likelihood of HABs and hypoxic conditions within the bay itself.

5.4 Discussion

5.4.1 *Impact of variable alongshore winds on nearshore retention*

Using a physical ocean model of the SBUS, in conjunction with a Lagrangian tracking code, the nearshore oceanographic dynamics under variable summer wind profiles were investigated to understand whether the alongshore wind phenology could influence the coastal retention times of nutrients and particles. In keeping with observational works such as Duncan & Nell (1969) and Fawcett *et al.* (2008), the numerical model was able to resolve sub-inertial current dynamics along the coast in response to the phasing of the synoptic southerly winds.

From Fig. (5.4) and Supp. (5.1S), the mean nearshore current velocities in the upper 50 m of the water column differed between parabolic and sustained wind-events. Prior to the onset of intense southerly winds, an equatorward surface current exists in the upper 20 m and below that a poleward undercurrent is present. This vertical current structure is typical for the region (Nelson 1989), with a summer MLD of 20 m (Fig. 3.9) dividing the surface equatorward and deeper poleward flows. During sustained wind-events, upwelling resulted in intense mixing within the upper 50 m, generating an equatorward flow throughout the water column that became less intense with depth. Parabolic events similarly had intense mixing within the upper 50 m when the winds were above the

threshold value, which resulted in a net equatorward flow; however, the onset of relaxation allowed for a re-stratification of the water column and a slumping of isopycnals below 30 m. The re-establishment of a poleward flow during wind relaxation as well as a reduction in offshore Ekman transport are important dynamics for elevating the nearshore retention of nutrients and potentially for the formation of HABs along the coast. How the poleward flow develops is important to understand.

The alongshore momentum equation can be written as follows (refer to Appendix section for all derivations):

$$\frac{Dv}{Dt} = \frac{-1}{\rho_0} \frac{\partial p}{\partial y} - fu + v(\nabla^2 v) \quad (5.3)$$

If we consider an initial condition along the coastal boundary such that $u=0$, an alongshore wind forcing τ_y exists and no steady-state assumptions are made, Eq. (5.3) can be rewritten:

$$\frac{Dv}{Dt} = \frac{\partial v}{\partial t} = \frac{-1}{\rho_0} \frac{\partial p}{\partial y} + v \frac{\partial^2 v}{\partial z^2} \quad (5.4)$$

In Eq. (5.4), the alongshore velocity component is influenced by the alongshore pressure-gradient force as well as vertical viscosity. Integrating Eq. (5.4) over the mixed-layer depth (H) and rearranging the terms yields:

$$\frac{\partial v}{\partial t} = \frac{-1}{\rho_0} \frac{\partial p}{\partial y} + \frac{\tau_y}{\rho_0 H} \quad (5.5)$$

Eq. (5.5) states that the coastal alongshore velocities are controlled by a balance between the meridional pressure-gradient force and the wind stress. During sustained wind events, the wind stress component exceeds the meridional pressure-gradient force generating an equatorward flow. During a relaxation phase in the wind, the wind stress weakens and the pressure-gradient force becomes the dominant term which results in the alongshore velocities becoming less intense or even reversing in direction.

The role of the alongshore winds in driving variability in the nearshore current velocities and alongshore fronts inevitably shapes the dynamics of retention along the coast. Consequently, HABs along the SBUS are likely to be more frequent during parabolic wind-events than sustained. During sustained wind-events, intense alongshore winds stimulate the formation of prominent alongshore fronts/jets which quickly transport particles and nutrients equatorward and offshore. The presence of a relaxation phase during parabolic wind-events not only results in the UF being found closer to the shore but also is conducive for the formation of a poleward flow along the coast (Eq. 5.5) which limits the northward and offshore transport of material. Therefore, the numerical modelling results corroborate the envisaged model of Pitcher & Nelson (2006) and support the idea that the elevated nearshore retention of dinoflagellate blooms during periods of wind relaxation will enhance HABs within St Helena Bay due to the development of a poleward flow along the coast that acts to transport nutrients and blooms south along the shelf region and into the bay itself (Fig. 2.4).

Another facet to nearshore retention is the role alongshore fronts play in restricting the offshore advection of material (Lamont *et al.* 2015; Flynn *et al.* 2020). Alongshore winds play a major role in shaping the alongshore frontal patterns in the region. Contrasting the two different wind profiles, sustained wind-events generally had more intense alongshore fronts, defined by greater SST gradients, as well as more frontal activity than parabolic events. From Tab. (5.1), alongshore transport was significantly greater during sustained wind-events than parabolic. Consequently, more intense equatorward jets will transport particles rapidly away from their upwelling point of origin. The spatial variability in the offshore position of the UF was greater during sustained wind-events such that the front was in general found further offshore; especially along the broad Namaqua shelf region. Furthermore, in contrasting Fig. (5.7) and Fig. (5.8), the UF tended to be less cohesive in alongshore structure for the sustained wind-event. While alongshore fronts will restrict the offshore advection of material, elevated mesoscale activity may in fact have a negative impact on productivity (Rossi *et al.* 2009; Gruber *et al.* 2011). Modelling work by Hauschildt *et al.* (2021) showed the subduction of recently upwelled nutrients along frontal interfaces to negatively impact primary productivity in the Peruvian Upwelling System. Consequently, elevated frontal activity during sustained wind-events may have a negative impact upon primary productivity. Therefore, the combination of a more spatially variable frontal environment, a less cohesive UF, more intense equatorward jets and greater offshore Ekman transport will not aid in the nearshore retention of matter during sustained wind-events.

5.4.2 Role of alongshore topography and bathymetry

The smaller scale orientation of the coastline and the presence of headlands and capes impacts the alongshore wind regimes such that regions of elevated upwelling as well as retention are blotted along the coastline of the SBUS (Pitcher & Jacinto 2019). From Fig. (5.9), there was evidence of bathymetric controls influencing the intensity of transport along the shelf as well as the meridional extents of the poleward undercurrents. Similar to Gan & Allen (2002a, b), the most intense alongshore current velocities were localized off of the prominent capes of Cape Point and Cape Columbine. The presence of capes act to amplify any shift in the alongshore pressure gradients in response to the phasing of the alongshore winds such that more intense poleward or equatorward currents can develop. The Cape Columbine headland creates a wind-shadow in the lee of the cape and from the Lagrangian experiments, weaker winds within St Helena Bay resulted in particles experiencing reduced displacement during both sustained and parabolic wind-events. The additional presence of a cyclonic eddy in the region also acts to reduce the equatorward and offshore transport of floats. Consequently, the alongshore coastal topography creates an ideal retention environment within St Helena Bay.

Along with coastal topography, the width of the shelf was found to impact the offshore positions of alongshore fronts as well as the formation of poleward currents along the coast. Taking a correlation of the total transport distance for floats (Fig. 5.9) relative to the width of the 200 m isobath (Supp. 5.2S) gives statistically significant correlations during parabolic wind-events as a correlation of 0.65 is calculated for the upper 20 m (Fig. 5.9a) and 0.30 for 30-50 m (Fig. 5.9b). However, no statistically significant correlations are seen for the sustained events. It is interesting that during parabolic events a statistical relationship is present between the total distance floats are transported and shelf width. Under sustained wind-events, bathymetric controls on retention were weak and the most intense offshore transport occurred where the shelf width was narrowest, between 31.5°S and 30°S (Fig. 5.9c, d). Although a narrower shelf does permit a shoreward migration of the alongshore fronts (Pfaff *et al.* 2022), Lachkar & Gruber (2012) found that a narrow continental shelf inhibits

net primary productivity as nutrients can quickly be advected offshore. The broad Namaqua shelf may not be conducive for the nearshore retention of nutrients, owing to the offshore veering of the surface equatorward currents, but particles are likely to be more easily retained on the coastal side of the 200 m isobath. An interesting result was the absence of a defined poleward flow during parabolic wind-events between 30°S and 28.5°S. Here, the widening of the Namaqua shelf results in the offshore veering of the surface equatorward currents. Applying the work of Gan & Allen (2002a, b), the absence of an intense nearshore current limits the development of a positive alongshore pressure-gradient (Eq. 5.4) such that a poleward flow cannot develop during wind relaxation. It seems then that a narrow shelf width favours the development of a poleward flow and thus blooms that are formed north of 30°S may be isolated from the rest of the shelf region and not be transported south.

5.4.3 *Acknowledging the importance of wind phenology in projections of change*

Interannual variability in wind intensity and frequency along the SBUS is driven by natural modes of variability associated with the position of the SAA, which can be impacted by ENSO as well as the Antarctic Annual Mode (Bordbar *et al.* 2023). In studying multi-decadal shifts in upwelling regimes within EBUS a common practice is to develop upwelling indexes that usually linearly relate upwelling intensity to productivity. Our results suggest that it is important to appreciate the differing affects of variable alongshore winds as they will generate distinct responses in the nearshore currents and alongshore fronts. Shifting these physical dynamics will impact the nearshore retention times of nutrients and consequently have an impact on biological productivity. Complementing our findings, Botsford *et al.* (2003, 2006) noted a drop-off in biological productivity associated with strong winds due to particles and nutrients being too rapidly transported offshore,

On synoptic scales there is a direct correlation between wind intensity and gross primary productivity (Botsford *et al.* 2006); however, the phenology of the wind itself will play a part in dictating the levels of primary productivity and may even impact the phytoplankton assemblages themselves (Fawcett *et al.* 2007; Pitcher *et al.* 2010). Consequently, the simple upwelling index of Bakun (1990) of increasing wind intensity generating greater primary productivity does not necessarily hold. As wind intensities are set to increase due to climate change, the interplay between stronger alongshore winds and increasing surface ocean stratification will be complex and non-linear (Bograd *et al.* 2023). How these non-linear physical interactions will impact primary productivity within EBUS is a topic of ongoing research. To add to this discussion is the role variable alongshore winds will play in impacting productivity in the nearshore regions of EBUS. Therefore, how synoptic cycles may shift in a changing climate warrants consideration when undertaking projections of change within EBUS.

5.5 Conclusion

Alongshore winds play a major role in influencing the physical oceanography and biological productivity within EBUS. Using a physical ocean model of the SBUS, this work has shown the role variable alongshore winds play in shaping the surface nearshore current velocities as well as the offshore position and intensity of the UF. Variability of these physical features influence the degree of nearshore retention for particles in the region. For the study, two types of summer, alongshore, seven-day, southerly wind profiles were investigated; parabolic and sustained wind-events and a threshold value was used to delineate significantly intense events.

In general, parabolic wind-events promote greater nearshore retention than sustained events due to the UF being in closer proximity to the coast as well as the presence of a relaxation phase in the wind that favours the development of a poleward current along the coast which reduces the equatorward and offshore advection of material. Sustained wind-events promote intense mixing within the upper 50 m of the water column; generating an equatorward flow that becomes less intense with depth. The absence of a wind-relaxation phase precludes the formation of a poleward counter-current such that particles and nutrients are quickly displaced offshore and equatorward. Furthermore, enhanced mesoscale activity during sustained wind-events may negatively impact primary productivity as a consequence of recently upwelled nutrients being quickly advected offshore or subducted along frontal interfaces before they can be assimilated. Variations in coastal topography and bathymetry also influence nearshore retention. The presence of the Cape Columbine headland creates a wind-shadow in the lee of the cape which creates an ideal retention environment within St Helena Bay as particles will experience reduced displacement. Furthermore, a narrow shelf width is more conducive for the formation of a poleward flow during parabolic wind-events. Therefore, HABs are likely to more frequent during parabolic wind-events and the development of a poleward flow along the coast will act to transport material south and into St Helena Bay which will add to nutrient loading in the region and potentially generate hypoxic conditions within the bay.

The alongshore upwelling wind phenology generates unique and differing responses in the nearshore current velocities and alongshore frontal positions such that it will impact biological productivity within the SBUS. Anthropogenic climate change is set to increase the intensity of alongshore coastal winds and recognizing the non-linear relation between wind intensity and productivity is an important dynamic to consider regarding the computation of upwelling indexes on productivity as well as forecasting biogeochemical shifts within EBUS. Alongshore winds show significant spatial variability along the SBUS due to the influence of the variable alongshore topography. Future work could explore intra-regional responses in the alongshore oceanography of the SBUS to variable alongshore wind profiles by sub-dividing the shelf region into smaller domains. One major limitation to the study was the differing number of parabolic vs sustained wind-events that were detected. Although the desire was to understand the 3-dimensional oceanographic responses during near realistic synoptic forcing winds; a future study could perturb the atmospheric forcing winds by increasing their intensity in order to study the responses of the coastal oceanography to more intense and sustained alongshore wind conditions.

Supplementary

5.1S Hovmöller plot of mean surface current velocities during sustained wind-events

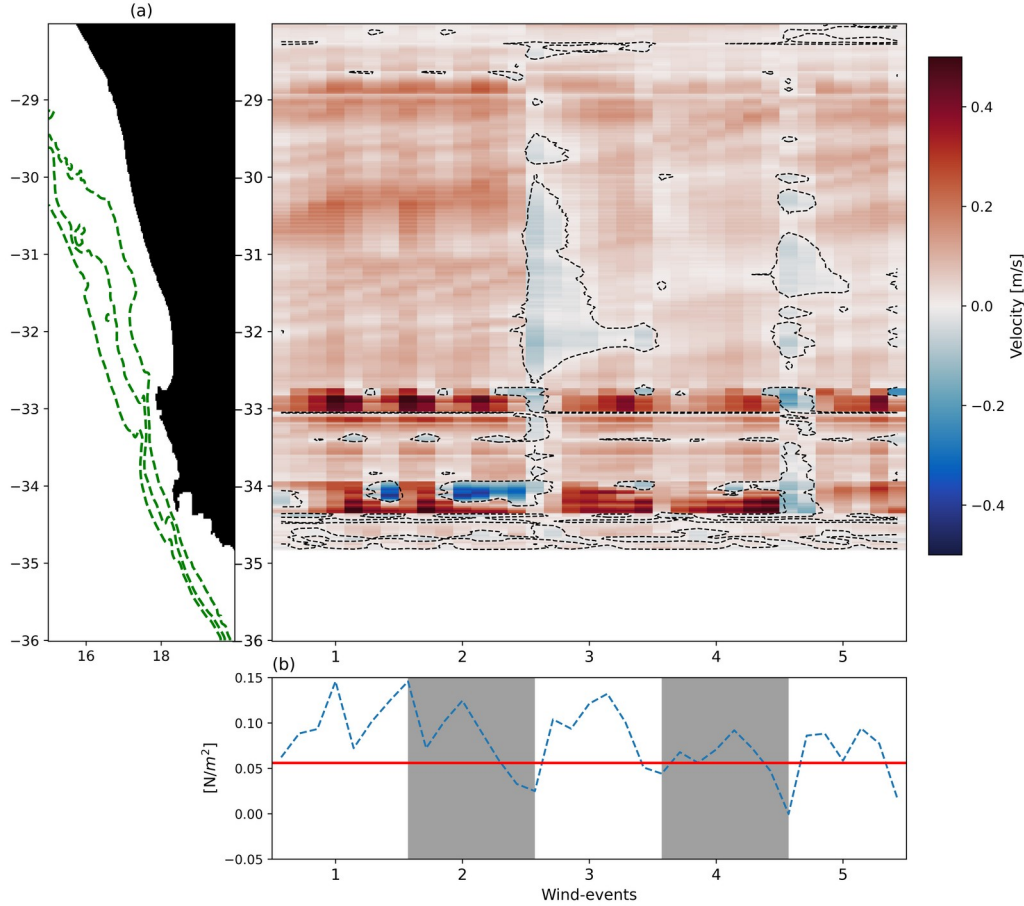


Figure 5i: (a) Hovmöller plot of the vertically averaged alongshore current between the surface and 50 m for the various discrete sustained summer wind-events. The green-dotted lines show the 200, 300 and 500 m isobaths. Important to note that the events are not consecutive. (b) The various alongshore wind profile, with the red-line denoting the threshold value. The grey and white banding in (b) just assists in readability.

5.2S Width of the 200 m isobath

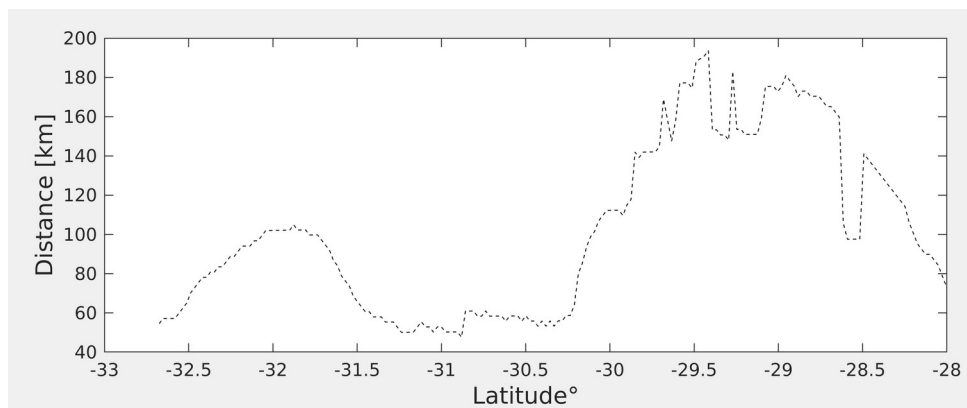


Figure 5.ii: Offshore distance of the 200 m isobath parallel to the shore

Chapter 6: Key findings and final discussion

Cogito, ergo sum – Descartes

Using a numerical model, this thesis has addressed the aims of the study by investigating and showcasing the major physical oceanographic features that contribute to and elevate the shelf-bound retention of matter and nutrients on the SBUS shelf. This chapter encompasses the major findings of chapters (3), (4) and (5). Three sections are presented: (1) the major retention processes in the SBUS, (2) uniqueness of retention dynamics and specificities to the SBUS, and (3) further avenues of research.

6.1 *Major retention processes in the SBUS*

The physical oceanographic variability of mesoscale/submesoscale fronts, shelf-based circulation patterns and nearshore currents influence the retention/residence times of matter along the SBUS. These physical features often coexist and are primarily driven by variability in the alongshore wind regimes, ocean stratification as well as variations in coastal topography and shelf bathymetry. When using the word ‘retention’, although somewhat of a scalar element, it is important to appreciate the spatial and temporal domains that define said retention. Consequently, scales of variability are important to consider; especially within the SBUS, as retention dynamics are not uniform along the shelf region and retention shows seasonal, interannual and synoptic scale variability.

At the seasonal and interannual scale, residence times of material along the SBUS shelf are primarily a function of the shelf-based current intensities. In summer, wind-driven equatorward surface jets, wind-stress induced poleward undercurrents and offshore Ekman transport are more intense resulting in passive Lagrangian particles being rapidly advected and transported out of the coastal shelf domain. Similar to the modelling results of Lett *et al.* (2006) and Miller *et al.* (2006), residence times throughout the upper 250 m of the water column were calculated to be a factor of 2-3 times longer in winter. It is important to appreciate that longer residence times in winter do not equate to elevated levels of biological activity or risks of hypoxia developing due to the absence of upwelling-favourable winds and an uncondusive frontal environment.

Superimposed on the broad seasonal shelf circulation patterns, perennial, seasonal and/or episodic frontal features create localised regions of enhanced retention. Critical to biological productivity is the role alongshore fronts play in enhancing the retention of nutrients and phytoplankton within the nearshore regions as well as restricting the advective losses of shelf-based biological material into the open ocean. A major finding was linking upwelling-favourable southerly winds and a strongly stratified surface water column during summer to the genesis of long and continuous frontal features which form closer to the coast and are defined by prominent SST gradients and associated with distinct Lagrangian Coherent Structures (LCS). Episodic southerly winds generate upwelling and divergence along the coast resulting in a contiguous UF forming parallel to the shore. Along the shelf-edge, the influence of warm Agulhas waters sustains intense cross-shore SST gradients which generates a near continuous and quasi-stationary SBF. Furthermore, southerly winds promote the formation of an upwelling plume off Cape Columbine and the oceanic appendage of this plume

manifests itself as the CF. The novel association of these frontal features, identified in the observational work of Lamont *et al.* (2015), with LCS furthers the knowledge base that fronts in the SBUS can act as cohesive barriers to the offshore transport of material. Furthermore, the results support the findings of Flynn *et al.* (2020) such that fronts locally enhance the retention times of nutrients and ‘trap’ them along the shelf which leads to greater primary productivity.

In winter, a reduction in solar insolation leads to an overturning and deep mixing of the water column resulting in a weakly stratified surface layer and a near homogeneous cross-shore SST environment. The coupled non-linear effect of weakened stratification and a shift to predominantly northerly winds was found to promote the formation of more numerous but less intense fronts along the shelf. Defined by weaker SST gradients and being statistically smaller in length, fronts in winter tend to be found further offshore and are not associated with defined LCS. From Callies *et al.* (2015), He & Mahadevan (2021) and Prants (2022), a more stratified surface layer in summer acts to limit weak mesoscale activity such that summer frontal features are associated with intense baroclinic instabilities. Consequently, weak stratification of the water column in winter was found to play a major role in shaping the high spatial variability of the seasonal frontal environment. The SBF remains quasi-stationary along the shelf-edge albeit less intense and the absence of upwelling-favourable winds precludes the formation of the CF. Between Cape Columbine and 30°S, the near overlap in the offshore positions of the SBF and UF was tantamount to the absence of a clear UF, which is expected; however, perennial southerly winds still sustained a front in close proximity to the shore along the Namaqua shelf region. Although fronts in winter do not play a dynamic role in retaining matter along the shelf, the seasonal erosion of prominent alongshore frontal gradients may facilitates the flushing and re-oxygenation of bottom waters along the shelf (Rixen *et al.* 2021) and elevate the cross-shore exchange of water masses.

Complementing the frontal features of the SBUS are several shelf-based surface and deep circulation patterns that contribute to the retention dynamics in the SBUS. Similar to the works of Penven *et al.* (2000), Hutchings *et al.* (2009) and Brandt *et al.* (2024), prominent circulation features that enhance retention along the coast include the perennial cyclonic circulation within St Helena Bay, cross-shore recirculation cells at 31°S and multiple upwelling cells along the broad Namaqua shelf. Crucially, these features were shown to have a seasonal cycle, being more intense in summer than winter. Far from being a constant value, residence times for summer and winter showed significant interannual variability throughout the upper 250 m of the water column. Residence times in the upper 50 m are driven by variability in the offshore position of the alongshore fronts and the intensity of offshore Ekman transport while the intensity of the broad poleward undercurrents are the main drivers of residence times for particles below 100 m. In general, negative/positive anomalies in Ekman transport correspond to fronts being found further/closer to the coast which results in lower/higher residence times for particles in the surface layer whereas positive/negative anomalies in the intensity of the poleward undercurrent generate longer/shorter residence times for particles below 100 m. These broad dynamical responses are applicable to the SBUS.

A major driver of interannual variability for the retention, transport and mixing patterns of material are the upwelling-favourable wind regimes. They impact the intensity of offshore Ekman transport, the positions of the alongshore fronts and through wind-stress curl affect the poleward undercurrents along the SBUS. Therefore, the role of remote forcing and large-scale modes of variability such as El Niño–Southern Oscillation (ENSO) and the Antarctic Oscillation can influence the atmospheric circulations along the Southeast Atlantic (Rouault, Pohl & Penven 2010;

Tim, Zorita & Hünicke 2015; Rouault & Tomety 2022). A key and novel finding was maximums in retention occurred at intervals of 4-5 years, corresponding to the periodicity of ENSO. Variability in residence times were not restricted to just the surface layer but seemed to be a transitive property of the whole water column. During La Niña, stronger upwelling-favourable winds are present whereas El Niño phases are associated with significantly weaker south-easterly winds during summer (Rouault *et al.* 2010; Rouault & Tomety 2022).

To investigate the link between interannual variability in shelf residence times and ENSO the Southern Oscillation Index (SOI) was used. The SOI is a standardised index based on the sea level pressure differences between Tahiti and Darwin, Australia (Ropelewski & Jones 1987). Prolonged periods of negative/positive SOI values correspond with El Niño/La Niña episodes. In summer, a negative correlation of -0.46 ($p \approx 0.08$) was seen for residence times vs SOI computed at the surface and -0.34 for -50 m ($p \approx 0.22$). Throughout the rest of the water column in summer and for all depths in winter p -values > 0.3 such that there was no relation. Though not significant at the 95% level, the pronounced negative correlations for the upper 50 m in summer gives some evidence for ENSO impacting the residence times of material on the SBUS shelf and if anything adds robustness to the approach used to compute the residence times because the results show some correspondence to remote forcing signals. Therefore, during El Niño/La Niña periods, the weaker/stronger upwelling-favourable winds during summer are likely to lead to higher/lower residence times along the shelf. This natural interannual variability in residence times along the shelf is important to consider when doing projections of marine productivity and risks of hypoxia developing.

The variability of alongshore synoptic winds near coastal margins play a major role in driving variability in the nearshore current profiles, offshore positions of alongshore fronts as well as significantly influencing the productivity of coastal systems through modulating the fluxes of nutrients. Episodic upwelling events operate on a synoptic scale of 2-8 days and the intensity and duration of these upwelling winds impacts the oceanography and retention dynamics along the coastal margins of the SBUS. Contrasting two types of seven-day, southerly, summer wind profiles; parabolic wind-events promoted greater nearshore retention than sustained events. Prior to the onset of intense southerly winds, an equatorward surface current is observed within the upper 20 m and below that a poleward undercurrent is present. During sustained wind-events, upwelling results in intense mixing within the upper 50 m of the water column, generating an equatorward flow that dissipates with depth. Parabolic wind-events similarly have intense mixing within the upper 50 m when the winds are of sufficiently high intensity which generates an equatorward flow; however, the onset of a relaxation phase allows for a re-stratification of the water column and a slumping of isopycnals below 30 m. The re-establishment of poleward flow during wind relaxation as well as a reduction in offshore Ekman transport are important dynamics for elevating the nearshore retention of nutrients and potentially for the formation of HABs along the coast.

Along with the southerly winds, the width of the shelf and coastal topography was also found to play a crucial role in shaping the nearshore current dynamics. The presence of the Cape Columbine headland creates a wind-shadow in the lee of the cape which creates an ideal retention environment within St Helena Bay. During both wind-events, particles experienced reduced displacement relative to other regions along the shelf. The high retentiveness of St Helena Bay has been noted in many studies such as Penven *et al.* (2000), Hutchings *et al.* (2009) and Pitcher *et al.* (2010) such that the finding aligns with the current literature. The formation of a poleward current following wind relaxation phases was crucially found to be limited to regions where the shelf-width was narrow. Using the work of Gan & Allen (2002a, b), the broadening of the Namaqua shelf between

30°S and 28.5°S steers the surface equatorward jets further offshore and this likely limits the formation of a positive alongshore pressure gradient near the coast. Consequently, during wind relaxation phases, a poleward current does not clearly develop in this specific region of the shelf. This may lead to blooms which form over the Namaqua shelf being isolated from the rest of the SBUS shelf region due to the absence of a defined poleward flow. Similar to Lachkar & Gruber (2012), particles were found to be rapidly displaced offshore where the shelf-width was narrow and this is likely to have a negative impact on productivity as nutrients may be advected out of the coastal domain before they can be assimilated. Of significance, the numerical experiments corroborated the envisaged scenario of Pitcher & Nelson (2006) and support the idea that the elevated nearshore retention of dinoflagellate blooms during wind relaxation will enhance HABs within St Helena Bay due to the development of a poleward flow along the coast that acts to transport nutrients and blooms south along the shelf region and into the bay itself.

6.2 *Uniqueness of retention dynamics and specificities to the SBUS*

Most of the physical dynamics that contribute to retention within the SBUS are shared in some respect by the other major EBUS and as such the critical distinguishing attribute for the SBUS is the close proximity and influence of the warm Agulhas Current. The ubiquitous nature of equatorward alongshore winds within EBUS originate from the basin-wide large-scale atmospheric circulations. Variations in coastal topography, such as changes in coastline orientation or the presence of capes and bays, modify these coastal winds which leads to spatial variability in upwelling intensity. Wang *et al.* (2015) showed SST frontal patterns to be remarkably similar in all EBUS, with high frontal activity existing close to the coast and decreasing offshore. The seasonal frontal probabilities computed in this study parallel the findings of Wang *et al.* (2015). Of significance, within the SBUS is a perennial SBF that is maintained by the intense cross-shore gradients between the cool shelf waters and warm offshore waters which are supplied by Agulhas leakage. Off Cape Columbine, the additional presence of the CF in summer adds to the number of alongshore frontal features; however, the presence of multiple alongshore fronts is not unique to the SBUS. Roy (1998) documented the presence of two cross-shore upwelling fronts off Senegal and Oerder *et al.* (2018) noted several alongshore jet features within a region of the Chile Upwelling System.

A comprehensive review by Pitcher *et al.* (2010) showcases the breadth and complexity of coastline features present within EBUS that can impact shelf circulation patterns and create zones of retention. Along with St Helena Bay, Pitcher *et al.* (2010) makes reference to features such as Monterey Bay in the California Upwelling System and Rías along the Iberian coast. Inevitably, a distinctive facet of any EBUS is their unique coastal topography. Within the SBUS, the cyclonic circulation within St Helena Bay as well as the varying width of the shelf creates a physical oceanographic environment that is especial to the region. In this study, the shelf-width was found to influence the offshore positions of the UF and SBF as well as impact the formation of a poleward flow during wind relaxation events. Along with coastal topography, EBUS experience variable oceanographic responses during ENSO, with EBUS in the Pacific showing greater sensitivity (Fréon *et al.* 2009).

The close proximity and influence of a warm western boundary current to an eastern boundary current system is unique to the SBUS and so too is the region's coastal topography. From the modelling work of Veitch & Penven (2017), warm waters from the Agulhas current that leak into the Atlantic Ocean are critical for sustaining an intense and perennial shelf-edge jet, referred to as the CP jet, between Cape Point and Cape Columbine. The CP jet plays a crucial role in the transport

of pelagic larvae from the Agulhas Bank to nursery grounds within St Helena Bay (Penven *et al.* 2001; Ragoasha *et al.* 2019) and the intense baroclinic nature of the jet was found to exert a significant influence on the trajectory patterns of passive floats below 100 m. While particles were transported poleward along the narrow shelf between Cape Columbine and Cape Point, the vertical extent of the CP jet acted as a barrier to the offshore advection of deep flowing particles. This finding contributes to the retentive importance of the jet and showcases a clear example where alongshore fronts can have a significant influence on particle trajectories throughout the upper 250 m of the water column. North of Cape Columbine, the jet feature was found to become more intense closer to the shelf-edge which is in keeping with previous findings such as the model work of Veitch *et al.* (2018) and the observational work of Lamont *et al.* (2015).

In conclusion, although the broad physics that govern upwelling are ubiquitous in all four EBUS, the influence of the Agulhas Current and especial coastal topography ultimately create a unique oceanographic environment for retention within the SBUS region. Consequently, it would be inappropriate to infer retention dynamics within other EBUS without proper consideration of the complex physical oceanographic interactions that take place. Therefore, regional specific studies are required to comprehensively understand retention dynamics and processes within an individual EBUS.

6.3 Further avenues of research

For the study, a physical ocean model configuration for the SBUS was designed and comprehensively evaluated against a combination of satellite and *in situ* data products in order to investigate the retention dynamics along the shelf region. Appreciating how significant water residence times are for biogeochemical cycles such as oxygen and nitrate cycling, a further research step would be to couple the current physical ocean model configuration with a biogeochemical model to resolve the biogeochemistry in the region. Coupled physical-biogeochemical model studies in the Benguela have previously been conducted by Koné *et al.* (2005) and Gutknecht *et al.* (2013). Both studies focussed on primary productivity and nutrient cycling within the region. Gutknecht *et al.* (2013) also highlighted the importance of resolving oxygen dynamics within EBUS.

Oxygen is fundamental to life and plays a significant role in the regulation of nutrients and the carbon cycle in the world's oceans (Breitburg *et al.* 2018). The concentration of oxygen in the ocean is dependent upon multiple non-linear processes such as air-sea fluxes, physical transport and biogeochemical processes that produce and consume oxygen in the water column and sediments (Peña *et al.* 2010). In general, oxygen is produced through photosynthesis and consumed during the respiration and remineralization of organic matter, nitrification and redox reactions. The availability of oxygen implicates other biogeochemical cycles of key nutrients such as phosphate, nitrate and iron. Furthermore, oxygen concentrations and availability have a pronounced impact on the biodiversity and habitability of marine ecosystems (Breitburg *et al.* 2018). Pertinent to the study of oxygen is the increase in hypoxia conditions ($O_2 < 60 \mu M$) throughout the world's oceans associated with global warming which results in a reduction in solubility of oxygen and increasing ocean stratification that hinders ventilation of the deep ocean (Breitburg *et al.* 2018). Extensive observational work has been conducted by Pitcher *et al.* (2014), Jarre *et al.* (2015), Pitcher & Probyn (2017), Pitcher & Jacinto (2019) and Pitcher, Probyn & du Randt (2022) on oxygen dynamics and cycling within the Benguela. Studies such as Jury (2014) have attempted to

understand the environmental forcing mechanisms conducive for the development of harmful red tides in an effort to develop methods of forecasting these events.

A coupled physical-biogeochemical model configuration of the SBUS presents an opportunity to conduct sensitivity experiments in order to investigate variously, the relative importance of wind forcing, stratification and alongshore pressure-gradients on hypoxia, subsurface nitrogen budgets and nutrient trapping efficiency along the shelf. Such research would further advance understanding of the key physical oceanographic processes that impact nutrient cycling and lead to the development of hypoxia within the SBUS.

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Appendix

Equations of motion on a static framework

The total derivative (Eq. 1.0) expresses the change of properties of a particle based on Eulerian fields and the vector $\mathbf{u} = (u, v, w)$.

$$\frac{D}{Dt} \equiv \frac{\partial}{\partial t} + \mathbf{u} \cdot \nabla \quad (1.0)$$

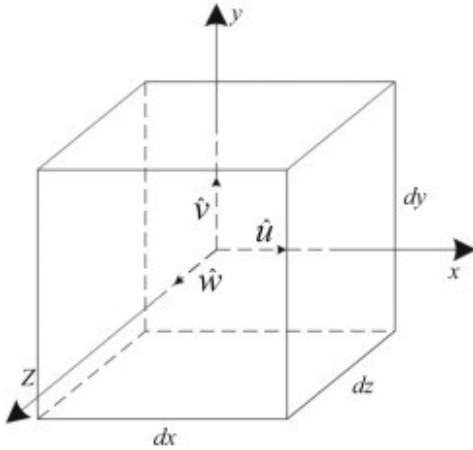


Figure 1A: Fixed volume

Fig. (1A) illustrates a controlled infinitesimal volume with mass (δM) and density (ρ). The combination of forces acting on the volume is governed by Newton's First Law (Eq. 1.1) and in the non-rotating framework this constitutes: gravity (F_g), pressure-gradient (F_p) and friction (F_f).

$$\mathbf{F} = \delta M \mathbf{a} = \rho \delta x \delta y \delta z \frac{D\mathbf{u}}{Dt} \quad (1.1)$$

The effects of gravity is independent of geometric considerations and only based on the mass of the infinitesimal volume. Using a constant gravitational constant (g), F_g is directed downward along the unit vector $\hat{\mathbf{k}}$ (Eq. 1.2).

$$F_g = -g \delta M \hat{\mathbf{k}} = -g \rho \delta x \delta y \delta z \hat{\mathbf{k}} \quad (1.2)$$

Pressure (p) along a plane acts at opposite sides of the volume. Along the x-component, the pressure on one side will be $p(x)$ while on the other side $p(x+\delta x) = p + \delta p$. Consequently, pressure along the x-component faces of the volume is written as:

$$F_p^x \hat{\mathbf{i}} = p(x) \delta y \delta z - [p(x) + \frac{\partial p}{\partial x} \delta x] \delta y \delta z$$

Solving and simplifying the pressure along F_p^y and F_p^z gives the net pressure (Eq. 1.3).

$$\mathbf{F}_p = -\nabla p \delta x \delta y \delta z \quad (1.3)$$

Frictional forces have no preferential direction and for the mean time will be written as forces per unit mass where $\mathfrak{F}$ represents the forces (Eq. 1.4).

$$\mathbf{F}_f = \rho \mathfrak{F} \delta x \delta y \delta z \quad (1.4)$$

Combining Eq. (1.2 – 1.4) with Eq. (1.1):

$$\rho \delta x \delta y \delta z \frac{D\mathbf{u}}{Dt} = \mathbf{F}_p + \mathbf{F}_g + \mathbf{F}_f$$

and substituting all the terms yields

$$\rho \delta x \delta y \delta z \frac{D\mathbf{u}}{Dt} = -\nabla p \delta x \delta y \delta z - g \rho \delta x \delta y \delta z \hat{\mathbf{k}} + \rho \mathfrak{I} \delta x \delta y \delta z$$

After simplifying, Eq. (1.5) gives the generic form of the equation of motion for a fluid particle in a static framework.

$$\boxed{\frac{D\mathbf{u}}{Dt} = -\frac{\nabla p}{\rho} - g \hat{\mathbf{k}} + \mathfrak{I}} \quad (1.5)$$

Hydrostatic balance

Hydrostatic balance (Eq. 1.6) in a fluid is achieved when gravity balances with the pressure-gradient force. Considering our fixed volume (Fig. 1A), it will be static only if all the forces acting on it are equal. Writing only the vertical component of Eq. (1.5) and neglecting friction gives

$$\frac{\partial w}{\partial t} = -\frac{1}{\rho} \frac{\partial p}{\partial z} - g$$

As all forces are balanced, there will be no acceleration in the vertical, $\frac{\partial w}{\partial t} = 0$. Consequently, the hydrostatic balance simplifies to

$$\boxed{\frac{dp}{dz} = -\rho g} \quad (1.6)$$

Continuity equation

Referring to Fig. (1A), the mass conservation equation can be derived from a Eulerian point of view using a reference frame which is fixed in space and the fluid flows through it. Only considering the flow of mass in (Q_{in}) and out (Q_{out}) of a control volume, fluid velocity is the only vector while density is a scalar field. Consequently, the Eulerian time rate of mass change in the fixed volume is obtained with the net flow through the faces (Eq. 1.7).

$$\frac{\partial \rho}{\partial t} \delta x \delta y \delta z = (Q_{in} - Q_{out})_x + (Q_{in} - Q_{out})_y + (Q_{in} - Q_{out})_z \quad (1.7)$$

The mass flow is the amount of mass per unit time passing through the faces. The x-component can be written as:

$$Q_{in} = [\rho(x)u(x)]_x \delta y \delta z$$

$$Q_{out} = [\rho(x+\delta x)u(x+\delta x)]_x \delta y \delta z$$

$$Q_{out} = [(\rho+\delta\rho)(u+\delta u)]_x \delta y \delta z$$

The x-component of the net flow can then be expressed as:

$$(Q_{in}-Q_{out}) = \rho u \delta y \delta z - [(\rho+\delta\rho)(u+\delta u)]_x \delta y \delta z$$

In solving the second term, we use the following relationships: $\delta u = \frac{\partial u}{\partial x} \delta x$ and

$\delta\rho = \frac{\partial\rho}{\partial x} \delta x$ such that the second term, when expanded using the substituted relationships, becomes:

$$(\rho u + \rho \delta u + u \delta\rho + \delta\rho \delta u) \delta y \delta z = (\rho u + \rho \frac{\partial u}{\partial x} \delta x + u \frac{\partial\rho}{\partial x} \delta x + O(\delta\rho \delta u)) \delta y \delta z$$

The x-component mass flow can now be expressed as

$$(Q_{in}-Q_{out})_x = \rho u \delta y \delta z - \rho u \delta y \delta z - (\rho \frac{\partial u}{\partial x} + u \frac{\partial\rho}{\partial x}) \delta x \delta y \delta z$$

$$(Q_{in}-Q_{out})_x = -(\rho \frac{\partial u}{\partial x} + u \frac{\partial\rho}{\partial x}) \delta x \delta y \delta z$$

After solving the x,y and z components, Eq. (1.7) becomes

$$\frac{\partial\rho}{\partial t} = -(\rho \frac{\partial u}{\partial x} + u \frac{\partial\rho}{\partial x}) - (\rho \frac{\partial v}{\partial y} + v \frac{\partial\rho}{\partial y}) - (\rho \frac{\partial w}{\partial z} + w \frac{\partial\rho}{\partial z})$$

Writing the above in terms of the total derivative yields the continuity equation.

$$\boxed{\frac{D\rho}{Dt} + \rho \nabla \cdot \mathbf{u} = 0} \quad (1.8)$$

Boussinesq equations

The Boussinesq approximation assumes that the density of an incompressible fluid remains constant. Incidentally, density (Eq. 1.9) does not change significantly from a reference value, ρ_0 , and temporal as well as spatial fluctuations ($\dot{\rho}$) are much smaller, $\frac{|\dot{\rho}|}{\rho_0} \ll 1$.

$$\rho = \rho_0 + \dot{\rho}(x, y, z, t); \quad |\dot{\rho}| \ll \rho_0 \quad (1.9)$$

Eq. (1.8) can be rewritten (Eq. 10) and then Eq. (1.9) can be substituted for density (Eq. 11).

$$\frac{\partial\rho}{\partial t} + \nabla \cdot (\rho \mathbf{u}) = 0 \quad (1.10)$$

$$\frac{\partial(\rho_0+\dot{\rho})}{\partial t} + \left(\frac{\partial(\rho_0+\dot{\rho})u}{\partial x} + \frac{\partial(\rho_0+\dot{\rho})v}{\partial y} + \frac{\partial(\rho_0+\dot{\rho})w}{\partial z} \right) = 0 \quad (1.11)$$

Dividing Eq. (1.11) by ρ_0 and rearranging the terms

$$\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) + \frac{\dot{\rho}}{\rho_0} \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \right) + \frac{1}{\rho_0} \left(\frac{\partial \dot{\rho}}{\partial t} + u \frac{\partial \dot{\rho}}{\partial x} + v \frac{\partial \dot{\rho}}{\partial y} + w \frac{\partial \dot{\rho}}{\partial z} \right) = 0$$

Noting that $\dot{\rho}$ is significantly smaller than ρ_0 , the smaller terms of the equation can be neglected

$$\boxed{\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = \nabla \cdot \mathbf{u} = 0} \quad (1.12)$$

Eq. (1.12) is the continuity equation for an incompressible fluid and it states that mass is conserved because volume is conserved.

The Boussinesq momentum equations are found by substituting Eq. (1.9) into Eq. (1.5). Solving for the x-component:

$$(\rho_0+\dot{\rho}) \left(\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} \right) = -\frac{\partial p}{\partial x} + (\rho_0+\dot{\rho}) \mathfrak{T}_x$$

and in dividing by $(\rho_0+\dot{\rho})$, the horizontal momentum equations can be simplified by neglecting terms multiplied by $\dot{\rho}$ such that the x and y components become

$$\boxed{\frac{Du}{Dt} = -\frac{1}{\rho_0} \frac{\partial p}{\partial x} + \mathfrak{T}_x} \quad (1.13a)$$

$$\boxed{\frac{Dv}{Dt} = -\frac{1}{\rho_0} \frac{\partial p}{\partial y} + \mathfrak{T}_y} \quad (1.13b)$$

The vertical component can only be partially simplified if we consider a pressure field that is in hydrostatic balance (Eq. 1.6) with a constant density.

$$\begin{aligned} (\rho_0+\dot{\rho}) \left(\frac{Dw}{Dt} \right) &= -\frac{\partial p}{\partial z} - g(\rho_0+\dot{\rho}) + (\rho_0+\dot{\rho}) \mathfrak{T}_z \\ \frac{Dw}{Dt} &= -\frac{1}{\rho_0} \frac{\partial p}{\partial z} - g \frac{(\rho_0+\dot{\rho})}{\rho_0} + \mathfrak{T}_z \end{aligned}$$

Using $p = p_0 + \dot{p}$ and the hydrostatic balance as $\frac{\partial p_0}{\partial z} = -\rho_0 g$

$$\boxed{\frac{Dw}{Dt} = -\frac{1}{\rho_0} \frac{\partial \dot{p}}{\partial z} - \frac{\dot{\rho}}{\rho_0} g + \mathfrak{T}_z} \quad (1.13c)$$

Equations of motion on a rotating framework

The angular velocity of the Earth's rotation is $\Omega = \frac{2\pi}{\iota}$, where ι is the length of the sidereal day (86164 s).

$$\Omega = 7.29 \times 10^{-5} \text{ rad/s}$$

Two apparent forces are generated due to the Earth's rotation; centrifugal force and Coriolis force. In Fig. (2A), a fixed point in a system of coordinates is identified with a position vector $\mathbf{r} \equiv (x, y, z)$. The time rate of change for a vector $\mathbf{A}$ in a rotating system is seen in the inertial system as

$$\left(\frac{D\mathbf{A}}{Dt} \right)_{\text{in}} = \left(\frac{D\mathbf{A}}{Dt} \right)_{\text{rot}} + \boldsymbol{\Omega} \times \mathbf{A} \quad (1.14)$$

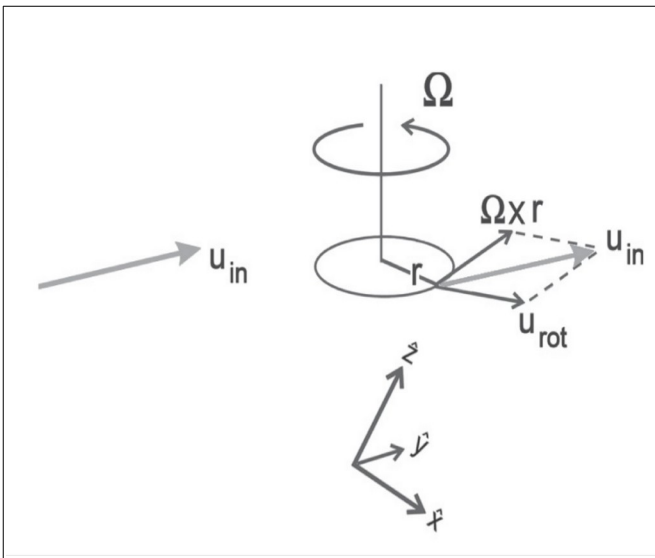


Figure 2A: Inertial (in) and rotating (rot) framework

Referring to Fig. (2A), using $\mathbf{r}$, the linear velocity of the point in the inertial framework is

$$\left(\frac{D\mathbf{r}}{Dt} \right)_{\text{in}} = \mathbf{u}_{\text{in}} = 0 + \boldsymbol{\Omega} \times \mathbf{r}$$

If the fixed point was moving with velocity $\mathbf{u}_{\text{rot}}$ as in Fig. (2A), then

$$\mathbf{u}_{\text{in}} = \mathbf{u}_{\text{rot}} + \boldsymbol{\Omega} \times \mathbf{r} \quad (1.15)$$

Computing the derivative of Eq. (1.15), noting that $\boldsymbol{\Omega}$ is a constant

$$\begin{aligned} \left(\frac{D\mathbf{u}_{\text{in}}}{Dt} \right)_{\text{in}} &= \left[\left(\frac{D}{Dt} \right)_{\text{rot}} + \boldsymbol{\Omega} \times \right] (\mathbf{u}_{\text{rot}} + \boldsymbol{\Omega} \times \mathbf{r}) \\ &= \left(\frac{D\mathbf{u}_{\text{rot}}}{Dt} \right)_{\text{rot}} + \boldsymbol{\Omega} \times \left(\frac{D\mathbf{r}}{Dt} \right)_{\text{rot}} + \boldsymbol{\Omega} \times \mathbf{u}_{\text{rot}} + \boldsymbol{\Omega} \times \boldsymbol{\Omega} \times \mathbf{r} \\ &= \left(\frac{D\mathbf{u}_{\text{rot}}}{Dt} \right)_{\text{rot}} + 2\boldsymbol{\Omega} \times \mathbf{u}_{\text{rot}} + \boldsymbol{\Omega} \times \boldsymbol{\Omega} \times \mathbf{r} \end{aligned}$$

The new total derivative function for equations of motion on a rotating framework can be written by combining the inertial solution (Eq. 1.5) with the above derivation

$$\frac{D\mathbf{u}}{Dt} = -\frac{\nabla p}{\rho} - g\hat{\mathbf{k}} - \underbrace{2\boldsymbol{\Omega} \times \mathbf{u}}_{\text{Coriolis}} - \underbrace{\boldsymbol{\Omega} \times \boldsymbol{\Omega} \times \mathbf{r}}_{\text{Centrifugal}} + \mathfrak{F} \quad (1.16)$$

However, Eq. (1.16) describes the motion of a fluid on a rotating disk. Consequently, the Coriolis force and centrifugal force must be translated onto a sphere. On a sphere $\boldsymbol{\Omega}$ has only two components, the N-S and vertical and is a function of latitude (θ): $\boldsymbol{\Omega} \equiv (0, \Omega \cos \theta, \Omega \sin \theta)$.

$$-2\boldsymbol{\Omega} \times \mathbf{u} = -2\Omega(w \cos \theta - v \sin \theta)\hat{\mathbf{i}} - 2\Omega u \sin \theta \hat{\mathbf{j}} + 2\Omega u \cos \theta \hat{\mathbf{k}}$$

Acknowledging that the vertical components are very small, and defining the Coriolis parameter as $f = 2\Omega \sin \theta$, the Coriolis force can be approximated

$$-2\boldsymbol{\Omega} \times \mathbf{u} \approx 2\Omega v \sin \theta \hat{\mathbf{i}} - 2\Omega u \sin \theta \hat{\mathbf{j}} + 0\hat{\mathbf{k}} = -f \hat{\mathbf{k}} \times \mathbf{u} \quad (1.17)$$

Gravity and the centrifugal force have adjusted the surface of the Earth to an equipotential surface – close to a reference ellipsoid. For centrifugal acceleration on a sphere, $\boldsymbol{\Omega} \times \boldsymbol{\Omega} \times \mathbf{r}$ is directed radially outwards with magnitude $-\Omega^2 r$ and r is the distance from the axis of the earth to the reference ellipsoid. Written in terms of the Earth's radius, $a = 6371 \text{ km}$ such that $r = (a+z)\cos \theta$, where z is the surface layer constituting the oceans and atmosphere.

Combining the centrifugal acceleration with gravity gives the resultant gravity and is measured in terms of gravitational potential (ϕ)

$$\phi = gz - \frac{\Omega^2 a^2 \cos^2 \theta}{2} \quad (1.18)$$

Rewriting Eq. (1.16) by substituting the Coriolis force (Eq. 1.17) and centrifugal force Eq. (1.18) for a sphere gives the equations of motion on a sphere

$$\boxed{\underbrace{\frac{D\mathbf{u}}{Dt}}_{\text{Total acceleration}} = - \underbrace{\frac{\nabla p}{\rho}}_{\text{Pressure gradient}} - \underbrace{-f \hat{\mathbf{k}} \times \mathbf{u}}_{\text{Coriolis}} - \underbrace{\nabla \phi}_{\text{Geopotential}} + \underbrace{\mathfrak{F}}_{\text{Friction}}} \quad (2.0)$$

Ekman transport

Friction in a moving fluid results from the transfer of momentum between different parts of the fluid. The physical process that transmits momentum is called viscosity (ν), which is seen as the resistance of a fluid to motion and has units of kg m s^{-1} . Stress (τ) applied to a surface will generate a momentum transfer and in a Newtonian fluid viscous stresses are proportional to the local shear in velocity.

Along the x-component, a stress is exerted, proportional to the velocity field

$$\vec{\tau}_u \propto \left(\frac{\partial u}{\partial x}, \frac{\partial u}{\partial y}, \frac{\partial u}{\partial z} \right)$$

Friction is proportional to the shear multiplied by the fluid viscosity.

$$\mathfrak{F}_x \propto \nu \nabla \cdot \vec{\tau}_u = \nu \left(\frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2} + \frac{\partial^2 u}{\partial z^2} \right)$$

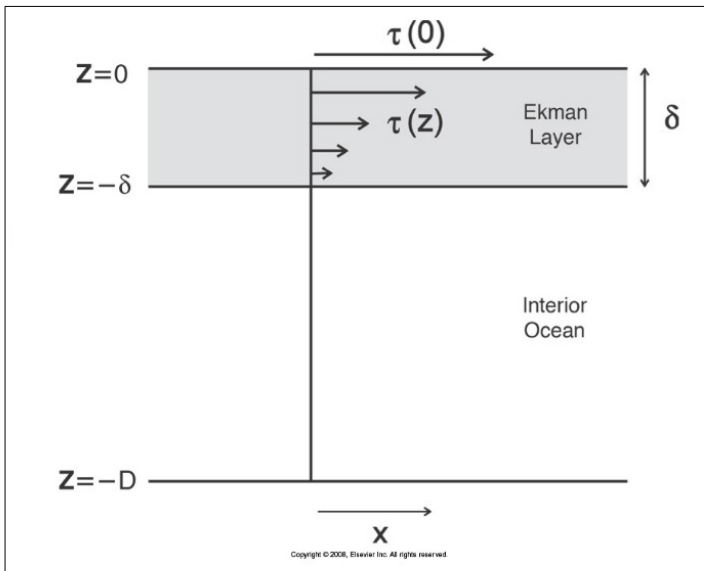
In resolving friction, the horizontal and vertical terms are separated. The molecular viscosity (A_H) refers to the horizontal diffusion of particles while eddy viscosity (A_v) is diffusion in the vertical.

$$\mathfrak{F}_x^{eddy} \propto \underbrace{A_H}_{\text{molecular viscosity}} \left(\frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2} \right) + \underbrace{A_v}_{\text{eddy viscosity}} \frac{\partial^2 u}{\partial z^2}$$

Eddy viscosity is significantly larger than molecular viscosity, allowing friction to be simplified, where $\mathbf{u}_H = (u, v)$. Consequently, friction can be described by Eq (2.1).

$$\mathfrak{F}^{eddy} \approx A_v \frac{\partial^2 \mathbf{u}_H}{\partial z^2} = \nu \nabla^2 \mathbf{u}_H \quad (2.1)$$

For a wind blowing over the ocean, the stress injected can be parameterized with a constant drag coefficient (c_D). Using the density of air (ρ_a) as well as the velocity of the wind measured at 10 m, the wind stress exerted on the surface layer of the ocean can be calculated with Eq. (2.2).



$$\tau = \rho_a c_D \mathbf{u}_H^2 \quad (2.2)$$

Ekman transport can be derived by making several assumptions:

- Laminar flow
- Surface ocean is in a steady state and velocity only varies in the vertical
- No horizontal pressure gradient
- The momentum flux at the surface (Fig. 3A) is equal to the wind-stress
- The momentum flux at a certain depth $-\delta$ is zero

Figure 3A: Wind-stress acting upon the surface layer (from Marshall & Plumb 2008)

$$\left(\nu \frac{\partial u}{\partial z} \right)_{z=0} = \frac{\tau_x}{\rho_0}; \quad \left(\nu \frac{\partial v}{\partial z} \right)_{z=0} = \frac{\tau_y}{\rho_0};$$

$$\left(v \frac{\partial u}{\partial z} \right)_{z=-\delta} = \left(v \frac{\partial v}{\partial z} \right)_{z=-\delta} = 0$$

Applying the hydrostatic balance (Eq. 1.6) as well as the Boussinesq approximations to the horizontal momentum components of Eq. (2.0) and using Eq. (2.1) to describe friction

$$\frac{D\mathbf{u}_H}{Dt} = -\frac{\nabla p}{\rho_0} - f \hat{\mathbf{k}} \times \mathbf{u}_H + \nu \nabla^2 \mathbf{u}_H \quad (2.3a)$$

Applying Ekman's assumptions to Eq. (2.3a), the equation simplifies to a balance between Coriolis force and friction.

$$f \hat{\mathbf{k}} \times \mathbf{u}_H = \nu \frac{\partial^2 \mathbf{u}_H}{\partial z^2} \quad (2.3b)$$

Rewriting Eq. (2.3b), acknowledging that ν can be substituted for A_v , the horizontal components can be written as

$$-f v = A_v \frac{\partial^2 u}{\partial z^2}$$

$$f u = A_v \frac{\partial^2 v}{\partial z^2}$$

The Ekman transport (M_E) is the vertical integral of the velocity in the Ekman layer (Fig. 3A), which is the depth at which the wind-stress has no impact on the fluid motion and has a thickness of D_E .

$$M_E^y = \int_{D_E} v(z) dz; \quad M_E^x = \int_{D_E} u(z) dz$$

Using the boundary condition assumptions and integrating the horizontal components

$$f \hat{\mathbf{k}} \times \int_0^{-\delta} \mathbf{u}_H(z) dz = A_v \int_0^{-\delta} \frac{\partial^2 \mathbf{u}_H}{\partial z^2} dz$$

Ekman transport becomes

$$\boxed{f \hat{\mathbf{k}} \times \mathbf{M}_E = \frac{\boldsymbol{\tau}}{\rho_0}} \quad (2.4)$$

Geostrophic balance

Conducting a scaling analysis of Eq. (2.0) relative to the Rosby number (R_o)

$$R_o = \frac{U}{fL}$$

where U is the magnitude of the horizontal velocities and L is the length scale of the horizontal components, shows that the major balance in the horizontal components of Eq. (2.0) is achieved between the Coriolis force and pressure-gradient force. Using the horizontal Boussinesq momentum equations (Eq. 1.13a,b) and considering only the major terms, the geostrophic balance can be derived (Eq. 2.5).

$$f \hat{\mathbf{k}} \times \mathbf{u}_H + \frac{\nabla_H p}{\rho_0} = 0 \quad (2.5)$$

Thermal wind relation

A horizontal gradient in density will give rise to a vertical shear in $\mathbf{u}_H$. Differentiating Eq. (2.5) with respect to depth and associating the geostrophic currents as $\mathbf{u}_g$ gives

$$\frac{\partial \mathbf{u}_g}{\partial z} = \frac{1}{f \rho_0} \hat{\mathbf{k}} \times \frac{\nabla_H p}{\partial z}$$

Using Eq. (1.6), $\frac{\nabla_H p}{dz}$ can be eliminated for $-g \nabla_H \rho$ such that

$$\frac{\partial \mathbf{u}_g}{\partial z} = -\frac{g}{\rho_0 f} \hat{\mathbf{k}} \times \nabla_H \rho \quad (2.6)$$

Potential Vorticity

Vorticity is the tendency for a particle to rotate. In the ocean, angular momentum of water parcels must be conserved such that vorticity is conserved. The relative vorticity (ζ) of a parcel is described by the vertical component of the curl of the velocity field $\nabla \times \mathbf{u}$

$$\zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \quad (2.7)$$

Planetary vorticity is described by the Coriolis parameter, f , such that the absolute vorticity experienced by a parcel of water is the summation of the relative and planetary vorticity. Consequently, absolute vorticity is the angular momentum relative to the inertial framework that must be conserved.

Starting with Eq. (2.0), we assume that we are away from frictional boundaries and thus friction can be neglected such that

$$\frac{D\mathbf{u}}{Dt} + f \hat{\mathbf{k}} \times \mathbf{u} = -\frac{\nabla p}{\rho}$$

Taking the curl (Eq. 2.7) of the horizontal momentum equations will yield the relative vorticity. Starting with the v-momentum equations

$$\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + f u = -\frac{1}{\rho} \frac{\partial p}{\partial y}$$

and differentiating along x

$$\begin{aligned} \frac{\partial}{\partial t} \left(\frac{\partial v}{\partial x} \right) + \left[\frac{\partial u}{\partial x} \frac{\partial v}{\partial x} + u \frac{\partial}{\partial x} \left(\frac{\partial v}{\partial x} \right) \right] + \left[\frac{\partial v}{\partial x} \frac{\partial v}{\partial y} + v \frac{\partial}{\partial y} \left(\frac{\partial v}{\partial x} \right) \right] + \left[\frac{\partial w}{\partial x} \frac{\partial v}{\partial z} + w \frac{\partial}{\partial z} \left(\frac{\partial v}{\partial x} \right) \right] + u \frac{\partial f}{\partial x} + f \frac{\partial f}{\partial x} \\ = -\frac{1}{\rho} \frac{\partial^2 p}{\partial x \partial y} \end{aligned}$$

$u \frac{\partial f}{\partial x} = 0$ and rearranging the remaining terms gives

$$\frac{\partial v}{\partial x} \left(\frac{\partial}{\partial t} + u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} + w \frac{\partial}{\partial z} \right) + \frac{\partial u}{\partial x} \frac{\partial v}{\partial y} + \frac{\partial v}{\partial x} \frac{\partial v}{\partial y} + \frac{\partial w}{\partial x} \frac{\partial v}{\partial z} + f \frac{\partial u}{\partial x} = -\frac{1}{\rho} \frac{\partial^2 p}{\partial x \partial y}$$

Similarly for the u-momentum equations along y

$$\frac{\partial u}{\partial y} \left(\frac{\partial}{\partial t} + u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} + w \frac{\partial}{\partial z} \right) + \frac{\partial u}{\partial y} \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \frac{\partial u}{\partial y} + \frac{\partial w}{\partial y} \frac{\partial u}{\partial z} - f \frac{\partial v}{\partial y} - v \frac{\partial f}{\partial y} = -\frac{1}{\rho} \frac{\partial^2 p}{\partial x \partial y}$$

Taking the difference and substituting the total derivative (Eq. 1.0)

$$\frac{D}{Dt} \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) + \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) + \frac{\partial w}{\partial x} \frac{\partial v}{\partial z} - \frac{\partial w}{\partial y} \frac{\partial u}{\partial z} + f \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) + v \frac{\partial f}{\partial y}$$

$\frac{\partial w}{\partial x} \frac{\partial v}{\partial z} - \frac{\partial w}{\partial y} \frac{\partial u}{\partial z}$ is negligible and substituting for ζ (Eq. 2.7) gives

$$\frac{D\zeta}{Dt} + \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) (\zeta + f) + v \frac{\partial f}{\partial y} = 0$$

Acknowledging that $\frac{Df}{Dt} = \frac{\partial f}{\partial t} + u \frac{\partial f}{\partial x} + v \frac{\partial f}{\partial y} + w \frac{\partial f}{\partial z} = 0$ and $\frac{\partial f}{\partial t}; u \frac{\partial f}{\partial x}; w \frac{\partial f}{\partial z} = 0$, the total derivative of Coriolis can be simplified to

$$\frac{Df}{Dt} = v \frac{\partial f}{\partial y}$$

Substituting this into the derivation gives

$$\frac{D(\zeta+f)}{Dt} + \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) (\zeta+f) = 0$$

Using the continuity equation for an incompressible fluid (Eq. 1.12)

$$\frac{D(\zeta+f)}{Dt} - \frac{\partial w}{\partial z} (\zeta+f) = 0$$

Integrating Eq. (1.12) over the thickness of a water column H and letting $w = \frac{dz}{dt}$ as well as assuming the height of the surface layer remains constant in time

$$\int_0^H \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) dz = - \int_0^H \frac{\partial w}{\partial z} dz$$

Substituting for w

$$\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) H = - \int_0^H \frac{\partial}{\partial z} \left(\frac{\partial z}{\partial t} \right) dz$$

and simplifying

$$\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) = - \frac{1}{H} \frac{dH}{dt}$$

Substituting back into the derivation

$$\frac{D(\zeta+f)}{Dt} - \left(\frac{1}{H} \right) (\zeta+f) \frac{\partial H}{\partial t} = 0$$

Assuming H only varies in time

$$\frac{D(\zeta+f)}{Dt} - \left(\frac{1}{H} \right) (\zeta+f) \frac{DH}{Dt} = 0$$

This can be rewritten as

$$\frac{D}{Dt} \left(\frac{\zeta+f}{H} \right) = 0$$

If the Potential Vorticity (PV) in a system must remain constant $\frac{D}{Dt}(PV) = 0$ then

$$\boxed{PV = \frac{\zeta+f}{H}} \quad (2.8)$$

Sverdrup Relation

Assuming a two layered ocean, where the upper column of water is defined as the Ekman layer (Fig. 3A), wind-stress curls give rise to patterns of Ekman pumping and suction. This results in a change in the height of the upper column. Consequently, to conserve potential vorticity in the interior water column, a meridional displacement occurs. Incidentally, the Sverdrup transport links the transport of the deep ocean to the surface wind stress curl.

Within the Ekman layer, we assume no friction and no acceleration such that the horizontal Boussinesq momentum equations (Eq. 13a,b) simplify to the geostrophic balance (Eq. 2.5).

$$f \hat{\mathbf{k}} \times \mathbf{u}_H = -\frac{\nabla_H p}{\rho_0}$$

Taking the curl (Eq. 2.7) of the geostrophic balance gives

$$\frac{\partial}{\partial x}(fu) - \frac{\partial}{\partial y}(-fv) = -\frac{1}{\rho_0} \left(\frac{\partial}{\partial x} \frac{\partial p}{\partial y} - \frac{\partial}{\partial y} \frac{\partial p}{\partial x} \right)$$

As $\frac{\partial f}{\partial x} u = 0$ and using $\beta = \frac{df}{dy}$, coming from the β -plane approximation (not shown)

$$f \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) + \beta v = 0$$

Using the continuity equation for an incompressible fluid (Eq. 1.12) and rewriting the equation

$$f \left(\frac{\partial w}{\partial z} \right) = \beta v$$

Integrating from the base of the Ekman layer to the bottom of the ocean (D) (Fig. 3A), assuming $w_D = 0$ and where $V = \int_D v dz$

$$f \int_D^{De} \frac{\partial w}{\partial z} dz = \beta \int_D^{De} v dz$$

$$f w_{De} = \beta V \quad (2.9)$$

Noting that the Ekman layer is defined as the balance between the Coriolis force and friction, taking the curl of the horizontal Ekman balance (Eq. 2.3b), noting that friction is proportional to wind-stress (Eq. 2.2)

$$\frac{\partial}{\partial x}(fu) - \frac{\partial}{\partial y}(-fv) = \frac{1}{\rho_0} \left(\frac{\partial}{\partial x} \frac{\partial \tau_y}{\partial z} - \frac{\partial}{\partial y} \frac{\partial \tau_x}{\partial z} \right)$$

Using Eq. (1.12) as well as β and rearranging the terms

$$f \left(\frac{-\partial w}{\partial z} \right) + \beta v = \frac{1}{\rho_0} \frac{\partial}{\partial z} (\hat{\mathbf{k}} \cdot \nabla \times \boldsymbol{\tau})$$

Integrating from the surface to the base of the Ekman layer, assuming $w_0 = 0$

$$-f \int_{De}^0 \frac{\partial w}{\partial z} dz + \beta \int_{De}^0 v dz = \frac{1}{\rho_0} \int_{De}^0 (\hat{\mathbf{k}} \cdot \nabla \times \boldsymbol{\tau}) \frac{\partial}{\partial z} dz$$

Noting that V is equal to the meridional component of Eq. (2.4)

$$f w_{De} + \beta \left(\frac{-\tau_x}{\rho_0 f} \right) = \frac{\hat{\mathbf{k}} \cdot \nabla \times \boldsymbol{\tau}}{\rho_0}$$

Making w_{De} the subject of the equation and seeing that $\beta \left(\frac{\tau_x}{\rho_0 f^2} \right) \approx 0$

$$w_{De} = \frac{\hat{\mathbf{k}} \cdot \nabla \times \boldsymbol{\tau}}{\rho_0 f} \quad (2.10)$$

Combining Eq. (2.9) and Eq. (2.10) gives the Sverdrup relation

$$\boxed{\beta V = \frac{\hat{\mathbf{k}} \cdot \nabla \times \boldsymbol{\tau}}{\rho_0}} \quad (2.11)$$

Thank you very much
for your help Jono xx

It is much appreciated ;)

- Sadiyah



Thank you so much Jono
We really appreciate all the help

- Ash and S

Thank you,
see you next
time ☺

Raguel

Thank you very very much
for all your help Jono.
I really appreciate it ;)

- S

Thanks again
Jono for the
lift ;)

MEOW



ASH

Hope to see you
some day!

Best,
Annette