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The variability of Lagrangian transport in the southern Benguela Current upwelling system

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Par

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**The variability of Lagrangian transport in the southern Benguela
Current upwelling system.**

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Abstract

This study analyses the physical mechanisms that impact Lagrangian pathways and transport in the southern Benguela upwelling system (SBUS), an environment in which currents are key components of many important ecological processes, including the dispersal of marine larvae. Physical advection by currents is an important mechanism for egg and larvae transport success in the SBUS since the spawning areas and recruitment areas are separated by a long distance. Numerical model simulations of the SBUS coupled with particle tracking experiments are used to investigate Lagrangian pathways between the Cape Peninsula (34°S) and St Helena Bay (32°S) and how they are linked to the oceanic circulation. Transport success, given by the ratio of the number of particles that reach St Helena Bay (32°) over the total number of particles released, is used to quantify the along-shore connectivity between the two regions.

Results of an interannual eddy-resolving (7 km) Coastal and Regional Ocean Community (CROCO) model simulation, coupled with a Lagrangian particle-tracking model (Pyticles) reveal a strong seasonal cycle in transport success that is governed by the wind-induced Ekman drift as well as the complex three-dimensional structure of the alongshore jets and their variability. An efficient and fast inshore route taken by particles in spring (September-November). This period is the optimal window, as the jet is in its intensified state, close to the shore, and relatively stable. At this time of the year, offshore Ekman transport is still low enough to allow particles to remain on the inshore side of the jet. During the peak upwelling season, this inshore path becomes less efficient due to high offshore Ekman transport and also the Benguela jet splits into multiple jets resulting in inshore and offshore routes. Particles advected in the offshore branch are exposed to

intense mesoscale variability and therefore are susceptible to chaotic trajectories, longer connection time and lower transport success.

The sensitivity of the alongshore connectivity to increasing ocean model horizontal resolution was investigated by developing a nested simulation based on a 3 km resolution grid. Increasing the model resolution from 7 km to 3 km reveals some changes in the intensity and the structure of the coastal circulation including the large-scale circulation associated with the Agulhas Current leakage. The increased resolution allows finer details of the coastal topography to be represented and this led to a Benguela jet that is narrower and more coherent (less diffused). Moreover, the 3 km simulation resolves a nearshore poleward current that is not present in the lower resolution simulation. A more realistic topography also led to a reduced leakage of warm Agulhas Current waters and instabilities associated with it. Our results however show that at a seasonal time-scale, there is not much difference in the alongshore connectivity since the Benguela jet that is responsible for this connectivity is present in both simulations.

The 3 km simulation is used to investigate the role of wind-driven coastal circulation and mesoscale variability on the interannual variability of transport success. This analysis reveals that the variability of transport success cannot be attributed to a single physical driver. The Benguela jet is found to be the dominant driver of transport success variability for particles released on the shelf, while the removal of these particles occurs through offshore Ekman transport and the nearshore poleward current could contribute to negative anomalies of transport success. Cross-shore transport is shown to contribute to the variability in transport success through the activity of mesoscale eddies. An infrequent interaction of a passing Agulhas ring with the shelf led to a strong offshore advection of particles into the open ocean (a zone of highest EKE) and negative transport success. The influence of an Agulhas ring is dependent on two factors: (i) the duration of its interaction, and (ii) its distance from the shelf-edge and the origin of particles. A shorter and closer interaction tends to have a greater impact. There were only 18 interactions that occurred over the 20 years model simulation and 13 of them resulted in negative transport success

anomaly of particles that originated on the outer shelf-edge. In contrast, the shelf-edge cyclonic eddies favour the onshore transport of particles that originated on the outer shelf-edge and thus a positive anomaly transport success. This onshore transport mostly occurs in an event when these particles do not initially experience any offshore transport away from the shelf-edge by anti-cyclonic eddies.

The findings of this study will provide a valuable information for the future studies on the role of the physical drivers that impact transport of larvae and eggs.

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General Introduction

Eastern Boundary coastal Upwelling Systems (EBUS) are some of the most productive marine systems. They cover less than 1% of the world's ocean surface but supports 20% of the marine fish catch (*Alves and Miranda, 2013*). The Benguela Current Large Marine Ecosystem (BCLME), situated off the west and south-west coasts of southern Africa, is one of the four major EBUS of the world's ocean (*Hill, 1998*). It extends from the Angola-Benguela Frontal Zone (ABFZ) situated between 15-17°S to 37°S in latitude and from 0°E to 20°E in longitude. This study focuses on the Southern Benguela Upwelling System (SBUS), a sub-region within the BCLME generally defined as bounded to the north by the Lüderitz cell (26.6°S), the largest and most perennial upwelling cell (*Shannon and Nelson, 1996; Weeks et al., 2006*), and to the south by the warm Agulhas Current.

Equatorward along-shore winds blow predominantly in summer along the west coast of South Africa, inducing an offshore transport of the coastal surface waters through Ekman dynamics. The no-inflow condition at the coast creates a divergence of the offshore Ekman transport that induces a vertical velocity directed towards the surface. This upwelling of deep, cold and nutrient-rich waters enriches the euphotic zone with nutrients, which stimulates and sustains a high primary production and a rich ecosystem and provides a suitable environment for small pelagic fish. Catches of sardines and anchovy are abundant in the region which drives an essential socio-economic activity.

In the EBUS, the abundance of small pelagic fish populations is regulated by a combination of several physical and biological processes (*Bakun, 1996*). Since 1983, repeated hydro-acoustic surveys have been carried out in the SBUS to monitor the populations

25 of sardines and anchovies. Their analysis shows significant interannual variability in the
 26 anchovy abundance and catches (*van der Lingen et al., 2002; Huggett et al., 2003*). Subse-
 27 quently, extensive research has been undertaken focusing on the early life history of small
 28 pelagic fish and on how this variability could be linked to changes in the environmental
 29 conditions (*Boyd et al., 1998; Hutchings et al., 1998; Huggett et al., 2003; Mullon et al.,*
 30 *2003; Parada et al., 2003; Lett et al., 2006; Miller et al., 2006; Parada et al., 2008; Koné*
et al., 2013).

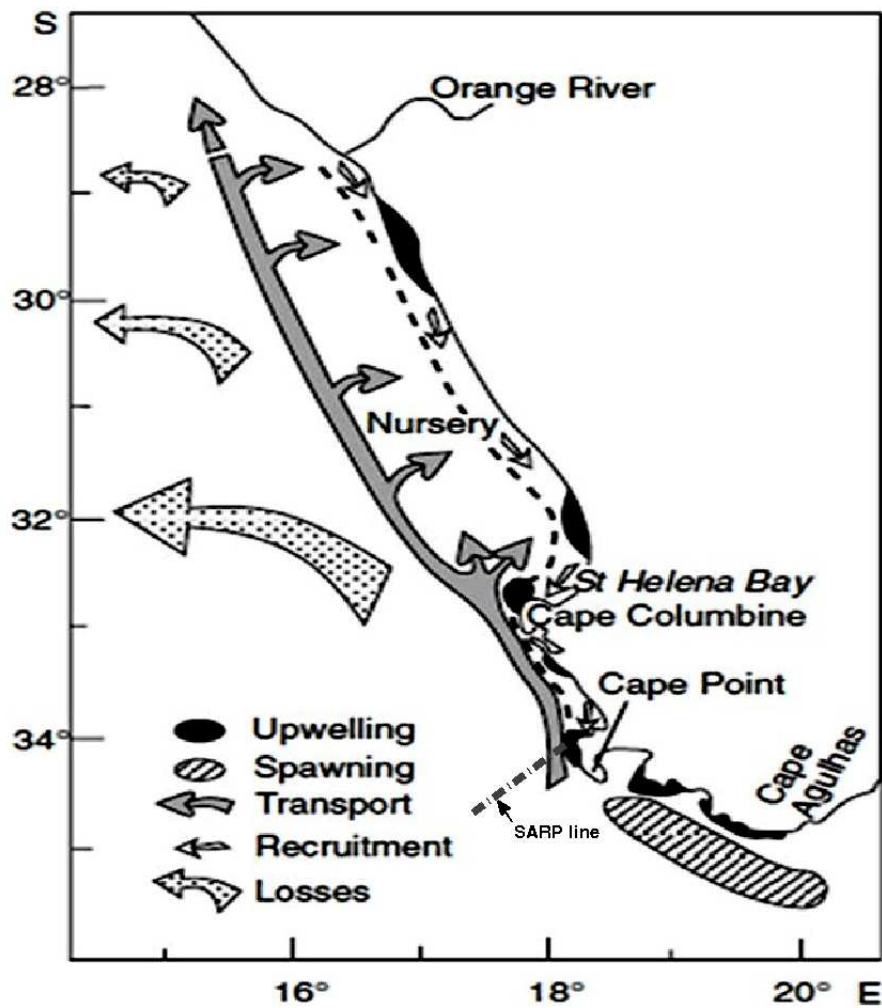


Figure 1: A conceptual diagram of the life history of anchovy in the southern Benguela. From *Hutchings et al. (1998)*. Fish eggs, spawned at the tip of Cape Agulhas are advected northward alongshore. They experience offshore losses along the way. Those that make it to the nursery area (St Helena Bay) are considered as recruits. In their early-stage, anchovy larvae have minimal swimming ability, which makes them behave as passive particles. Therefore, their dispersion and transport is directly dependant on the oceanic circulation (*Pagès et al., 1991*).

31

32 Anchovy eggs are spawned during the spring and summer season, in the western Agul-

33 has Bank (WAB), a 200 m depth large coastal plateau that extends up to 400 km offshore
34 at the south tip of the South African continent (*Hutchings, 1992; Hutchings et al., 1998;*
35 *Huggett et al., 2003*) (Fig. 1). Since 1997, an eastward extension of the spawning ground
36 has been observed at the eastern side of the bank (*Roy et al., 2007*) but with the west coast
37 being the main nursery. In order to survive, eggs and larvae must be transported from
38 the spawning to the nursery grounds in less than 30 days. Nevertheless, in the SBUS,
39 the latter is situated more than 400 km away, off the southern African west coast, be-
40 tween Cape Columbine and the Orange River. This makes the SBUS distinctive from its
41 northern Benguela counterpart and the other major Eastern Boundary Upwelling Systems
42 (EBUS) where spawning and nursery grounds overlap. The success of transport to the
43 nursery ground is considered to be critical for the anchovy stock, and one of the main
44 drivers of the recruitment variability of anchovy (*Shelton and Hutchings, 1982; Hutchings,*
45 *1992; Hutchings et al., 1998*). However, in their early-stage, anchovy larvae have minimal
46 swimming ability, which makes them behave as passive particles. Therefore, their disper-
47 sion and transport, from the Agulhas Bank to the nursery area, are directly dependant on
48 the oceanic circulation (*Pagès et al., 1991*).

49

50 *Hutchings (1992)* presented a schematic picture of the major processes impacting the
51 eggs and larvae journey from the western Agulhas Bank to the west coast upwelling region
52 (Fig. 1). This scheme was based on knowledge of the coastal circulation and numer-
53 ous observations of eggs and larvae distribution (*Bang and Andrews, 1974; Shelton and*
54 *Hutchings, 1982; Shannon, 1985; Armstrong et al., 1987; Shelton and Hutchings, 1990*).
55 A coastal current flowing northward and following the shelf, namely the Benguela Jet,
56 is considered as the major conveyor belt linking the western Agulhas Bank and the west
57 coast upwelling area. According to *Nelson (1989)*, the Benguela Jet is a narrow current
58 (~ 20 km) that flows northward along the South African west coast between Cape Point
59 (34.15°S , 18.4°E) and Cape Columbine (33°S , 17.5°E). The jet can sustain high velocities,
60 ranging from 25 cm s^{-1} to 75 cm s^{-1} (*Bang and Andrews, 1974; Gordon, 1985*). It lies
61 over a poleward undercurrent (*Nelson and Hutchings, 1983*), that flows over the shelf edge
62 and occasionally reaches the surface. The jet varies in position and strength and responds

63 rapidly to changes in wind stress magnitude and direction. It follows the seasonal cycle of
64 the upwelling (*Armstrong et al.*, 1987) and intensifies in summer when coastal upwelling
65 of cold waters contributes to reinforcing the across-shore density gradient (*Veitch et al.*,
66 2006, 2017). In winter, when the wind direction reverses, this across-shore density gra-
67 dient weakens but persists due to the intrusions of warm Agulhas waters from the south
68 (*Twatwa Mhlongo et al.*, 2005).

69

70 *Hutchings'* (1992) scheme for Lagrangian transports of fish eggs and larvae has been
71 widely used to set-up numerous Lagrangian modelling studies in the Southern Benguela
72 (*Mullon et al.*, 2003). *Huggett et al.* (2003), followed by *Parada et al.* (2003; 2008), carried
73 out pioneering work demonstrating that the dominant anchovy spawning patterns in the
74 Southern Benguela could be reproduced when combining a 3D hydrodynamic model with
75 a Lagrangian particle-tracking tool. They also quantified the impact, on eggs and larvae
76 transport success, of parameters such as the initial location, density and patchiness of
77 eggs, as well as behavioural processes such as diurnal vertical migration, mortality due to
78 lethal temperatures or predation. However, neither the local pathways nor the physical
79 processes responsible for the transport of particles was studied in detail, even though the
80 success of their experiments was obviously linked to the capacity of the hydrodynamical
81 model to reproduce the main characteristics of the currents in the region.

82

83 The modelling works of *Veitch et al.* (2006) and *Blanke et al.* (2009) provided insights
84 on the importance of the Benguela Jet for the connectivity between the western Agulhas
85 Bank and the west coast upwelling area. More recently, in a detailed modelling study of
86 the shelf edge currents, *Veitch et al.* (2017) described the bifurcation of this jet between
87 Cape Point and Cape Columbine into a weak, narrow branch flowing over the shelf edge
88 and a strong offshore branch flowing in the north-westward direction. A bifurcation of
89 the jet, as well as other physical processes such as induced offshore Ekman transport and
90 occasional entrainment by Agulhas rings, have been suggested as important physical pro-
91 cesses impacting the alongshore transit of fish eggs and larvae (*Hutchings et al.*, 1998;
92 *Skogen et al.*, 2003; *Veitch et al.*, 2006; *Garavelli et al.*, 2012).

94 This thesis builds upon these recent physical modelling studies to investigate in de-
95 tails how the horizontal and vertical structure of the shelf edge jets and the presence of
96 mesoscale turbulence impact the alongshore trajectory of passive particles in the SBUS.
97 The approach consists in a set of Lagrangian particle tracking experiments based on a
98 robust hydrodynamical model capable of simulating the dominant physical processes in-
99 fluencing the along-shore transport of particles, *i.e.* a realistic seasonal coastal circulation,
100 a realistic mesoscale eddy field, Agulhas leakage bringing warm Agulhas waters offshore
101 of the southern Benguela shelf. While our study remains motivated by questions related
102 to the dynamics of anchovy reproductive patterns and recruitment variability, it distin-
103 guishes from the others as it focuses on the physical transport processes that impact
104 the survival of individuals at early stages. We are convinced that an important step for
105 improving our understanding of the functioning of marine ecosystems is first to gain a de-
106 tailed understanding of the physical processes that impact the life cycle. We believe that
107 this understanding is of primary importance before adding other biological compartments,
108 such as predation and mortality.

109

110 The Lagrangian experiments are built upon the pioneering work of *Huggett et al.* (2003)
111 to focus exclusively on the northward alongshore portion of the hypothesised trajectory of
112 anchovy larvae from the Agulhas Bank to St Helena Bay, *i.e.* the portion between Cape
113 Point and St Helena Bay. If the Benguela Jet is assumed to play a dominant role in the
114 transport of fish larvae, the success of this route should be in theory more predictable than
115 the cross-shelf transport in the vicinity of the turbulent Agulhas retroflexion. Compared
116 to the model used by *Huggett et al.* (2003) and *Parada et al.* (2003, 2008), the simulations
117 used for this study were created using a more recent version of Regional Ocean Modelling
118 System (ROMS) and a regional configuration with higher vertical and horizontal resolu-
119 tions to allow an accurate representation of the mesoscale and sub-mesoscales structures
120 generated by the Benguela Jet instabilities, as well as the inner shelf (less than 100 m
121 depth) poleward jets.

122

123 This study is dedicated to assessing the role of locally generated physical processes on
124 the along-shore pathways. Our analysis is focused on the physical processes related to (i)
125 the wind surface forcing (Ekman drift, Benguela Jet, coastal poleward current) and (ii) the
126 role of the mesoscale dynamics at play in the area. Remotely forced ocean physics, such as
127 southward propagating coastal Kelvin waves, that have been demonstrated to be of minor
128 importance in the southern Benguela (*Illig et al.*, 2018) or large scale interannual variabil-
129 ity of Agulhas leakage into the South Atlantic Ocean (*Loveday et al.*, 2014) is neglected.
130 On the other hand, the full spectrum of atmospheric forcing variability is considered in
131 order to correctly represent the locally generated physical processes (upwelling and Ek-
132 man dynamics forced by the local wind) and its interaction with the intense mesoscale
133 dynamics at play in this region. Keeping the full variability of the atmospheric forcing to
134 run a 20-year simulation allows deriving a much more realistic seasonal cycle than if an
135 average surface forcing from a climatology was used. This is a major difference between
136 our set-up and those used by studies published in the early 2000s. In our set-up, the local
137 wind-forced physical processes (such as Ekman drift, upwelling cells, poleward undercur-
138 rent) are different and much more realistic.

139

140 The outline of the thesis is as follows: Chapter 1 presents a literature review of the
141 dynamics of the SBUS. A brief description of the circulation in the SBUS is given, focusing
142 on the main oceanographic and atmospheric features that may influence the transport of
143 anchovy larvae during their early stages. Chapter 2 provides a detailed description of
144 the numerical model used for this study and Chapter 3 describe the reference numerical
145 simulation designed, and its validation. Our reference simulation is based on a 7.5 km
146 horizontal resolution grid, with 100 vertical levels, and it encompasses all the Benguela
147 Upwelling System, the Southern Ocean up to 54°S and the Mozambique Channel. Unlike
148 the climatologically forced model of *Veitch et al.* (2017), our ocean model is forced by
149 6-hourly surface forcing (wind, heat and freshwater fluxes) over the 1989-2010 period.
150 A careful assessment of this reference simulation against available in-situ and satellite
151 datasets is presented.

152 Chapter 4 focuses on the physical processes that influence the seasonal cycle of the

153 alongshore connectivity between the southernmost portion of the SBUS (Cape Point at
154 34°S) and St. Helena Bay (32°S). The Lagrangian experiments aim at addressing the
155 following: (1) the identification of the dominant pathways that ensure this connectivity;
156 (2) their characteristics and robustness, and (3) the ocean dynamics that are responsible
157 for the seasonal alongshore transport success variability. Results are based on the 7.5 km
158 horizontal resolution reference simulation. The details of the Lagrangian experimental
159 set-up used in the subsequent chapters are also presented here.

160 Chapter 5 discusses the sensitivity of these results to the resolution of the hydrody-
161 namical model. For this purpose, one nested simulation based on 3 km resolution grid is
162 developed and also run for 20 years. Increasing the spatial resolution allows some fine-scale
163 shelf dynamical features, such as the very inshore poleward coastal jet, to be modelled.
164 Furthermore, the Agulhas leakage was also unexpectedly changed in the higher resolution
165 model, with some important implications on the stratification of the region. These gains
166 of realism have an impact on the alongshore connectivity, in particular on its variability.
167 However, they confirm the significant role of the Benguela Jet in modulating the seasonal
168 cycle of the alongshore connectivity.

169 Having identified the mechanisms that drive the seasonal cycle allows us to investigate
170 how the interannual variability of each of these processes individually impact the interan-
171 nual variation of transport success. This is the focus of Chapter 6. This study is based on
172 the 3 km simulation nested within the 7.5 km simulation. The 7.5 km simulation considers
173 a monthly climatology at lateral boundaries, which filters out the interannual signal, but
174 keeps the seasonal one. It is important to stress that the objective of this work is not to
175 study the impact of remotely generated interannual variability, but rather to identify the
176 significant local dynamical processes influencing the seasonal cycle of transport between
177 Cape Point and St Helena Bay, i.e. a realistic Benguela Jet, coastal poleward current,
178 offshore Ekman drift, and off-shelf eddy field.

179 Finally, conclusions and future directions of this research are discussed in Chapter 7.
180 A better understanding of the processes controlling Lagrangian transport dynamics in the
181 SBUS allows for better connections to be made between the physics and the biology of
182 both lower (primary productivity) and middle trophic-levels (pelagic fish eggs and larvae).

183 It also enlightens some potential implications for the understanding of the propagation of
184 harmful algal blooms and/or oil spills and pollutants.

185 Chapter 1

186 Literature Review

187 This chapter presents a literature review of the current knowledge on the ocean circulation
188 of the southern Benguela upwelling system (SBUS). The literature review will first focus
189 on the atmospheric forcing that drives the upwelling dynamics and then the main regional
190 circulation.

191 1.1 Atmospheric forcing

192 1.1.1 Seasonal variability

193 The primary atmospheric forcing of the southern Benguela upwelling system is the semi-
194 permanent South Atlantic Anticyclone (SAA), the presence of continental low-pressure
195 systems and the eastward propagating mid-latitude cyclones south of the continent at
196 2-8 day periods (*Risien et al.*, 2004; *Shillington et al.*, 2006; *Burls and Reason*, 2008).
197 The SAA, denoted by the 150 m isoline in Figure 1.1, undergoes a seasonal latitudinal
198 migration. It moves southward (northward) during austral summer (winter). It also shifts
199 longitudinally such that it is found closer to the coast in spring (Fig 1.1d) and moves
200 offshore in summer and autumn (Fig 1.1a,b). This migration results in the variability
201 of the upwelling favourable winds in the region between the 25°S-35°S. The presence of
202 a continental low-pressure trough over the south of the African continent results in the
203 strengthening of the alongshore equatorward winds in summer. On the other hand, the
204 presence of the continental high and the northward movement of SAA in winter weakens

205 the upwelling-favourable winds and results in the dominance of westerlies over the region
 206 (Fig 1.1c) (*Shelton et al.*, 1985). As a result, the upwelling-favourable winds reach a
 207 maximum during summer and downwelling favourable winds occur in winter with spring
 208 and autumn being transition periods.

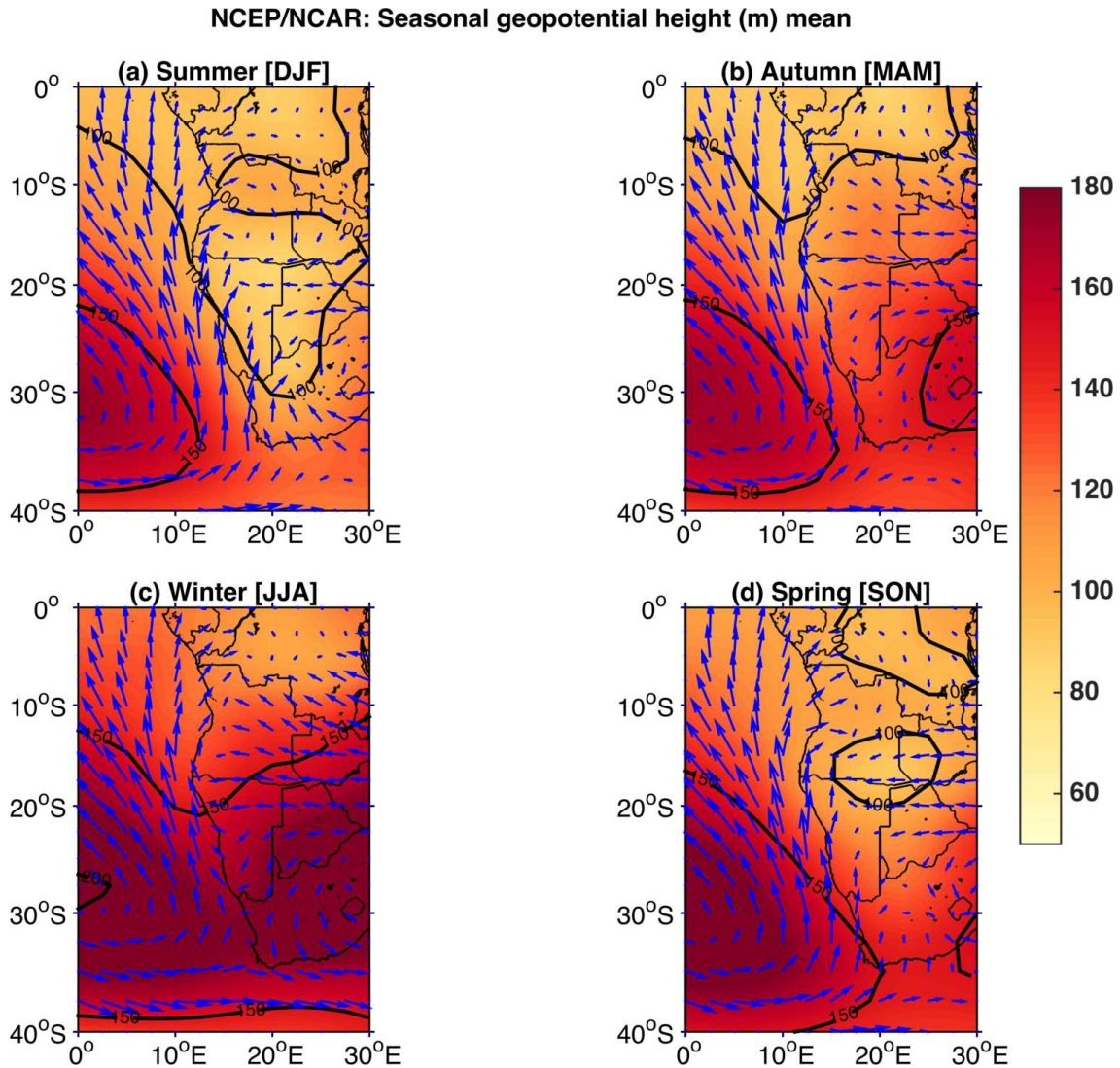


Figure 1.1: Mean seasonal composite of 1000 mb geopotential height (m) over the South Atlantic ocean from NCEP (1980-2012) data: (a) summer (December-February), (b) autumn (March-May), (c) winter (June-August) and (d) spring (September-November). The contours shown are of the 100 m and 150 m geopotential height and represent the high and low pressure, respectively. The surface wind speed and direction is represented by the overlaid blue arrows. Data used was downloaded from NCEP website¹

¹<https://www.esrl.noaa.gov/psd/cgi-bin/data/composites/printpage.pl>

209 1.1.2 Intra-seasonal and short-term variability

210 The large-scale seasonal variability of the atmospheric forcing over the SBUS region co-
211 exists with some short-term and intra-seasonal atmospheric features. The equatorward
212 upwelling favourable winds that blow predominantly during summer are subject to sub-
213 stantial intra-seasonal variability. Synoptic features such as the passage of cold fronts,
214 cut-off lows and mesoscale features such as the Berg winds², coastal lows and diurnal sea
215 breeze (*Shillington et al.*, 2006) can disrupt the favourable upwelling winds at different
216 temporal and spatial scale (*Shillington et al.*, 2006) and can promote specific upwelling or
217 downwelling events (*Fennel*, 1999).

218 The eastward passage of cyclonic storms south of South Africa, may generate winds with
219 poleward and onshore components driving downwelling events (Fig. 1.2) (*Bakun*, 2001).
220 On the other hand, the occurrence of coastal lows are characterised by the poleward prop-
221 agation of a low pressure system along the west coast of South Africa (*Jury et al.*, 1990).
222 This phenomenon causes the strengthening (weakening) of favourable upwelling winds in
223 its leading (trailing) edge (*Jury and Brundrit*, 1992; *Shillington et al.*, 2006) and this in
224 turn may induce more/less upwelling (*Payne et al.*, 1992).

225 1.1.3 Interannual and long-term variability

226 *Blamey et al.* 2012 analysed in detail the Cape Point anemometer time series over the
227 period 1965-2011. Figure 1.3, extracted from their paper, shows the annual average time
228 series of the north-south component of the wind speed responsible for the coastal di-
229 vergence of Ekman transport. The south-easterlies vary on the interannual time scale.
230 Although the signal has some chaotic behaviour, some clear periods of positive/negative
231 upwelling index can be distinguished. *Blamey et al.* (2012, 2015) identified these periods
232 and characterised some of the regime shifts. The period of interest of this study goes from
233 1990 to 2012. During this period, a net amplification of the upwelling favourable winds
234 occurred during the 1990s, followed by a net decrease in the early 2000s.

235 Large scale patterns of atmospheric variability may explain partly these regime shifts.
236 At interannual time scales, the Benguela atmospheric circulation is influenced by large-

²Berg winds are offshore directed and blow mostly during winter when a high-pressure system over the continent co-exists with a low-pressure system along the coastal area.

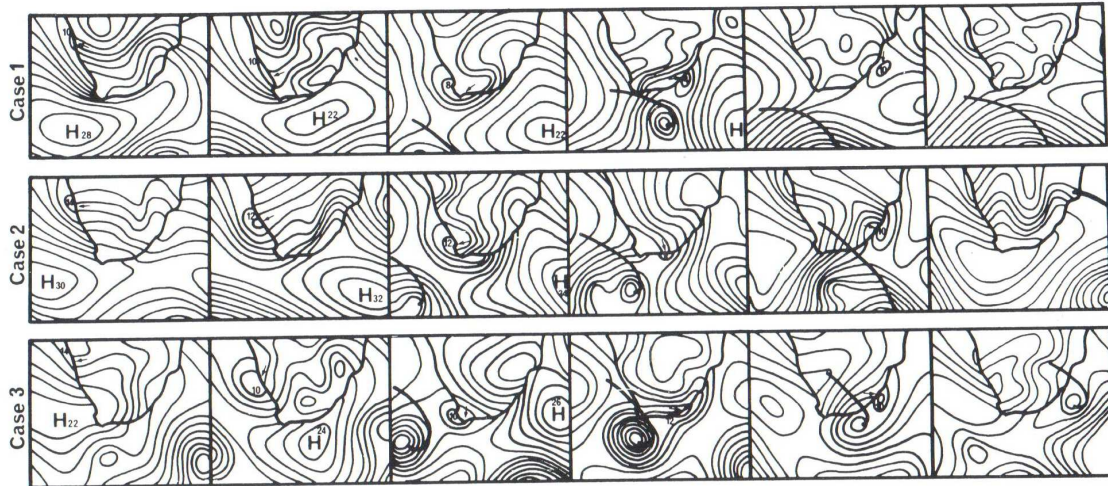


Figure 1.2: Surface synoptic pressure maps of the eastward propagating cyclones passing south of the continent obtained from South African Weather Services (from *Jury et al.* (1990)) and the formation of a trapped coastal low that can promote pulsed upwelling at 3-10 day intervals (*Hutchings et al.*, 2009). The H in each figure refers to the ridging high pressure cell, the bold line to the trailing edge of frontal system, and the arrow points the position of the coastal low. The three cases corresponds to days extending over 6-11 February 1981, 15-20 September 1985 and 18-23 April 1980, respectively.

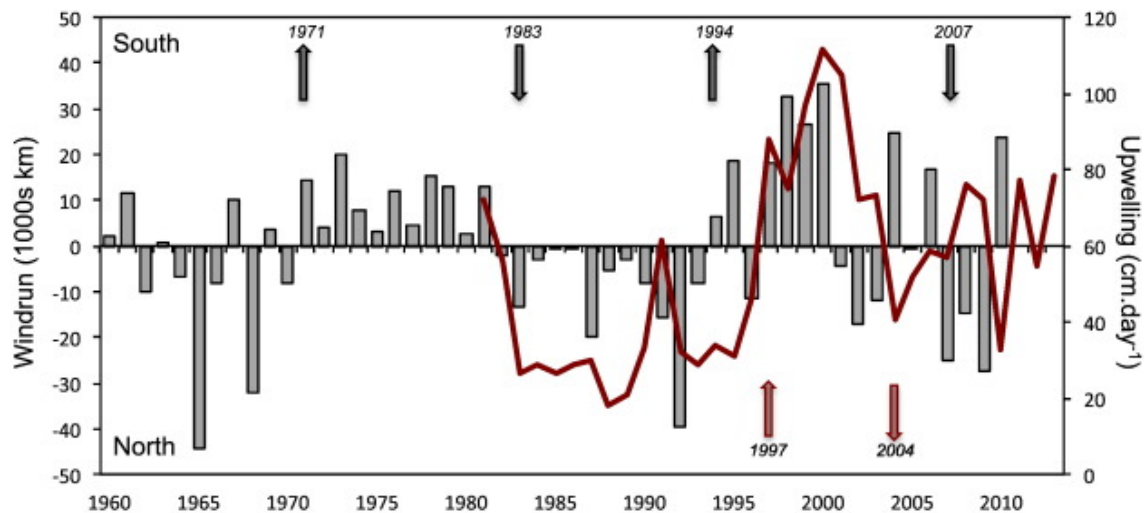


Figure 1.3: Annual anomalies (1960-2010 period) of north/south wind component measured as wind-run (grey bars, km) and summer upwelling rates (cm day^{-1}) for Cape Point (taken from *Blamey et al.* (2015)). Positive/negative bars indicate a dominance of southerlies/northerlies. Positive/negative wind regime shifts and upwelling are denoted by grey and red arrows, respectively. There are some periods where the dominance of southerlies (northerlies) coincide with intensified (reduced) upwelling rates. The early 1980s to early 1990s period was predominately associated with reduced upwelling and northerlies and upwelling was intensified between 1997 and 2000.

237 scale interannual variability of the El Niño-Southern Oscillation (ENSO) and the Southern
238 Annular Mode (SAM) (*Pitcher and Pillar, 2010; Tim et al., 2015*). According to *Tim et al.*
239 (2015), these climate modes of variability influence upwelling through the modulation of
240 local atmospheric conditions.

241

242 Seasonal composites of sea-level pressure (SLP) anomalies indicated that La Niña/El
243 Niño events are characterised by positive/negative SLP anomalies over the South At-
244 lantic mid-latitudes (*Colberg et al., 2004*). Because higher SLP over the South Atlantic
245 mid-latitudes enhance the strength of the south easterly upwelling favourable winds, this
246 positive/negative correlation between an ENSO index and SLP anomalies over the South
247 Atlantic suggests that stronger/weaker south easterlies may be expected during El Niño/La
248 Niña years (*Risien et al., 2004; Dufois et al., 2012*).

249

250 The SAM is the dominant mode of climate variability in the extra-tropical Southern
251 Hemisphere (*Hall and Visbeck, 2002*) and results in meridional shifts of the mid-latitude
252 westerlies on various time scales. These shifts can be monitored using the Marshall Index,
253 defined as the meridional differences in normalised sea level pressure between the Antarc-
254 tic (65°S) and the southern hemisphere mid-latitudes (40°S) (*Marshall, 2003*). Positive
255 SAM indices are associated with a poleward shift of the atmospheric jet stream, which in
256 turn leads to a weakening of the westerlies between 45° S and 20° S (*Hall and Visbeck,*
257 *2002*). Several studies suggest that this weakening of the westerlies is accompanied by a
258 strengthening of the south-easterly upwelling favourable winds (*Hall and Visbeck, 2002;*
259 *Hutchings et al., 2009*). In fact, positive correlations exist between upwelling favourable
260 wind anomalies over the SBUS and summer averages (December-February) of the SAM
261 index (*Tim et al., 2015*).

262 1.2 Ocean circulation

263 The mean regional circulation in the SBUS and its seasonal cycle are described in *Nelson*
264 *and Hutchings (1983)*. Its main features are summarised in Figure 1.4 extracted from
265 *Hardman-Mountford et al. (2003)*.

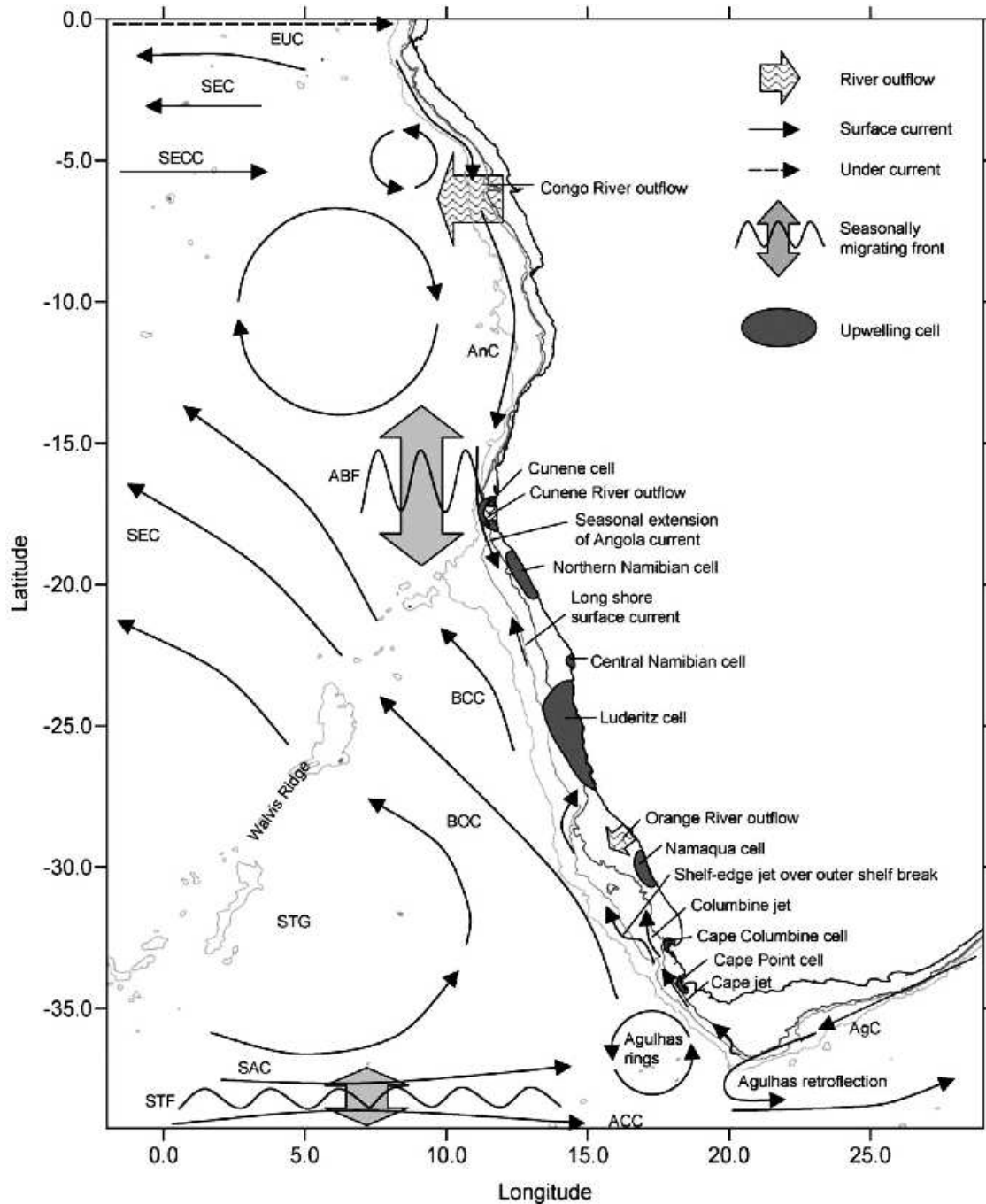


Figure 1.4: Map of the South East Atlantic showing surface and near-surface currents, frontal zones, upwelling cells, major areas of freshwater input and bathymetry adapted from *Hardman-Mountford et al. (2003)*. These are main oceanographic features that could impact biology in the Benguela upwelling system at different temporal and spatial scale (*Shannon et al., 1996*). EUC, Equatorial Under Current; SEC, South Equatorial Current; SECC, South Equatorial Counter Current; AnC, Angola Current; BOC, Benguela Oceanic Current; BCC, Benguela Coastal Current; SAC, South Atlantic Current; AgC, Agulhas Current; ABF, Angola-Benguela front; STF, subtropical front; STG, subtropical gyre; ACC, Antarctic Circumpolar Current. The Benguela Current and Benguela Jet are labelled BOC and 'Cape Jet', respectively.

266 **Some intense alongshore surface and sub-surface currents**

267 The surface circulation in the region is mainly dominated by equatorward currents. The
268 broad (100 km) Benguela Current that constitutes the eastern limb of the South-Atlantic
269 Subtropical Gyre (*Peterson and Stramma, 1991*) is located offshore beyond the shelf edge.
270 Over the shelf edge, another northward along-shore flow co-exists, namely the Benguela
271 Jet. The latter flows northwards along South Africa's west coast between Cape Point
272 and Cape Columbine. It is much narrower (a few tens of kilometres) than the Benguela
273 Current, but has higher velocities ranging from 25 cm s^{-1} to 75 cm s^{-1} (*Bang and Andrews,*
274 *1974; Gordon, 1985*). It is in thermal wind balance with the cross-shore density front that
275 results from the upwelling of cold waters inshore and the presence of warmer waters offshore
276 (*Blanke et al., 2009; Veitch et al., 2017*). Even though the Benguela Jet gets intensified
277 in summer (*Armstrong et al., 1987*), when the coastal upwelling of cold waters occurs, it
278 can be observed throughout the year thanks to the leakage of warm Agulhas waters into
279 the SBUS (*Nelson and Hutchings, 1983; Strub et al., 1998*). Another feature of the SBUS
280 mean circulation is the presence of a poleward undercurrent (PUC) that flows over the
281 continental shelf edge underneath the Benguela Jet (*Shelton and Hutchings, 1982; Nelson*
282 *and Hutchings, 1983*). This is a common feature between all main four EBUS of the world
283 ocean. It is worth noting that this undercurrent may occasionally reach the surface. The
284 surfacing of the poleward undercurrent is not specific to the SBUS. When it happens, it
285 can either split the surface equatorward jet, or shift it offshore (*Song et al., 2011*). This
286 leads to the co-existence of a complex network of alongshore multi-core jets in most EBUS.

287 A theoretical study by *Fennel et al. (2012)* shows that the cross-shore horizontal and
288 vertical structure of these jets is governed by the bathymetry, the alongshore wind-stress,
289 and the curl of the wind stress. The latter results mainly from the coastal wind drop-off. Its
290 signature is the presence of an along-shore 300 km width coastal band of cyclonic negative
291 wind stress curl (WSC). Increasing the cyclonic WSC over the shelf edge strengthens the
292 poleward undercurrent, which in turn displaces the surface equatorward jet offshore. As
293 expected, in the absence of WSC, the shelf edge equatorward jet relocates closer to the
294 coast (*Song et al., 2011*). However, the poleward undercurrent is always present with or
295 without WSC. A study by *Junker et al. (2015)* suggests that the poleward undercurrent

296 may be modulated by remotely generated coastal Kelvin wave signals.

297 **An intense mesoscale eddy activity**

298 Eddies and filaments are general mesoscale features of the SBUS (*Rubio et al.*, 2009).
299 They can be generated either locally from mixed barotropic/baroclinic instabilities of the
300 density fronts (*Veitch et al.*, 2010), or remotely in the Agulhas retroflection region. Their
301 presence modulates the schematic picture of the circulation given by the mean flow and
302 potentially play a large role in structuring the ecosystem (*Hutchings et al.*, 1998; *Rubio*
303 *et al.*, 2009).

304 In EBUS, mesoscale eddies play a significant role in the cross-shore transport of buoyancy,
305 balancing the net surface heat losses and gains (*Marchesiello et al.*, 2003; *Capet et al.*,
306 2008b; *Colas et al.*, 2013). Their ability to generate cross-isobaths flow breaks the dynam-
307 ical barrier set up by the shelf slope and facilitates cross-shore exchanges of water masses
308 between the coastal upwelling zone and the adjacent open ocean. Eddies and filaments
309 may thereby entrap coastal shelf biological material and drag it out into the open ocean,
310 impacting the overall cumulative cross-shore mass transport of biota (*Hernández-Carrasco*
311 *et al.*, 2014; *Veitch et al.*, 2010). In addition, vertical velocities are triggered on their lat-
312 eral boundaries when horizontal density gradients are enhanced by frontogenesis (*Capet*
313 *et al.*, 2008b; *Lévy et al.*, 2012).

314

315 **A unique location in the vicinity of a western boundary current**

316 The SBUS is distinct by having the warm Agulhas Current with its ring shedding as its
317 southern boundary, unlike any other Eastern Boundary Upwelling Systems. The Agulhas
318 Current forms the western branch of the South Indian Ocean subtropical gyre and flows
319 southward along the east coast of South Africa. As it reaches the southern tip of the
320 African continent at 20°E, it turns back and flows eastward into the South Indian Ocean.
321 This process results in the generation of intense mesoscale activity including the generation
322 of shelf-edge cyclonic eddies as it interacts with the Agulhas Bank (*Penven et al.*, 2001).
323 The retroflection of the Agulhas current is also accompanied by a leakage (18 Sv) of warm

324 and salty waters of Indian Ocean origin into the South Atlantic Ocean (*Loveday et al.*,
 325 2014). This leakage occurs through the shedding of large (radius > 100 km) anticyclonic
 326 Agulhas Rings, smaller (radius < 50 km) cyclonic eddies and filaments. Agulhas rings are
 327 shed at a rate of 5-6 per year when the Agulhas Current retroflects eastward. They then
 328 propagate into the South Atlantic. Each of them transports large volumes (0.5-1.5 Sv)
 329 of Indian Ocean waters (*Schouten et al.*, 2000; *van Aken et al.*, 2003; *Doglioli et al.*,
 330 2006). This leakage is partly responsible for the enhanced cross-shore density gradient
 331 observed in the SBUS, which explains why the Benguela Jet is such a narrow and intense
 332 current, squeezed onto the shelf edge in between the 200 m and 500 isobaths. In addition,
 333 this leakage also explains why the SBUS is characterised by a strong offshore mesoscale
 334 variability reflected by high EKE values in excess of $\sim 500 \text{ cm}^{-2}\text{s}^{-2}$ that extends to -600 m
 335 depth in contrast to the nearshore region (EKE $\sim 100 \text{ cm}^2.\text{s}^{-2}$) and the $\sim 120 \text{ cm}^{-2}\text{s}^{-2}$
 336 in the offshore California Current System (*Capet et al.*, 2008b; *Veitch et al.*, 2009; *Veitch*
 337 *and Penven*, 2017).

338 The presence of this very intense mesoscale turbulent activity in the coastal transition
 339 zone can also drive a strong intra-seasonal variability of the shelf circulation. There are
 340 also some rare observations of Agulhas rings interacting with the Benguela upwelling front
 341 (*Duncombe-Rae et al.*, 1992), triggering mesoscale instabilities leading to the formation
 342 filaments of cold water that extends further offshore in the open ocean (*Payne et al.*, 1992).

343 **The coastal poleward current**

344 Shelf and nearshore currents vary in response to the short-term fluctuations of the atmo-
 345 spheric forcing (*Nelson*, 1989). Variations of the nearshore negative cyclonic wind stress
 346 curl associated with atmospheric boundary layer dynamics and orographic effects may
 347 modulate the intensity and occurrence of poleward surface flow in the first 200 m depth
 348 coastal band (*Brown and Hutchings*, 1987; *Marchesiello et al.*, 2003; *Veitch et al.*, 2009).
 349 The presence of cyclonic wind stress curl on the innershelf may induce a poleward counter
 350 current consistent with the Sverdrup relation (*Vallis*, 2006). This innershelf counter cur-
 351 rent may sometimes merge with the offshore poleward undercurrent located on the shelf
 352 slope (*Junker et al.*, 2015).

353 In addition, periodic current reversals can also occur during episodes of relaxation of
354 favourable upwelling wind stress and are characteristic of all EBUS. In the California
355 mid-shelf (depths over 100 m), they have been extensively documented (*Send et al.*, 1987;
356 *Winant et al.*, 1987; *Huyer and Kosro*, 1987; *Kosro*, 1987; *Largier et al.*, 1993). In-situ ob-
357 servations of the California mid-shelf region, show that these episodes generate a complex
358 inner shelf circulation characterised by a strong counter-current (*Cudaback et al.*, 2005;
359 *Roughan et al.*, 2006) and poleward flow of warm waters that can accumulate inside bays
360 (*Gan and Allen*, 2002b; *Melton et al.*, 2009; *Washburn et al.*, 2011).

361 In the SBUS, there is also evidence of current reversals associated with the poleward prop-
362 agation of warm waters on the innershelf (*Fawcett et al.*, 2008). They can lead to a large
363 increase of temperature up to $5^{\circ}\text{C day}^{-1}$ in the very nearshore region and inside of St
364 Helena Bay. This occurs when the equatorward wind stress relaxes and the nearshore up-
365 welling circulation is no longer in geostrophic balance resulting in a counter-current that
366 flows around the headlands and transports some warm waters (*Gan and Allen*, 2002a).

367 In EBUS, the response of the shelf-circulation to the short-term variability of the atmo-
368 spheric forcing has been investigated using regional ocean models. The importance of
369 using a realistic, time-varying surface forcing was demonstrated by *Blanke et al.* (2005)
370 who found that a daily wind product is more efficient in simulating the observed nearshore
371 interannual variability of sea surface temperature anomalies (warm and cold events) and
372 short-term upwelling events than a weekly product (*Blanke et al.*, 2002).

373 **1.3 Conclusion**

374 The literature review presented in this chapter on the southern Benguela ocean circulation
375 has identified complex physical features with different temporal and spatial scales. The
376 Benguela Jet plays a role as a vehicle for fish eggs and larvae from the southern tip of
377 the SBUS to St Helena Bay on the west coast. However due to lack of longer in-situ
378 measurements, the variability of the jet in terms of its structure and intensity has not
379 been thoroughly studied. Hence, it is not fully known how this variability in the jet and
380 mesoscale variability influences the alongshore transport of fish eggs and larvae. Therefore,
381 other tools such as numerical models can be used to provide us with desired spatially and

382 temporally cohesive outputs in order to study the ocean circulation of the SBUS. The
383 next chapter deals with the ROMS model that is used in this thesis to simulate the ocean
384 circulation of the SBUS.

385 Chapter 2

386 The ROMS model

387 This chapter focuses on the description of the ocean model employed in this study to
388 simulate the ocean circulation in the Benguela Upwelling System. Here, we focus on
389 the numerical details of the Regional Oceanic Modeling System (ROMS; Shchepetkin
390 and McWilliams 2005; Shchepetkin 2015) in its Coastal and Regional Ocean Community
391 (CROCO¹) version (*Debreu et al.*, 2012).

392 Readers who are familiar with the ROMS model may skip this chapter and move
393 directly to chapter 3 in which the reference simulation is described. However, readers
394 who wish to learn more about the components necessary to develop a numerical ocean
395 circulation model may find here some relevant information about the equations that are
396 time integrated, the vertical and lateral boundary conditions, the numerical schemes, and
397 the horizontal and vertical discretization.

398 2.1 The primitive equations

399 The primitive equations are a set of partial differential equations. They result directly from
400 the very general momentum conservation Navier-Stokes equations for Newtonian rotating
401 fluids (*Kundu et al.*, 2015), when one considers together the flow to be incompressible,
402 hydrostatic, of Boussinesq type (*Vallis*, 2006). This set of equations is completed by an
403 equation of state that provides the fluid density as a function of pressure, temperature
404 and salinity. Therefore, two tracer conservation equations are added for temperature and

¹<http://www.croco-ocean.org/>

405 salinity respectively. After combining all together, we obtain the following set of equations,
 406 here written in Cartesian coordinates:

- 407 • 2 prognostic equations for u and v derived from the conservation of horizontal mo-
 408 mentum

$$\frac{\partial u}{\partial t} + \vec{v} \cdot \vec{\nabla} u - fv = -\frac{\partial \phi}{\partial x} + D_u \quad (2.1)$$

$$\frac{\partial v}{\partial t} + \vec{v} \cdot \vec{\nabla} v + fu = -\frac{\partial \phi}{\partial y} + D_v \quad (2.2)$$

- 410 • 1 diagnostic equation for the pressure derived from the vertical momentum conser-
 411 vation equation with the hydrostatic assumption

$$\frac{\partial \phi}{\partial z} = -\frac{\rho g}{\rho_o} \quad (2.3)$$

- 412 • 2 prognostic equations for potential temperature T and salinity S derived from the
 413 conservation of tracer content

$$\frac{\partial T}{\partial t} + \vec{v} \cdot \vec{\nabla} T = D_T \quad (2.4)$$

$$\frac{\partial S}{\partial t} + \vec{v} \cdot \vec{\nabla} S = D_S \quad (2.5)$$

- 415 • 1 diagnostic equation for w derived from for the continuity equation (mass conser-
 416 vation + incompressible flow)

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \quad (2.6)$$

- 417 • 1 diagnostic equation for the water density derived from for the equation of state for
 418 seawater (*Shchepetkin and McWilliams, 2005*)

$$\rho = \rho(T, S, P). \quad (2.7)$$

419 x, y, z are the independent variables related to the east-west, north-south, vertical direc-
 420 tions respectively, and t the time independent variable. The prognostic variables are the

421 zonal (u), meridional (v) and vertical (w) components of the velocity vector \vec{v} , the poten-
422 tial temperature (T) and the salinity (S). ρ is the seawater density, $\rho_0 = 1025 \text{ kg m}^{-3}$ the
423 reference seawater density, $f = 4\pi/86400 \sin(\lambda)$ the Coriolis parameter, λ the latitude,
424 ϕ the dynamic pressure, \vec{g} the constant of gravity. D_u , D_v , D_T and D_S are the diffusive
425 terms for momentum, temperature and salt respectively.

426 The hydrostatic assumption neglects the vertical acceleration terms, and the Boussi-
427 nesq assumption neglects the variation of density in the horizontal momentum equations.
428 The diffusive terms of the primitive equation include second order partial derivatives ac-
429 cording to x , y and z and some knowledge of the diffusion coefficients for momentum and
430 heat and water fluxes. They can be decomposed into a vertical part ($_v$ suffix) and an
431 horizontal part ($_h$ suffix).

$$D_u = D_{uh} + D_{uv} \quad (2.8)$$

$$D_v = D_{vh} + D_{vv} \quad (2.9)$$

$$D_T = D_{Th} + D_{Tv} \quad (2.10)$$

$$D_S = D_{Sh} + D_{Sv} \quad (2.11)$$

For the momentum equations , we have

$$D_{uh} = \frac{\partial}{\partial x} \left(K_{Mh} \frac{\partial u}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{Mh} \frac{\partial u}{\partial z} \right) \quad (2.12)$$

$$D_{uv} = \frac{\partial}{\partial z} \left(K_{Mv} \frac{\partial u}{\partial z} \right) \quad (2.13)$$

$$D_{vh} = \frac{\partial}{\partial x} \left(K_{Mh} \frac{\partial v}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{Mh} \frac{\partial v}{\partial z} \right) \quad (2.14)$$

$$D_{vv} = \frac{\partial}{\partial z} \left(K_{Mv} \frac{\partial v}{\partial z} \right) \quad (2.15)$$

where K_{Mh} and K_{Mv} are the coefficients of horizontal and vertical eddy viscosity for

momentum. For the tracers equations , we have

$$D_{Th} = \frac{\partial}{\partial x} \left(K_{Th} \frac{\partial u}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{Th} \frac{\partial u}{\partial z} \right) \quad (2.16)$$

$$D_{Tv} = \frac{\partial}{\partial z} \left(K_{Tv} \frac{\partial u}{\partial z} \right) \quad (2.17)$$

$$D_{Sh} = \frac{\partial}{\partial x} \left(K_{Sh} \frac{\partial v}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_{Sh} \frac{\partial v}{\partial z} \right) \quad (2.18)$$

$$D_{Sv} = \frac{\partial}{\partial z} \left(K_{Sv} \frac{\partial v}{\partial z} \right) \quad (2.19)$$

432 where K_{Th}/K_{Sh} and K_{Tv}/K_{Sv} are the coefficients of horizontal and vertical eddy diffu-
433 sivity for heat/salt.

434 The presence of first order and second order spatial derivatives in the primitive equa-
435 tions implies the use of boundary conditions. These boundary conditions are used to
436 specify the fluxes of momentum, heat and fresh-water at the lateral and surface bound-
437 aries. In the presence of hard boundaries (bottom, land), the no-flux condition across the
438 boundary is used.

439 2.2 The vertical boundary conditions

440 The presence of vertical partial derivatives in the primitive equation requires surface and
441 bottom boundary conditions to be specified. These conditions are of particularly impor-
442 tance because they control the inputs and losses of momentum, heat and fresh-water fluxes
443 through the surface and bottom boundaries.

444 2.2.1 Surface boundary condition at $z=\eta(\mathbf{x},\mathbf{y},\mathbf{t})$

445 At the surface, at $z=\eta(\mathbf{x},\mathbf{y},\mathbf{t})$ where η is the free surface elevation, the ocean receives in-
446 put of momentum through wind stress, as well as inputs or losses of heat/salt trough
447 heat/fresh-water surface fluxes. These fluxes relate to either the local atmospheric com-
448 ponents of the wind stress, or the heat and freshwater surface fluxes:

$$K_{Mv} \frac{\partial u}{\partial z} = \frac{\tau_x^s}{\rho_o}(x, y, t) \quad (2.20)$$

$$K_{Mv} \frac{\partial v}{\partial z} = \frac{\tau_y^s}{\rho_o}(x, y, t) \quad (2.21)$$

$$K_{Tv} \frac{\partial T}{\partial z} = \frac{Q_{net}}{\rho_o C_p} \quad (2.22)$$

$$K_{Sv} \frac{\partial S}{\partial z} = \frac{(E_v - P_r) SSS}{\rho_o} \quad (2.23)$$

449 τ_x^s, τ_y^s are the surface wind stress in x and y directions respectively, where as Q_T is the
 450 net surface heat flux, $(E_v - P_r)SSS$ is the net fresh-water flux, with E_v the net evaporation
 451 rate, P_r the precipitation rate freshwater flux and SSS the sea surface salinity.

452 A quadratic law is used to express the surface wind stress components as a function of the
 453 atmospheric wind velocity components (u_{10}^a and v_{10}^a) 10 m above sea level, the air density
 454 ρ_{air} , and the surface drag coefficient C_D^s .

$$\tau_x^s = \rho_{air} C_D^s \sqrt{u_{10}^2 + v_{10}^2} u_{10} \quad (2.24)$$

$$\tau_y^s = \rho_{air} C_D^s \sqrt{u_{10}^2 + v_{10}^2} v_{10}. \quad (2.25)$$

456 In addition a dynamical condition is imposed on the vertical velocity so that the free
 457 surface behaves as a material derivative:

$$w = \frac{\partial \eta}{\partial t} + u \frac{\partial \eta}{\partial x} + v \frac{\partial \eta}{\partial y} \quad (2.26)$$

458 **2.2.2 Bottom boundary condition at $z = -H(x, y)$**

459 At the bottom of the ocean, a condition of no flow is imposed, as well as zero heat and
 460 freshwater fluxes across the seafloor:

$$w = - \left(u \frac{\partial H}{\partial x} + v \frac{\partial H}{\partial y} \right) \quad (2.27)$$

$$K_{Tv} \frac{\partial T}{\partial z} = 0 \quad (2.28)$$

$$K_{Sv} \frac{\partial S}{\partial z} = 0. \quad (2.29)$$

461 Momentum is dissipated through bottom friction and the bottom boundary conditions
 462 are

$$K_v \frac{\partial u}{\partial z} = \frac{\tau_x^b}{\rho_o}(x, y, t) \quad (2.30)$$

$$K_v \frac{\partial v}{\partial z} = \frac{\tau_y^b}{\rho_o}(x, y, t), \quad (2.31)$$

463 with τ_x^b, τ_y^b the bottom stress in the in x and y directions respectively. The expression for
 464 bottom stress uses a combination of linear, quadratic, or logarithmic terms (*Hedström,*
 465 2009):

$$\tau_x^b = (-r + C_D^b \sqrt{u^2 + v^2}) \cdot u_b \quad (2.32)$$

466 and

$$\tau_y^b = (-r + C_D^b \sqrt{u^2 + v^2}) \cdot v_b \quad (2.33)$$

467 where r is a friction velocity and C_D^b bottom a non dimensional **bottom** drag coefficient.

- "Linear friction",

Linear friction can be implemented by setting $\mathbf{C}_D^b = \mathbf{0}$, which implies:

$$(\tau_b^x, \tau_b^y) = -r (u_b, v_b)$$

468 .

469

- "Quadratic friction",

Quadratic friction can be implemented by setting $\mathbf{r} = \mathbf{0}$, which implies:

$$(\tau_b^x, \tau_b^y) = C_D^b \sqrt{u_b^2 + v_b^2} (u_b, v_b)$$

470 .

- "Quadratic friction using a Von Karman Prandtl logarithmic layer "

471 The "Von Karman Quadratic Friction" has the same formulations as the "classic"
 472 quadratic friction ($\mathbf{r} = \mathbf{0}$). The only difference lies in the the parameterization of
 473

474 the drag coefficient C_D^b :

$$C_D^b = \left(\frac{\kappa}{\log[\Delta z_b/z_r]} \right)^2 \quad \text{if } C_D^{min} < C_D^b < C_D^{max}, \text{ otherwise} \quad (2.34)$$

$$C_D^b = C_D^{min} \quad \text{or} \quad C_D^b = C_D^{max} \quad \text{if } C_D^b \text{ is outside this range} \quad (2.35)$$

475 $\kappa = 0.41$ represents the Von Karman constant and Δz_b , the thickness from the first
 476 level to the bottom. C_D^b becomes variable and depends on the thickness of the first
 477 vertical level, reaching its maximum value if the vertical resolution near the bottom
 478 is close to the scale of roughness z_r .

479 2.3 Surface fluxes forcing in the model

480 2.3.1 Momentum surface fluxes

481 The wind forcing is imposed in the form a surface stress τ^s (Eqs. 2.24 and 2.25) in the
 482 surface boundary conditions of the horizontal momentum equations. τ^s is calculated at
 483 each point on the model grid from the wind speed components 10 m above sea level
 484 u_{10}^a, v_{10}^a , the air density and the drag coefficient C_D^s . Several parametrizations exist to
 485 estimate C_D^s . The latter is usually calculated using a so-called "bulk" parametrization.
 486 In the following, we chose the one proposed by *Fairall et al.* (1996). All the atmospheric
 487 data required to calculate the surface wind stress are provided by the 6 hourly Climate
 488 Forecast System Reanalysis (CFSR) atmospheric reanalysis (*Saha, 2010*).

489 2.3.2 Heat fluxes and fresh water flux

490 The surface boundary conditions in the two prognostic equations for temperature and
 491 salinity require the specification of a net heat and fresh-water flux at each grid point of
 492 the model grid. The net heat flux is the sum of four components:

$$Q_{net} = Q_{solar_net} + Q_{long-wave_net} + Q_{latent} + Q_{sensible}, \quad (2.36)$$

493 where Q_{solar_net} is "short wave" radiative flux due to solar radiation penetrating into the
 494 ocean's surface top layers, $Q_{long-wave_net}$ the "long wave" infrared flux emitted from the

495 Earth to the atmosphere, Q_{latent} the latent heat flux due to condensation/evaporation
 496 processes at the ocean-atmosphere interface and $Q_{sensible}$ the sensitive heat flux due to
 497 heat conduction at the ocean-atmosphere interface. It is worth noting that the latent and
 498 sensitive heat fluxes are "turbulent" heat fluxes.

499 The formulation of the net fresh water fluxes EmP_{net} has already been presented in Eq. 2.23
 500 and is computed as :

$$EmP_{net} = (E_v - P_r) SSS \quad (2.37)$$

501 with E_v the net evaporation rate, $Prate_{net}$ the precipitation rate and SSS the sea surface
 502 salinity.

503

504 To be consistent with the wind stress parameterization described above, the explicit
 505 calculation of Q_{net} and EmP_{net} are based on the "bulk" parametrization of *Fairall et al.*
 506 (1996), and all the necessary atmospheric data required in calculations are provided by
 507 the 6 hourly CFSR reanalysis atmospheric reanalysis (*Saha, 2010*).

508 2.4 Vertical mixing and Turbulent closure scheme

509 Subgrid unresolved turbulent processes are parametrized using a downgradient diffusive
 510 law. The vertical turbulent diffusivity coefficient is calculated using the non-local K-
 511 profile planetary (KPP) scheme (*Large et al., 1994*). In the interior, vertical mixing is
 512 mainly due to shear instabilities associated with internal wave activity, as well as double
 513 diffusion. Shear instabilities occur when the Richardson number becomes higher than
 514 0.25. The KPP mixing scheme consists in calculating the boundary layer depth (h) at
 515 each grid point as a function of a bulk Richardson number, and the surface/bottom wind
 516 stress and buoyancy forcing. The vertical diffusivity coefficient is then increased artificially
 517 within these boundary layers. Some special functions are used to insure smooth transitions
 518 between the interior and surface/bottom vertical mixing coefficients. The KPP scheme
 519 has proven to work well at mid-latitudes and in the tropics. There, it allows a fair realistic
 520 representation of the surface mixed layer during convective and storm events, as well as
 521 its diurnal cycle (*Large et al., 1994*).

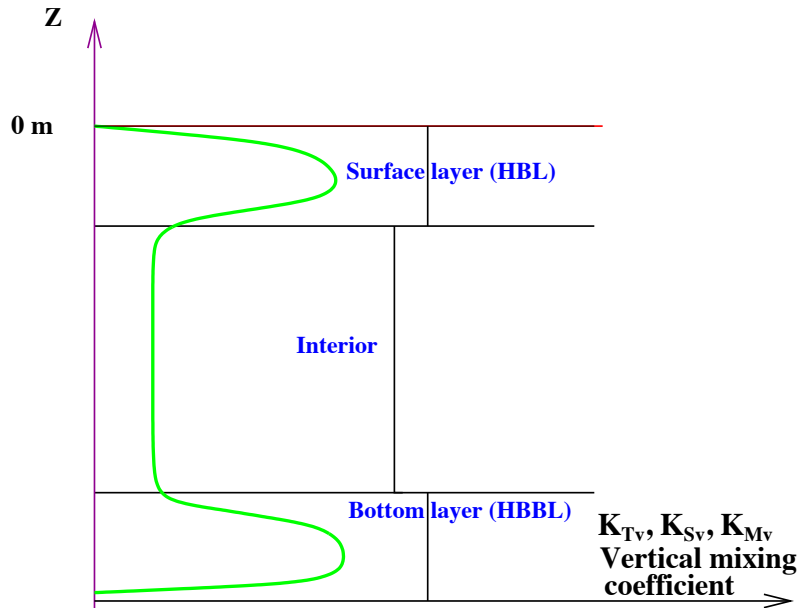


Figure 2.1: Schematic diagram of the KPP closure model used. The mixing coefficients computed by the KPP closure model are the vertical thermal diffusivity K_{Tv} , vertical haline diffusivity K_{Sv} and the vertical momentum viscosity K_{Mv} .

522 2.5 Horizontal mixing

523 Horizontal turbulent mixing processes occur through the diffusive terms of the primitive
 524 equations, generally in the form a laplacian or bilaplacian operator. The values of the
 525 diffusive coefficients for momentum K_{Mh} , heat K_{Th} , and salt K_{Sh} are "turbulent" values.
 526 Their expression should result from a parametrization of the turbulent sub-grids non
 527 resolved processes. Nevertheless, in practise, they are set to a minimum value chosen to be
 528 high enough so that numerical artefacts are dissipated. This minimum value is generally
 529 higher then the real value, and very fine scale dynamics tend to be over-dissipated in
 530 the simulations, resulting in over-smoothed solutions. In ROMS, the use of third-order
 531 advection scheme with implicit dissipation theoretically prevents the user from adding
 532 horizontal viscosity.

533 2.6 Lateral Boundary Conditions

534 The presence of horizontal partial derivatives in the momentum equations requires speci-
 535 fying some lateral boundary conditions for u , v , T and S .

536 **The case of land-boundary**

537 In the case of land-boundary, conditions are similar to some extent to the conditions
538 imposed at the seabed, *i.e.* no inflow and zero heat and freshwater flux across the land
539 boundary:

$$u_{\perp} = 0 \tag{2.38}$$

$$K_{Th} \frac{\partial T}{\partial x_{\perp}} = 0 \tag{2.39}$$

$$K_{Sh} \frac{\partial S}{\partial x_{\perp}} = 0 \tag{2.40}$$

540 where x_{\perp} is the direction perpendicular to the land boundary. For the velocity component
541 parallel to the land-boundary, a no-slip or free-slip condition (*Hedstrom, 1997*) is imposed.

542 **The case of open boundary**

543 In the case of open ocean borders, boundary conditions are slightly different. One wishes to
544 impose the surrounding ocean known temperature, salinity and velocities at the boundary,
545 as well as to allow disturbances and waves originating from the domain to leave without
546 influencing the interior solution. In ROMS, open boundary conditions are implemented
547 with a method that combines outward oblique radiation and adaptive nudging towards a
548 known external data (*Marchesiello et al., 2001*). "Sponge" layers, within which the lateral
549 viscosity coefficient is smoothly increased from the background interior value towards a
550 maximum value at the boundary are also implemented. Their role is to smooth out model-
551 boundary inconsistencies, prevent spurious reflection of the external/internal solution and
552 maintain long-term model stability. The width of the sponge layer is typically 10 grid
553 cells. The external values of temperature, salt and velocity components at the boundary
554 are either specified from observations or models with larger domains.

555 **2.7 Time discretization**

556 To limit computational cost, ROMS uses a split-explicit time-stepping integration tech-
557 nique that decomposes the three-dimensional fields into a fast barotropic and a slow baro-

558 clinic mode (*Shchepetkin and McWilliams, 2005*). The primitive equations are vertically
559 integrated to provide the prognostic variables of the barotropic mode: η , \bar{u} , \bar{v} . The verti-
560 cally integrated two-dimensional (2D), momentum equations are solved using a time-step
561 typically 10 times smaller than the baroclinic step used to solve the three-dimensional (3D)
562 baroclinic momentum and tracer equations (Fig. 2.2). Before being re-injected into the
563 into the three-dimensional equations, the barotropic variables are time-averaged over the
564 baroclinic time-step in order to limit aliasing errors that may result from high frequencies
565 dynamics present in the barotropic 2D equations (*Shchepetkin and McWilliams, 2005*).

566 The time stepping numerical scheme for the 2D and 3D equations follows a robust plan
567 that combines a Leap-Frog predictor scheme and a corrector Adams-Moulton scheme. This
568 technique allows the use of larger time-steps by a factor of about four (*Shchepetkin and*
569 *McWilliams, 2005*).

570 This method was found to be more robust because it is more efficient to resolve two-
571 dimensional than three-dimensional equations over a short-time step. It also allows to
572 overcome the limitations imposed by the Courant-Friedrichs-Levy (CFL) stability condi-
573 tion (*Ezer et al., 2002; Shchepetkin and McWilliams, 2005*).

574

575 2.8 Horizontal discretization

576 In the horizontal direction, the model variables are discretized on a numerical staggered
577 Arakawa-C grid (Figure 2.4). This grid is well suited for models with resolution finer than
578 the internal Rossby radius (*Arakawa and Lamb, 1977*). The density (ρ), depth (h), Cori-
579 olis parameter (f), vertical velocity (ω), sea surface height and tracers (temperature and
580 salinity) are defined at the centre of the grid-cell, while the horizontal velocity components
581 (u)/(v) are defined at the middle of its eastern/northern and western/southern boundary.

582 Note that ROMS uses an orthogonal curvilinear coordinate system (Figure 2.4). This
583 system can follow an irregular lateral boundary (grid rotation, increased resolution toward
584 the coast, etc.). It is advantageous because the computational grid placing is quite flexible
585 and can be optimised for a given problem.

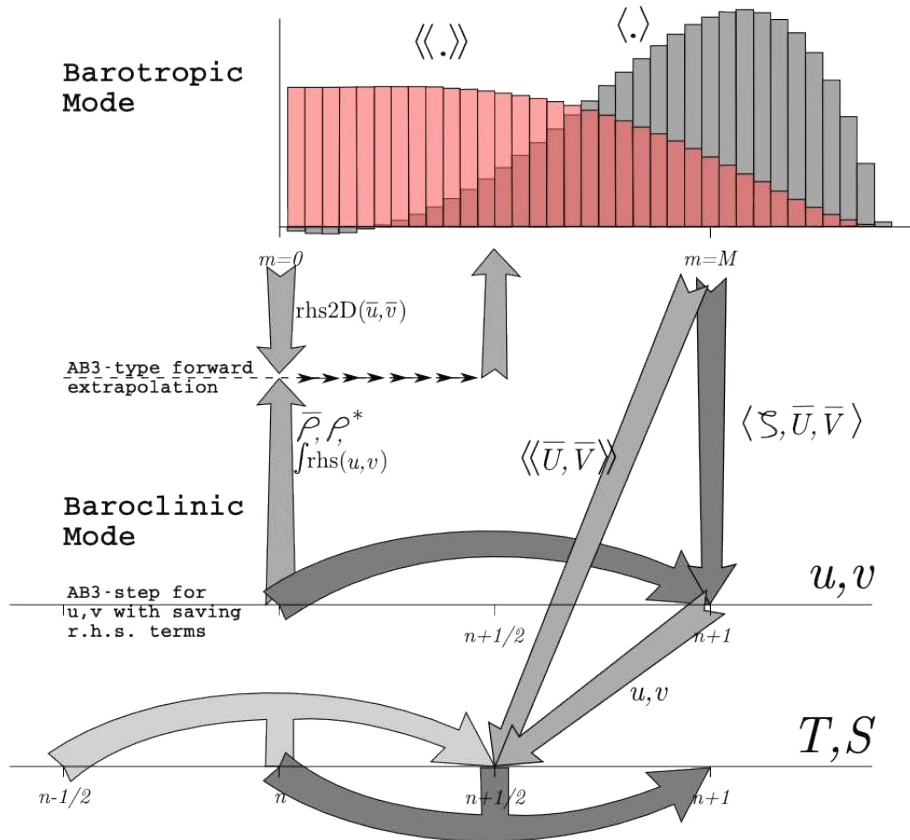


Figure 2.2: Schematic view of the ROMS predictor-corrector hydrostatic kernel. From (Shchepetkin and McWilliams, 2005). Ξ is here the free surface elevation (η in the text).

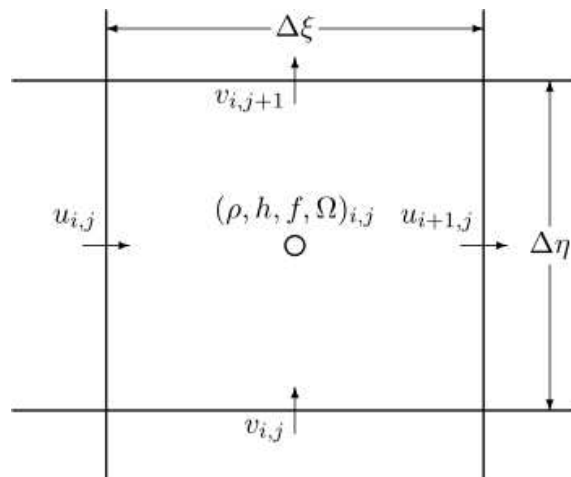


Figure 2.3: The placement of variables on the Arakawa-C grid used in the horizontal discretization of ROMS. From Hedström (2009).

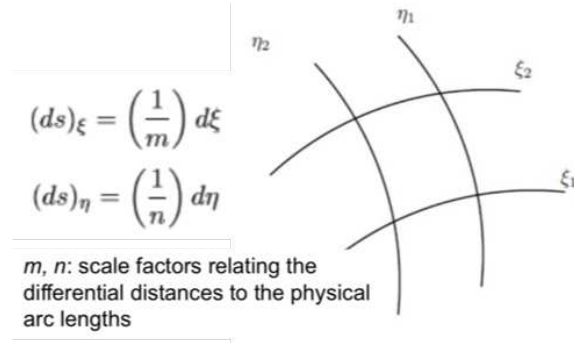


Figure 2.4: Curvilinear horizontal coordinate

586 2.9 Vertical discretization

587 In the vertical direction, ROMS uses a staggered grid (Fig. 2.5) where ρ points are inter-
 588 posed between two vertical velocity Ω points.

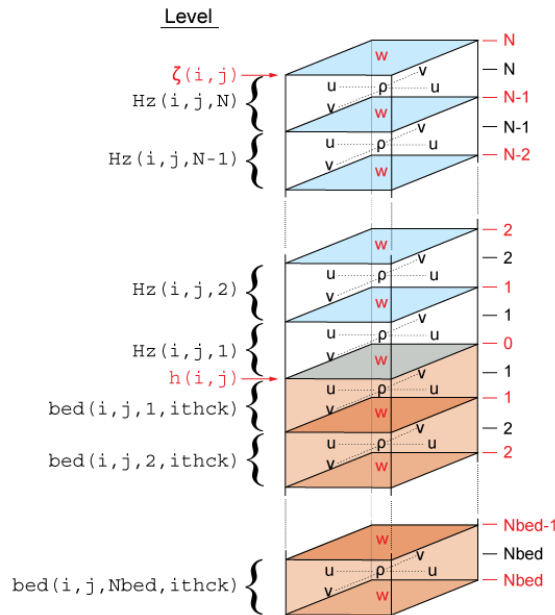


Figure 2.5: Vertical grid used in the vertical discretization of ROMS (from www.myroms.org).

589 ROMS follows a sigma or terrain-following stretched coordinate system which improves
 590 the representation of the topography, increasing the vertical resolution over the continental
 591 shelf (*Song and Haidvogel, 1994*). Sigma coordinates are particularly adapted to study
 592 shelf dynamics and to enhance resolution in the surface mixed layer and at the bottom for
 593 the proper representation of bathymetry-flow interaction (*Veitch et al., 2006*).

594 The calculation of the depth of sigma layers at each grid point takes into account the

$$z(x, y, \sigma, t) = \zeta(x, y, \sigma) + [\zeta(x, y, t) + h(x, y)] z_0(x, y, \sigma) \quad (2.41)$$

$$z_0(x, y, \sigma) = \frac{h_c \sigma + h(x, y) Cs(\sigma)}{h_c + h(x, y)} \quad (2.42)$$

596 where $\zeta(x, y, t)$ is the free-surface elevation, $h(x, y)$ the ocean bottom depth, σ a frac-
 597 tional vertical stretching coordinate ($-1 \leq \sigma \leq 0$), h_c a positive thickness control-
 598 ling the stretching, $Cs(\sigma)$ a non-dimensional monotonic vertical stretching parameter
 599 ($-1 \leq Cs(\sigma) \leq 0$), $z_0(x, y, \sigma)$ a nonlinear vertical transformation.

$$sc = \frac{\sigma - N}{N} \quad (2.43)$$

$$csf = \frac{1. - \cosh(\theta_s sc)}{\cosh(\theta_s) - 1.} \quad \text{if } \theta_b > 0, \quad csf = -sc^2 \quad \text{otherwise} \quad (2.44)$$

$$Cs(\sigma) = \frac{e^{\theta_b csf} - 1.}{1. - e^{-\theta_b}} \quad \text{if } \theta_s > 0, \quad Cs(\sigma) = csf \quad \text{otherwise} \quad (2.45)$$

600 The parameters, θ_s , θ_b and h_{cline} control the vertical grid resolution and stretching.

601 • h_{cline} thus defines the thickness at which to increase the resolution in vertical grid
 602 surface

603 • θ_s ($\theta_s > 0$), θ_b defines the rate of resolution increase in the surface layer.

604 • θ_b ($0 < \theta_b < 1$) controls the resolution increase in the bottom layer, $\theta_b = 1$ correspond
 605 to the same stretching than the surface. $\theta_b = 0$ correspond to no refinement of
 606 resolution at the bottom.

607 Figure 2.6 illustrates the sigma levels through a vertical section in the model configuration
 608 that was developed and that is presented in chapter 3.

²For simplicity we only present the latest transformation available in ROMS, NEW_S_COORD option, other are still available and can be found in the online documentation <http://www.croco-ocean.org>

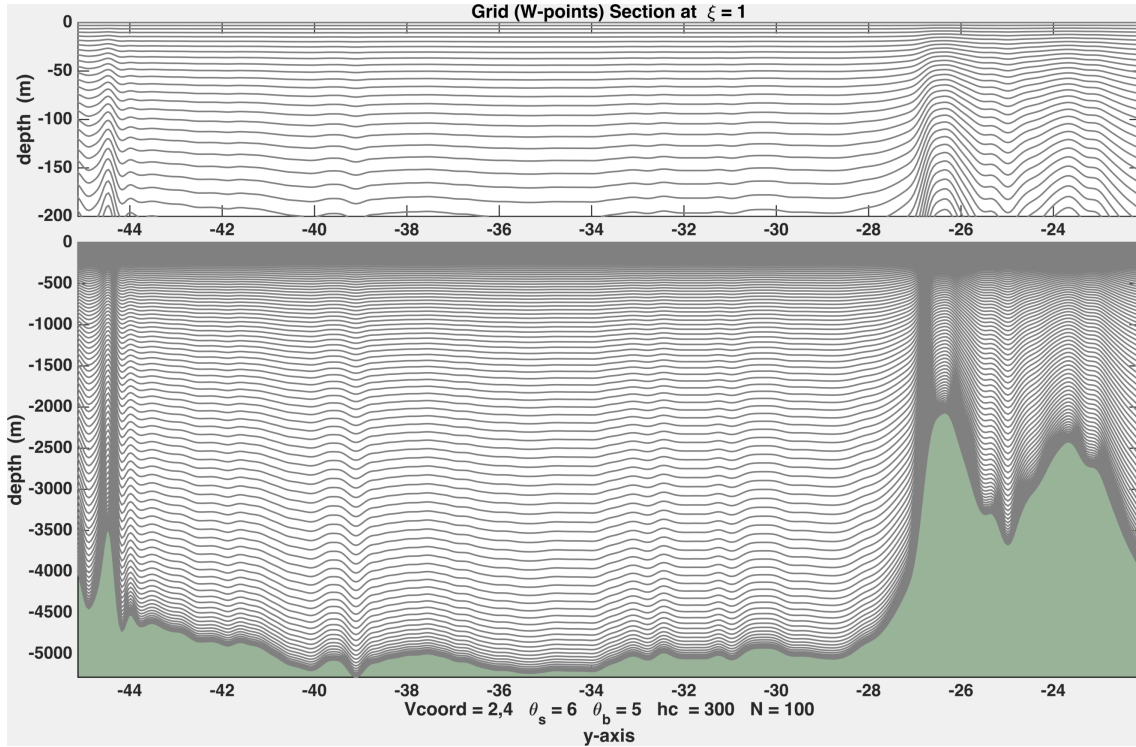


Figure 2.6: An example of the BENGR15 vertical grid as prescribed by the σ -coordinate system. A zoom is made in the first 200 m of the water column in order to accentuate the distribution of the vertical grid in upper layer. Here the sigma coordinate parameters are $h_{cline} = 300$ m, $\theta_b = 6$, and $\theta_s = 5$

609 2.10 Advection schemes

610 The advection scheme is third-order upstream biased which reduces dispersion errors nec-
611 essary to maintain smoothness. This scheme enhances precision for a given grid resolution
612 (*Shchepetkin and McWilliams, 1998*). However, the implementation of the third-order
613 advection-diffusion scheme has lead to the development of spurious diapycnal mixing in σ -
614 coordinate models. A solution to this problem was addressed by *Marchesiello et al. (2009)*
615 and involves the split of advection and diffusion for the tracer equation, and the latter
616 appears as a biharmonic operator. This solution was implemented in our configuration to
617 preserve the low diffusion and dispersion benefits of the original scheme while maintaining
618 water mass characteristics.

619 **2.11 The ROMS model: summary**

620 ROMS is a free-surface, terrain-following σ -coordinate model with split-explicit time step-
621 ping. It integrates in time the primitive equations with state of the art numerical methods
622 and with Boussinesq and hydrostatic approximations. The advection scheme is third-
623 order upstream biased, which reduces dispersion errors, essentially enhancing precision for
624 a given grid resolution (*Shchepetkin and McWilliams, 1998*). The diffusive part of the
625 advection scheme is rotated along the isopycnal surfaces to avoid spurious diapycnal mix-
626 ing (*Marchesiello et al., 2009; Lemarié et al., 2012*). Adaptive open boundary conditions
627 combine outward radiation and nudging towards prescribed external boundary conditions
628 (*Marchesiello et al., 2001*). Additional eddy viscosity is added in the sponge layer at open
629 boundaries, where it increases smoothly over several grid points. Subgrid-scale vertical
630 mixing is governed by a non-local, K-profile parametrization (KPP) scheme (*Large et al.,*
631 *1994*). A quadratic bottom drag is applied.

632 Chapter 3

633 A new simulation of the Southern 634 Benguela Upwelling System

635 This chapter focuses on the description of the new eddy resolving simulation that is used
636 to study the along-shore connectivity in the SBUS. The latter is based on the analysis of
637 Lagrangian trajectories advected by oceanic currents produced by this simulation. Our
638 first goal is to build a robust hydrodynamical model with the major ingredients, *i.e.* a
639 realistic mesoscale eddy field, a realistic Agulhas leakage that brings warm waters off
640 the Benguela shelf and enhances the offshore turbulence, and a realistic wind field with
641 its full (seasonal+ intraseasonal + interannual) variability. To satisfy these needs, the
642 horizontal resolution of this simulation was chosen to be $1/15^\circ$ (7.5 km), over a domain that
643 encompasses large portions of the ocean adjacent to the whole South African continent.
644 In order to reproduce a "statistically" realistic representation of the Agulhas Current and
645 its retroflection, the domain was extended as far 54° S and also includes the southern part
646 part of the Mozambique Channel where the Agulhas Current initiates. Throughout this
647 manuscript, this simulation will be thereafter considered as our reference simulation. It
648 will be used in all subsequent chapters, either directly in chapter 4 to study the physical
649 processes impacting the seasonal cycle of the along-shore connectivity, or indirectly in
650 chapter 5 and 6 to provide lateral boundary conditions for higher resolution nested models.

651 A monthly climatology of heat, fresh-water and momentum fluxes will be used at the
652 lateral boundaries, and 22 years of 6-hourly atmospheric conditions (winds, moisture, tem-

653 perature, infrared and short wave radiations) will be used as surface boundary conditions.
654 As a result, the inter-annual variability produced by this simulation over the Benguela
655 region will come from the local forcing and the turbulent eddy field, not from the lat-
656 eral boundaries. Therefore, the influence of remote dynamical features such as equatorial
657 coastal Kelvin waves will be neglected. However, the contribution of remote forcing to
658 the local dynamics is much less important in the Southern Benguela than in the northern
659 Benguela and north of the Angola-Benguela front (*Illig et al.*, 2018). Our priority is to
660 focus on the role dynamical processes occurring locally, or nearby like for example eddies
661 generated during the Agulhas retroflexion.

662 The chapter is constructed as follows. Section 3.1 provides a description of model
663 reanalysis, in-situ and satellite data sets used either to build and force our simulation, or
664 to compare the results of our simulation against. The configuration itself and its modelled
665 oceanic circulation is described in section 3.2.

666 **3.1 The regional model configuration developed: BENG15**

667 A regional configuration of the ocean surrounding the Southern African region, includ-
668 ing the Mozambique Channel, the Agulhas retroflexion, and the Benguela, is developed
669 specifically to study ocean circulation patterns of the SBUS. It is composed of a $1/15^\circ$ (
670 ~ 7.5 km) grid resolution extending from 54.5°S to 4.75°S and 2.5°W to 55.16°E (Fig.
671 3.1) and has 909×866 grid points. The vertical resolution consists of 100 vertical terrain-
672 following levels (sigma coordinate) that are stretched to obtain a vertical resolution from
673 1 m in the surface layers to 100 m in the bottom layers. The vertical grid uses stretching
674 surface and bottom parameters of $h_{cline} = 300$ m, $\theta_b = 6$, and $\theta_s = 5$.

675 **3.1.1 Bathymetry**

676 The bottom bathymetry used in this configuration is derived from the Shuttle Radar
677 Topography Mission (SRTM 30 plus) product (available at <http://topex.ucsd.edu/WWW.html/srtm30.plus.html>) based on the 1-min *Smith and Sandwell* (1997) global datasets
678 and higher-resolution surveys when available. It is interpolated on the model grid. To
679 avoid pressure gradient errors induced by terrain-following (sigma) coordinates in shallow
680

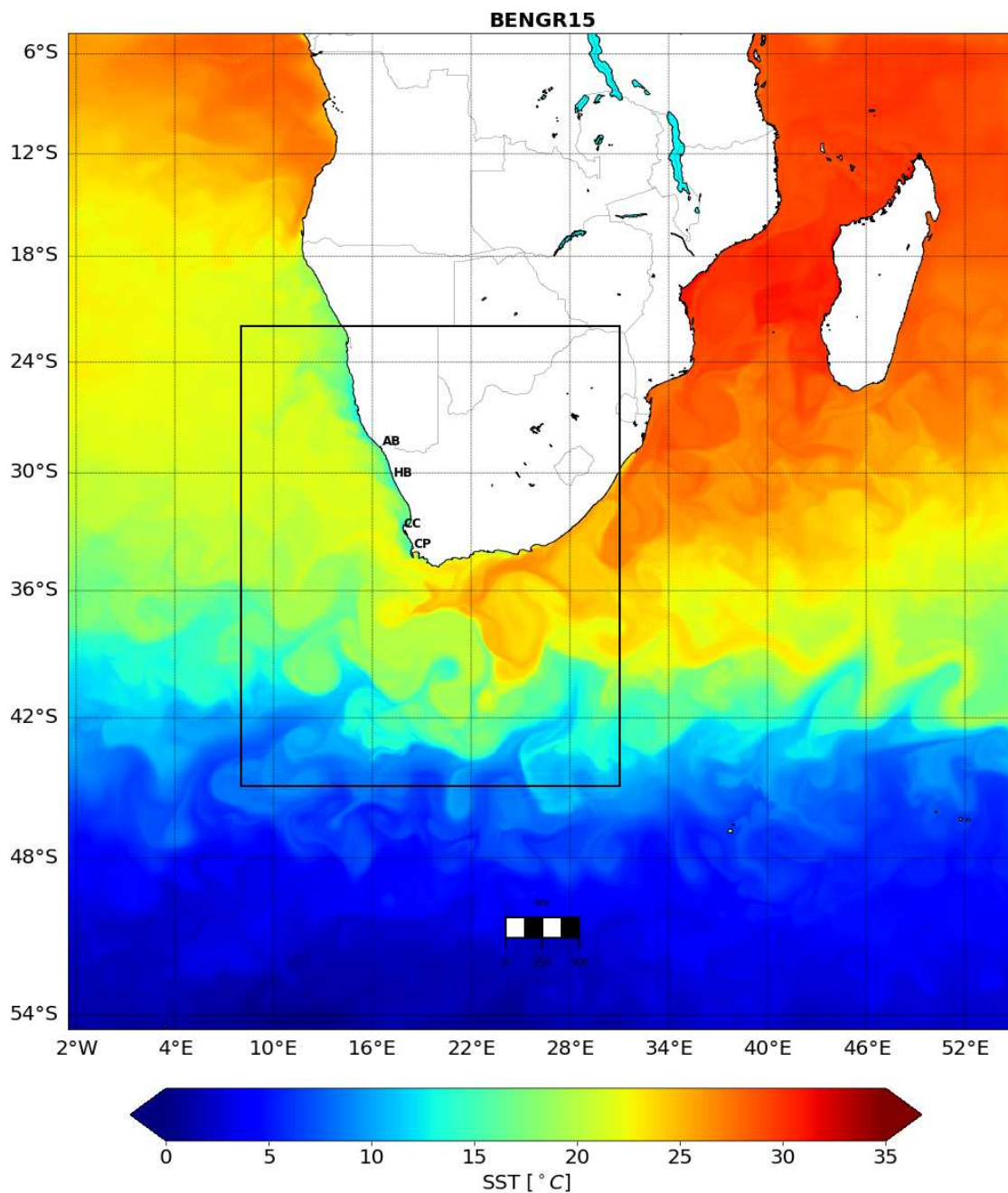


Figure 3.1: Model year 2005, January average snapshot of sea surface temperature (SST) for the Benguela upwelling system configuration ("BENGR15"). The domain encompasses the region covering 2.5°W - 55.16°E and 54.5°S - 4.75°S. The box represents an extracted region over the southern Benguela (45.1°S to 22.1°S and from 5.0°E - 34.9°E) used in our study.

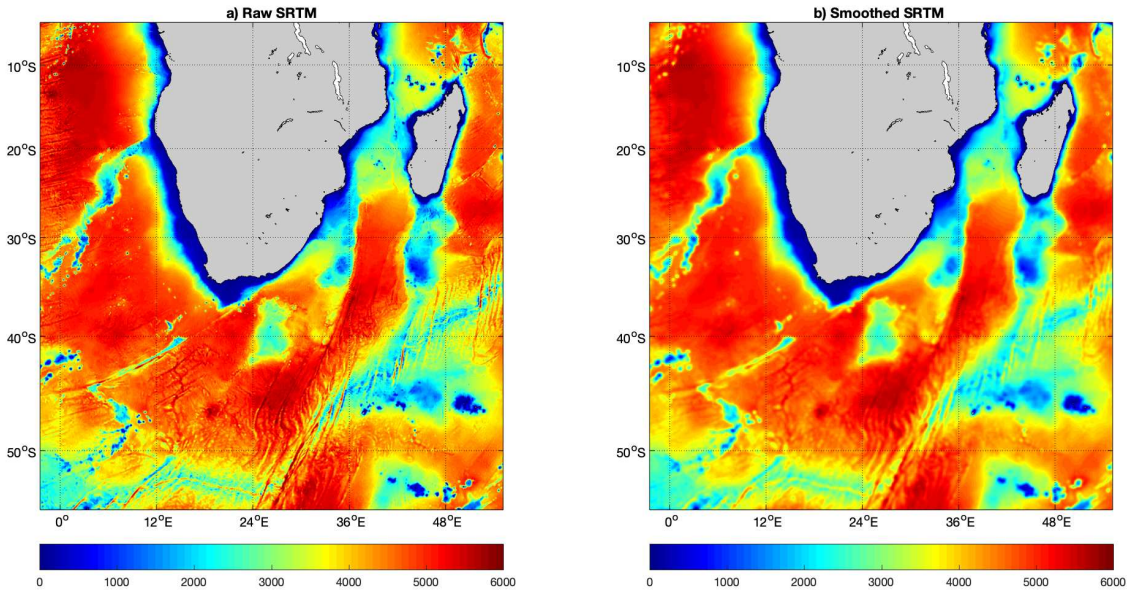


Figure 3.2: a) The original raw SRTM 30' bathymetry extracted over the BENG15 domain. b) The bathymetry of the BENG15 model configuration obtained after smoothing of the raw SRTM 30' bathymetry. c) The difference between raw SRTM 30' and smoothed BENG15 bathymetry

681 regions with steep bathymetry, the model bottom topography is smoothed where the
682 steepness of the topography ($r = \frac{\delta h}{2h}$) exceeded 0.2 (*Beckmann and Haidvogel, 1993*).
683 When the depth is shallower than 10 m, it is reset to 10 m.

684 A comparison between the smoothed bathymetry interpolated on the model grid and the
685 original SRTM30' bathymetry is presented on Fig. 3.2a,b. Despite some obvious smoothing
686 of some small scale bathymetric features (Fig. 3.3), one sees that the main topographic
687 features of the region are reproduced in the BENG15 topography. On the eastern flank
688 of South Africa, where the Agulhas Current flows southward, the shelf is very narrow. It
689 widens at the tip of the African Continent to form a large plateau, the Agulhas Bank, that
690 extends south, up to 400 km offshore at the longitude of Cape Agulhas. On the South-East
691 Atlantic side, the Agulhas ridge delimits the entrance of the South Atlantic. It stretches
692 along the South-West/North-East direction. It forms the southern boundary of the Cape
693 basin, an abyssal plain that lies offshore the SBUS and extends as far north as 20°S.
694 Some isolated seamounts are present on its eastern flank. In between these seamounts,
695 and the South-African west coast, there is a narrow passage that allows Agulhas leakage
696 into the South Atlantic. At 20° S, the Walvis Ridge forms the northern boundary of the

697 Cape Basin. It also stretches along the South-West/North-East direction. The SBUS is
698 characterised by a narrow shelf between Cape Point and Cape Columbine. The latter
widens slightly north of St Helena Bay.

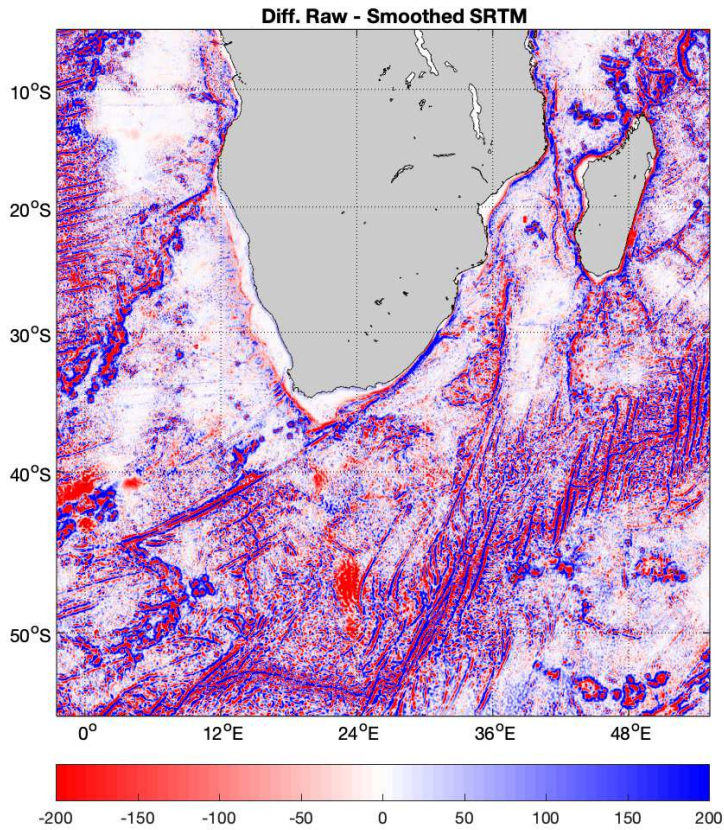


Figure 3.3: The difference between raw SRTM 30' and smoothed BENG15 bathymetry (m).

699

700 3.1.2 Atmospheric surface forcing

701 In EBUS, the along-shore wind forcing is the main ingredient that forces the circulation.
702 It is therefore essential to force our model with a wind product that reproduces the main
703 observed patterns of the wind forcing; i.e. its seasonal cycle and its spatial variability. A
704 good representation of the wind stress curl is of particular importance because it relates
705 dynamically to the Ekman pumping and can therefore enhance upwelling (*Colas et al.*,
706 2011). It also plays a role on the dynamics of the poleward undercurrent (*Capet*, 2004).
707 In EBUS, it is not rare to observe some intensification of the wind forcing around capes
708 that coincide with enhanced upwelling cells.

At the surface, atmospheric forcing (heat, fresh water and momentum fluxes) are calcu-

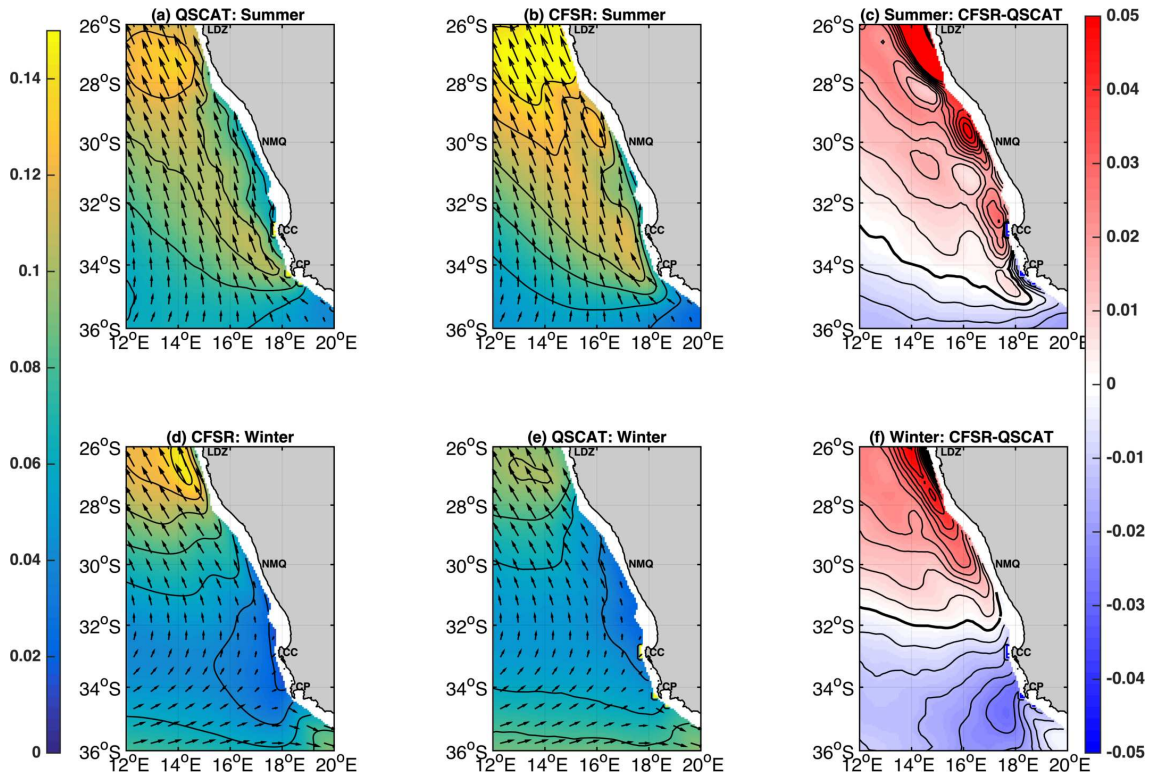


Figure 3.4: Summer and winter mean meridional wind stress from (a,d) QuikSCAT, (b,e) CFSR (colors, in N m^{-2} , contour interval: 0.02 N m^{-2}) and (c,f) their differences (reanalysis - satellite winds) for the 2000-2008 period. Positive (negative) values in the difference maps are regions where the reanalysis is stronger (weaker) than satellite. The bold black line represent the zero difference and the contour interval is 0.005 N m^{-2} .

709

710 lated using bulk formulae (Fairall et al., 1996) from the NCEP Climate Forecast System
 711 Reanalysis (CFSR) with a 6-hourly sampling and a spatial resolution of 0.3° (Saha, 2010).
 712 CFSR is a global coupled ocean-atmosphere-land-sea ice modelling system that uses the
 713 NCEP Global Data Assimilation System, and it is available over 31 years from 1979 to the
 714 end of 2017. To evaluate the CFSR reanalysis in SBUS, the wind stresses and wind-stress
 715 curl computed using the bulk formule are compared to wind stresses observed from the
 716 NASA scatterometer QuikSCAT satellite. QuikSCAT wind used here are homogeneous
 717 temporal series of daily mean wind stress fields, on a global $0.5^\circ \times 0.5^\circ$ resolution grids,
 718 generated from L2B product distributed by the French ERS Processing and Archiving
 719 Facility CERSAT¹. It is worth noting that near-coastal regions QuikSCAT wind stress
 720 data are masked due to land contamination and are not available within 25 km off the

¹<http://www.ifremer.fr/cersat/en/data/overview/gridded/mwfsqcat.html>

721 coast.

722 **Wind stress**

723 Comparing the summer and winter climatology of wind stress from the CFSR reanaly-
724 sis and QuickSCAT satellite observations show some good agreements between the two
725 over the SBUS (Figure 3.4). The summer and winter seasons are defined to encompass
726 the upwelling period from October to March, and the non-upwelling period from April to
727 September.

728 As expected, in summer, the direction of the alongshore wind stress is directed equator-
729 ward. The maximum wind stress core corresponds with the Cape Point upwelling cell at
730 34°S and the Lüderitz cell at 26°S (*Hutchings et al.*, 2009, *Weeks et al.*, 2006) (Fig. 3.4 a,b).
731 In winter, the equatorward winds weaken and become predominately westerlies south of
732 Cape Point (Fig. 3.4 d,e) consistent with the northward movement of South Atlantic Cy-
733 clone. Favourable upwelling winds persist year-round at Lüderitz cell, but they are much
734 weaker in winter. Although CFSR winds seem to be overestimated in regions coinciding
735 with the major upwelling cells of the SBUS (Fig. 3.4c), the upwelling favourable winds are
736 reproduced with a good degree realism.

737 **Wind stress curl**

738 The seasonal distribution of the wind stress curl over the southern Benguela is presented
739 in Figure 3.5. There is an along-shore coastal band of negative (cyclonic) wind-stress curl
740 associated with the along-shore wind drop-off. The position and local maximum of the cy-
741 clonic wind stress curl varies seasonally according to the seasonal cycle of the atmospheric
742 forcing. It is maximum and moves poleward during summer months whereas it retreats
743 northward during winter months. It then becomes weaker south between Cape Point
744 and Cape Columbine. There are some areas where the along-shore wind stress is locally
745 enhanced, therefore enhancing the cyclonic wind stress curl (*Bakun and Nelson*, 1991).
746 In the SBUS, three of these specific regions can be distinguished clearly: between Cape
747 Columbine (33°S) and Cape Point (34°S), off Namaqua shelf (30°S) and nearby Lüderitz
748 (*Weeks et al.*, 2006). Although, the CFSR wind stress cyclonic curl appears stronger than

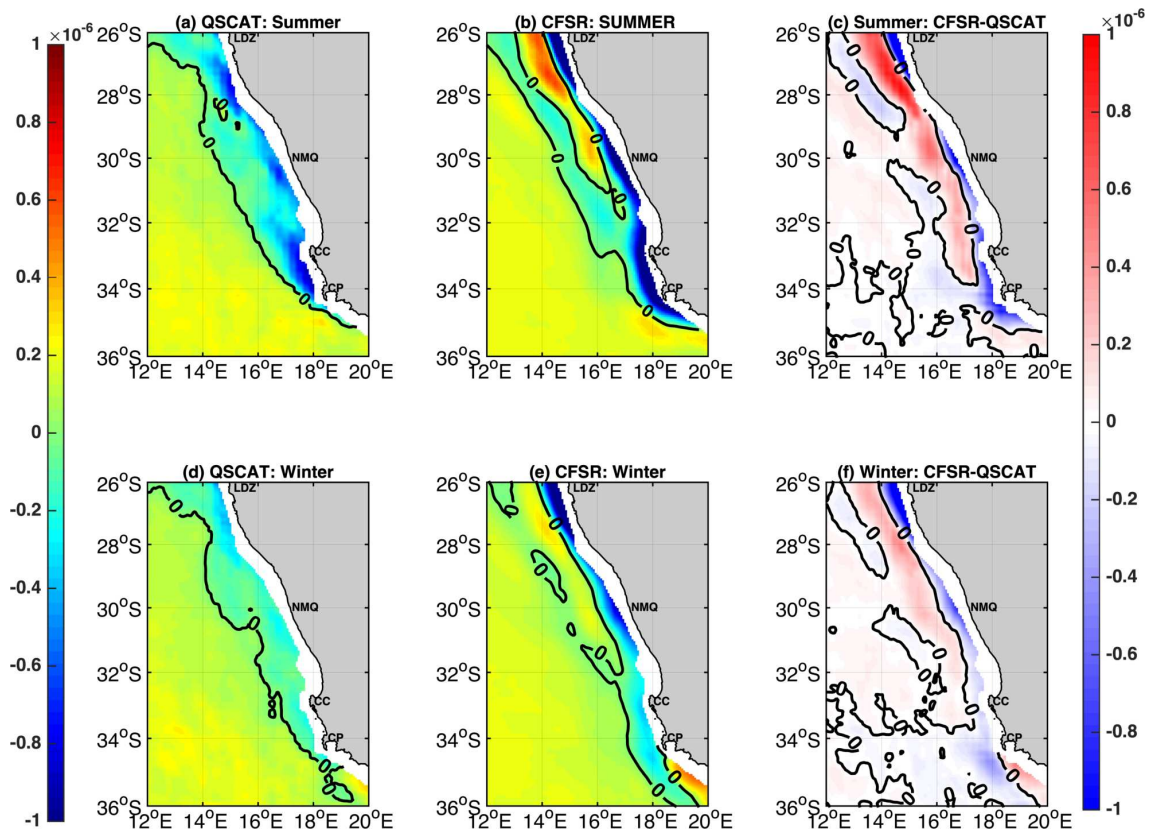


Figure 3.5: Summer and winter mean wind stress curl from (a,d) QuikSCAT and (b,e) CFSR (c,f) their differences (reanalysis - satellite wind stress curl) (colors, in Nm^{-3}). Cyclonic (anticyclonic) curl is negative (positive) and line bold black line represent the region of zero wind stress curl.

749 the one derived from QuickSCAT observations (Fig. 3.5c,f), the three upwelling cells are
750 consistently represented and match with the observations.

751 **3.1.3 Lateral oceanic forcing**

752 The conditions for temperature, salinity, horizontal velocity and sea level at the lateral
753 boundaries were provided by a monthly seasonal climatology derived from the SODA
754 2.3.2 reanalysis. SODA is a global ocean model with a horizontal resolution of 0.25° in
755 longitude and 0.4° in latitude. It has 40 vertical levels with a 10-m spacing near the surface
756 (*Carton and Giese, 2008*). This reanalysis covers the period 1958-2005 and assimilates
757 observational data in a general circulation model based on the Parallel Ocean Program
758 (*Smith et al., 1992*). The ROMS model was initialised from January 1979 using 3-D
759 current, temperature, salinity and SSH derived from SODA climatology with a 10-year
760 spin-up.

761 The SODA reanalysis is evaluated with World Ocean Atlas 13 (WOA13). The World
762 Ocean Atlas 2013 provides objectively analysed fields of temperature, salinity as well as
763 several nutrients. The spatial resolution of these data are from 1° to 0.25° globally, and the
764 data is distributed between the surface and 5500 m (*Locarnini et al., 2013*). In contrast to
765 the previous WOA09 (2009) that consisted of 32 depth levels, the WOA13 has increased
766 vertical resolution of 102 levels. The data are provided at a temporal resolution of annual,
767 seasonal or monthly climatologies. WOA13 is used to compare temperature sections based
768 on SODA open boundary conditions data used in the model.

769 The SODA vertical structure of temperature taken at the southern boundary of the
770 domain during summer and winter is examined and compared with the one obtained from
771 WOA13. The thermal gradient in the surface layers is stronger in summer (Fig. 3.6a,b)
772 indicative of higher stratification than in the winter (Fig. 3.6c,d) and the layer of temper-
773 atures $>16^\circ\text{C}$ is thicker in winter. This layer is warmer ($\sim 18^\circ\text{C}$) towards the east, near
774 the Agulhas Current region in WOA13 during this period. The seasonal vertical position
775 of the 16°C isotherm and the V-shaped structure of the isotherms are consistent in both
776 WOA13 and SODA. Since we apply SODA at the boundaries ~ 1500 - 2000 km away (see
777 full domain in Fig. 3.1) from the interior of the domain, the discrepancies, although minor,

778 will have reduced effect on the ocean model solution in the SBUS region. Hence we are
 779 confident with using SODA at the lateral boundaries.

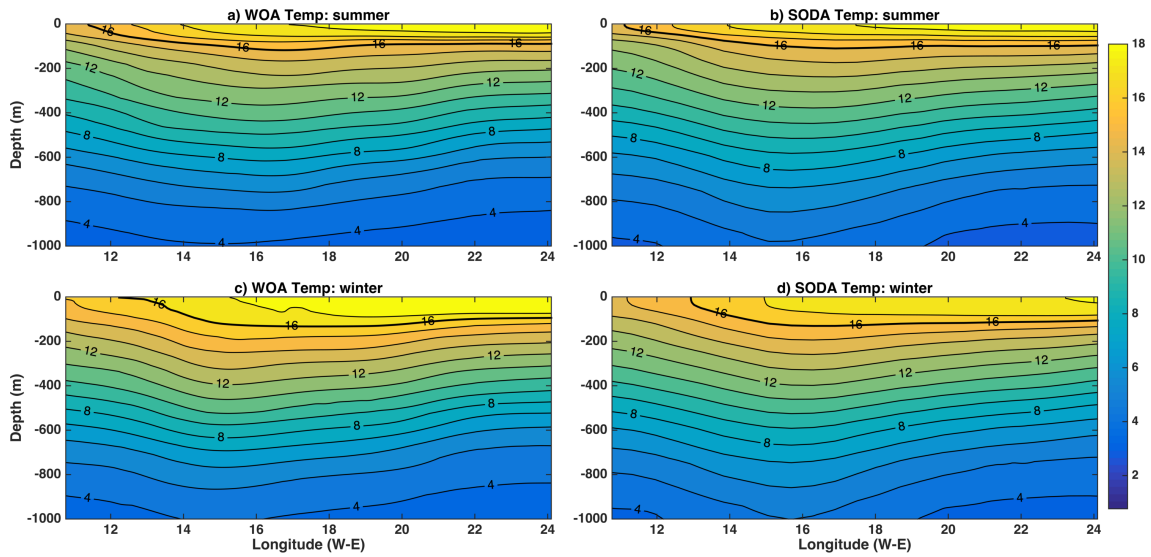


Figure 3.6: Summer and winter mean cross section of temperature taken from the southern boundary: (a,c) WOA13 and (b,d) SODA open boundary conditions with 1°C contour interval. The thick line represents the 16°C isotherm.

780 3.1.4 Model spin-up

781 The BENGR15 simulation is initialised with the January 1979 conditions given by the
 782 SODA reanalysis, and it is run for a total of 33 years (1979-2011). A 10-year spin-
 783 up is carried out for the model to reach a statistical equilibrium. To minimise storage
 784 requirements, we only downloaded model outputs from model year 11 to year 33 (1989 –
 785 2010). Therefore the diagnostic analysis presented in Figure 3.7 was done on the 1989 –
 786 2010 output variables averaged and stored every three days. Figure 3.7b,c shows that the
 787 change in kinetic energy is minimal and it fluctuates around its mean, which is indicative
 788 that the model has reached a statistical equilibrium.

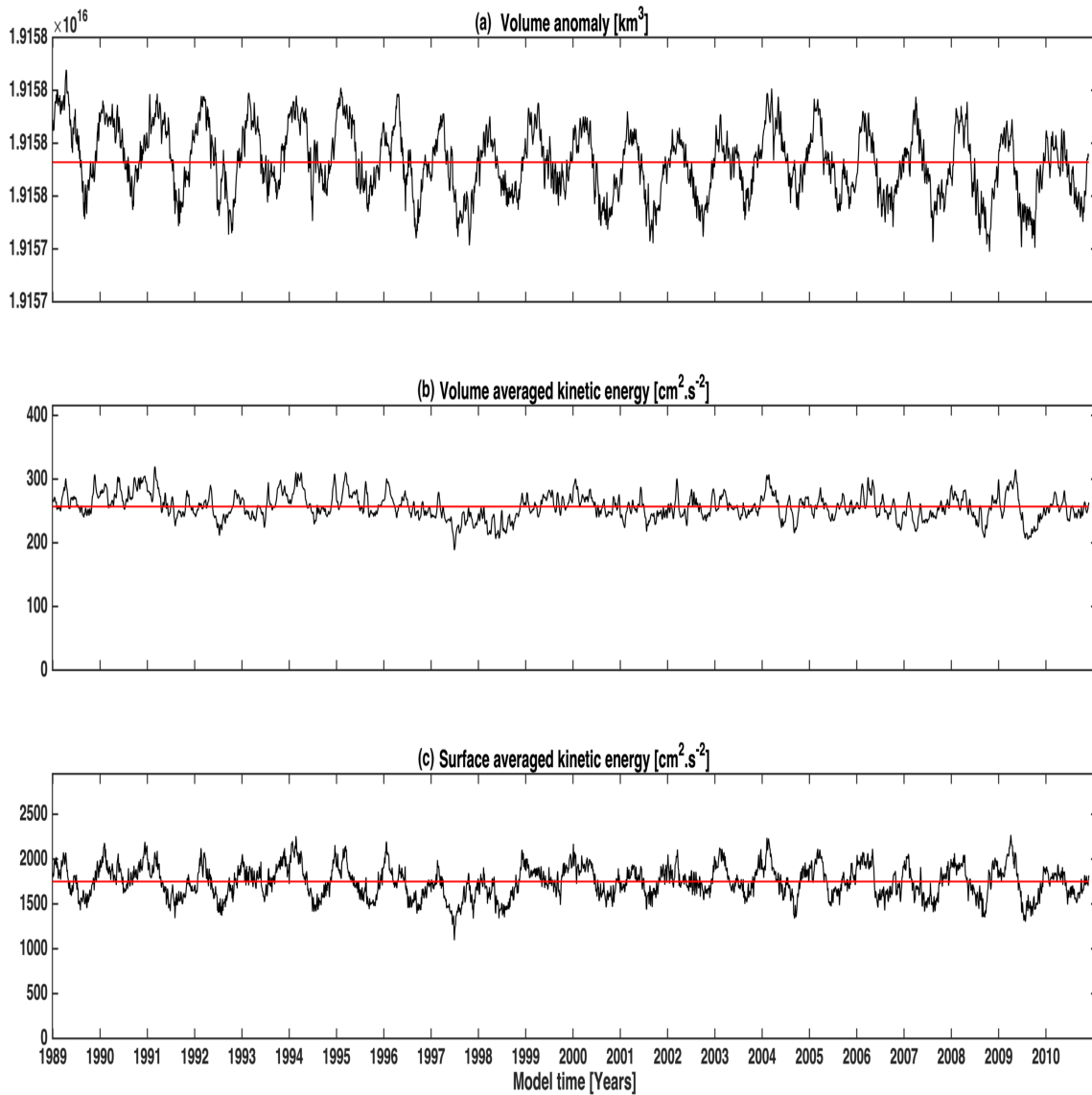


Figure 3.7: 3-day mean volume anomaly, volume averaged kinetic energy and surface averaged kinetic energy for the 22 years (1989-2010) BENGR15 simulation. The red horizontal line represents the mean value of the data.

789 3.1.5 Summary of the model configuration BENGR15

790 Parametrizations and main characteristics of the BENGR15 model configurations are pre-
 791 sented Table Tab. 3.1 below.

Parameters and Unit	Values
Horizontal grid size ($N_x N_y$)	$N_x = 909$ et $N_y = 866$
Number of vertical levels N_z	$N_z = 100$
Vertical resolution refinement σ -coordinate parameter at surface [$0 < \theta < 8$]	$\theta = 6$
Vertical resolution refinement σ -coordinate parameter at depth $0 \leq b \leq 1$	$b = 5$
Minimal depth h_{min} en m	$h_{min} = 10$
Thick of the maximum refinement layer	$h_c = 300$ m
Slow 3D-barocline time step (s) $0 < \Delta t_{bc} < \min(\frac{N_{bt} L_x}{N_x \sqrt{gh_{max}}}, \frac{L_x^2}{N_x^2 A_h^{sp}}, \frac{\Delta z_B}{r})$	$\Delta t_{bc} = 540$ s
Number of barotropic iterations during one barocline time step ($N_{bt} = \frac{\Delta t_{bc}}{\Delta t_{bt}}$)	$N_{bt} = 60s \rightarrow \Delta t_{bt} = 9s$
Bottom friction ($m^2 \cdot s^{-2}$)	Quadratic logarithmic parametrisation, $z_r = 0.001,$ $C_d^{min} = 1e^{-4}$ et $C_d^{max} = 2.5e^{-3}$
Nudging time (in days) for tracers	$\tau_{in}^T = 1$ & $\tau_{out}^T = 360$
Nudging time (in days) for momentum	$\tau_{in}^M = 3$ & $\tau_{out}^M = 360$
Sponge layer	Sponge layer width $L^{sp} = 10$ grid points Viscosity $K_M h^{sp} = XX$ [automatic computation] Diffusivity $K_T h^{sp} = XX$ ”
Bathymetry	SRTM30' + smoothing at $r = \frac{\delta h}{2h} \leq 0.2$
Surface Atmospheric forcing	Bulk formulation <i>Fairall et al.</i> (1996) based on the 6-hourly CFSR atmospheric reanalysis
Lateral Oceanic forcing	Initial and open boundary conditions computed from monthly climatology of the oceanic reanalysis SODA 2.3.2
Outputs frequency	Averaged outputs every 3 days
Period cover by the simulation	1979-2011 (33 years, spin-up over 1979-1989), analysis period of 23 years

Table 3.1: BENGRI15 configuration, at 7.5 km spatial resolution main parameters

792 **3.2 Model evaluation**

793 In order to study the variability of passive particles' transport, it is essential that our model
794 reproduces the seasonal cycle of the main currents within the region, before analysing
795 intraseasonal and interannual time scales.

796 **3.2.1 Mean Surface Circulation features**

797 **Surface Eddy Kinetic Energy (EKE)**

To estimate mesoscale activity in the upper-ocean, surface geostrophic eddy kinetic energy (EKE) is computed using geostrophic velocities derived from sea level anomaly (SLA). In the model, sea level anomaly (η') is computed by subtracting the monthly climatology sea surface height ($\bar{\eta}$) from the 3-daily sea surface height model outputs to remove the seasonal cycle. The SLA (η') is then used to calculate the non-seasonal surface geostrophic velocity anomalies as follows:

$$u'_g = -\frac{g}{f} \frac{\partial \eta'}{\partial y} \quad (3.1a)$$

$$v'_g = -\frac{g}{f} \frac{\partial \eta'}{\partial x} \quad (3.1b)$$

798 where g is the acceleration due to gravity (9.80 m s^{-2}), f is the Coriolis parameter, x and
799 y are the longitude and latitude positions, respectively.

800 The u'_g and v'_g are then used to calculate the non-seasonal surface geostrophic EKE:

$$EKE = \frac{1}{2}(u'^2_g + v'^2_g) \quad (3.2)$$

801 Regarding the observations, the delayed-time absolute dynamic topography (ADT), sea
802 level anomalies (SLA) and geostrophic velocities obtained from a gridded data product pro-
803 duced by DUACS are used to evaluate simulated surface geostrophic flows and eddy kinetic
804 energy. AVISO supplies the Data Unification and Altimeter Combination System (DU-
805 ACS) gridded product (CLS, Toulouse, France, available at <http://www.aviso.altimetry.fr/duacs/>)
806 from 14 October 1992 to 27 December 2001, with a nominal resolution of $1/3^\circ$ every
807 week. These measurements are obtained from merged TOPEX, POSEIDON, ERS and

808 Jason gridded data (Ducet et al., 2000).

809 Figure 3.8 shows a comparison between simulated, and altimetry derived EKE. Both
810 the simulated spatial pattern of the mean surface large-scale circulation and its variability
811 compare well with the EKE derived from satellite altimetry. In general, the most sig-
812 nificant EKE values ($>2000 \text{ cm}^{-2} \text{ s}^{-2}$) coincide with the energetic Agulhas Current, the
813 Agulhas retroflexion and the return current. The position of the Agulhas retroflexion
814 in the model (Figure. 3.8b) also appears to agree with satellite data. This retroflec-
815 tion drives an intense mesoscale activity that leaks into the south-eastern Atlantic Ocean
816 (Veitch et al., 2006) reflected by high values of EKE ($\sim 300\text{-}500 \text{ cm}^{-2} \text{ s}^{-2}$) in the offshore
817 side of the upwelling region between 36°S and 33°S . However, the simulated and observed
818 EKE are qualitatively similar in their spatial patterns but not in magnitude, as the model
819 underestimates EKE values in the Agulhas retroflexion. It is important to note that the
820 simulated coastal geostrophic EKE minima off the west coast are not captured in the
821 altimetry due to the incapability of the satellite to retrieve data close to the coast.

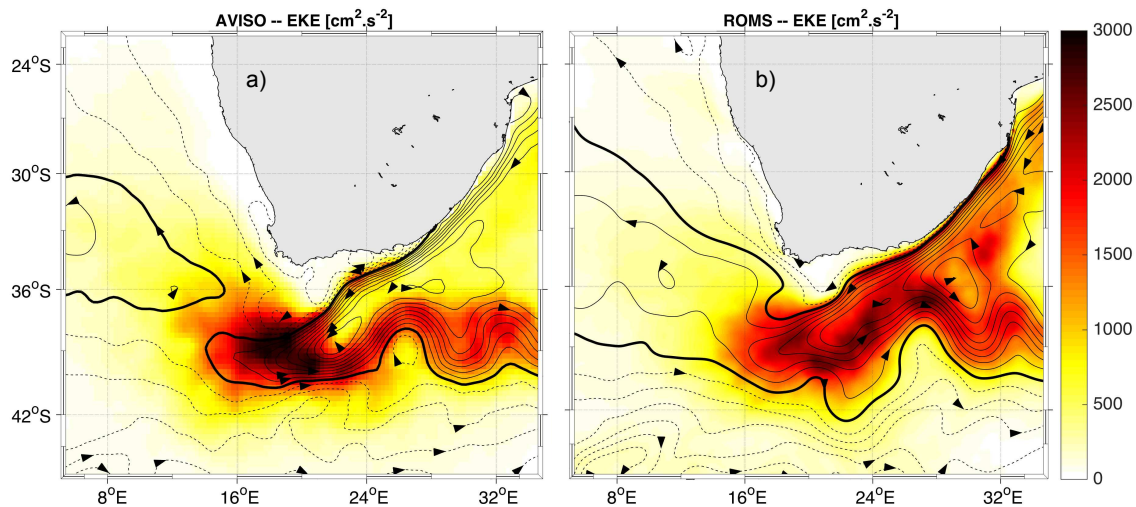


Figure 3.8: Maps of mean of the geostrophic surface eddy kinetic energy: a) computed from satellite (AVISO) data for the period between 1993 to 2015 and b) model-derived for the 1989-2010 period, both in $\text{cm}^{-2} \text{ s}^{-2}$ (color). The contours represent the mean SSH, with a contour interval of 10 cm. The overlaid arrows indicate the sense of associated geostrophic currents.

822 Sea surface heights from the model and altimetry show that the general climatological
823 characteristics of the large-scale ocean circulation are well reproduced in our simulation
824 (Fig. 3.8). The complex and chaotic dynamics of the Agulhas Current and its retroflexion,

825 that in turn leads the leakage of warm Indian ocean waters into the Atlantic, is well
 826 represented. The Agulhas Current flows south-westward following the east coast of South
 827 Africa. When it passes the tip of the Agulhas Bank, it abruptly veers eastward and forms
 828 the Agulhas return Current. Although the simulated retroreflection tends to occur 100 km
 829 east from the observed position, the levels of mean eddy kinetic energy (EKE) found in
 830 the model and observations are similarly high within this region. The retroreflection is
 831 in fact accompanied by the shedding of large anticyclonic eddies and filaments, some of
 832 which enter into the South Atlantic Ocean and form the Agulhas leakage (*Loveday et al.*,
 833 2014). The signature of this leakage in the form of a tongue of high EKE values, spreading
 834 northwestward into the Atlantic, can be seen in the model as in the observations. Finally,
 835 the long term average of this mesoscale eddy field propagating northwestward into the
 836 South Atlantic results in the eastern branch of the South Atlantic subtropical gyre.

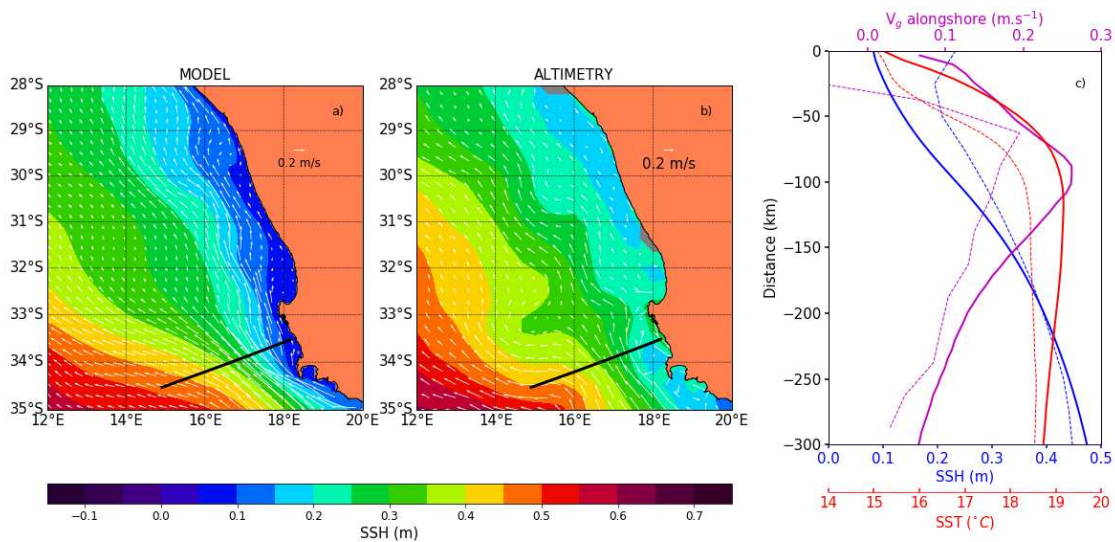


Figure 3.9: Annual average of sea surface height (SSH) with corresponding surface geostrophic currents superimposed: a) model outputs averaged over 1989-2010; b) satellite absolute dynamic topography (ADT) data averaged over 1993-2010; c) across-shore profiles of model (solid) and satellite (dashed) annual averages of SST (red), ADT (blue) and along-shore geostrophic current (magenta) off Grotto Bay (black line in a) and b). The plotted model SSH was offset by +0.407 m so that the model and satellite spatially average fields matched over the domain of interest.

837 Zooming in Fig. 3.8 over the SBUS to get a closer view of the along-shore current
 838 shows that the modelled surface geostrophic currents display similar characteristics to the
 839 observations (Fig.3.9 a,b). The eastern branch of the South Atlantic subtropical gyre

840 flows northwestward. There is an intensified inshore branch, between the 200 and 500 m
 841 isobaths, the Benguela Jet, with average velocities reaching 20 cm s^{-1} in the model, and
 842 slightly less (15 cm s^{-1}) in the observations (Fig.3.9 c). However, this bias could be linked
 843 to the coarse resolution of altimetry near the coast.

844 Sea surface temperature (SST)

845 The comparison of the modelled and infrared ODYSEA (*Piollé and Autret, 2011*) sea
 846 surface temperature (SST) shows that the cooling of coastal waters in summer consistently
 847 occurs in the model and observations (Fig. 3.10). There is an inshore warm bias of roughly
 848 1.5°C in the simulation, but such bias in regional models of EBUS is quite common ².

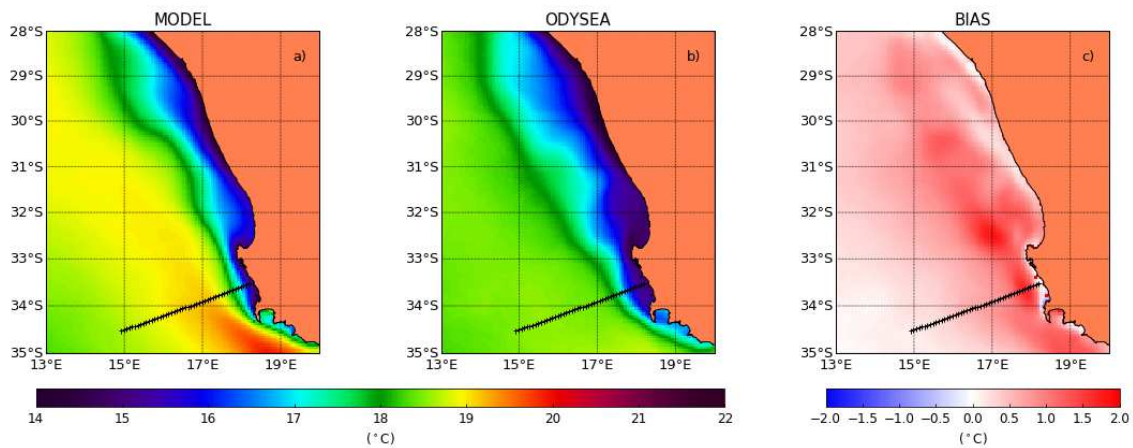


Figure 3.10: Annual average of Sea Surface Temperature (SST): a) satellite derived infrared data averaged over 2010-2017 (ODYSSEA); b) model data averaged over 1989-2010; c) Annual average of the SST bias ($SST_{\text{model}} - SST_{\text{ODYSEA}}$).

849 A second satellite observed SST product, ESA-CCI SST, has been used to evaluate the
 850 model SST bias through the strong seasonal cycle encountered in the regions. CCI-SST is
 851 a daily satellite SST from the European Space Agency's Climate Change Initiative (CCI;
 852 <http://www.esa-sst-cci.org/>) launched by European Space Agency (ESA). The CCI
 853 data has a $\sim 0.05^\circ$ resolution and spans from 1991 to 2010 with a 0.1 K accuracy (*Hollmann*
 854 *et al.*, 2013). The daily data were then averaged to obtain monthly data.

855 In summer, SST varies from about 15°C at the coast to $21\text{-}22^\circ\text{C}$ offshore (Fig. 3.11a,b).
 856 Lower temperatures are observed in winter where the temperature drops down to $13\text{-}14^\circ\text{C}$

²Misrepresentation of surface heat fluxes, wind-stress and vertical mixing in the coastal region could partially explain this bias.

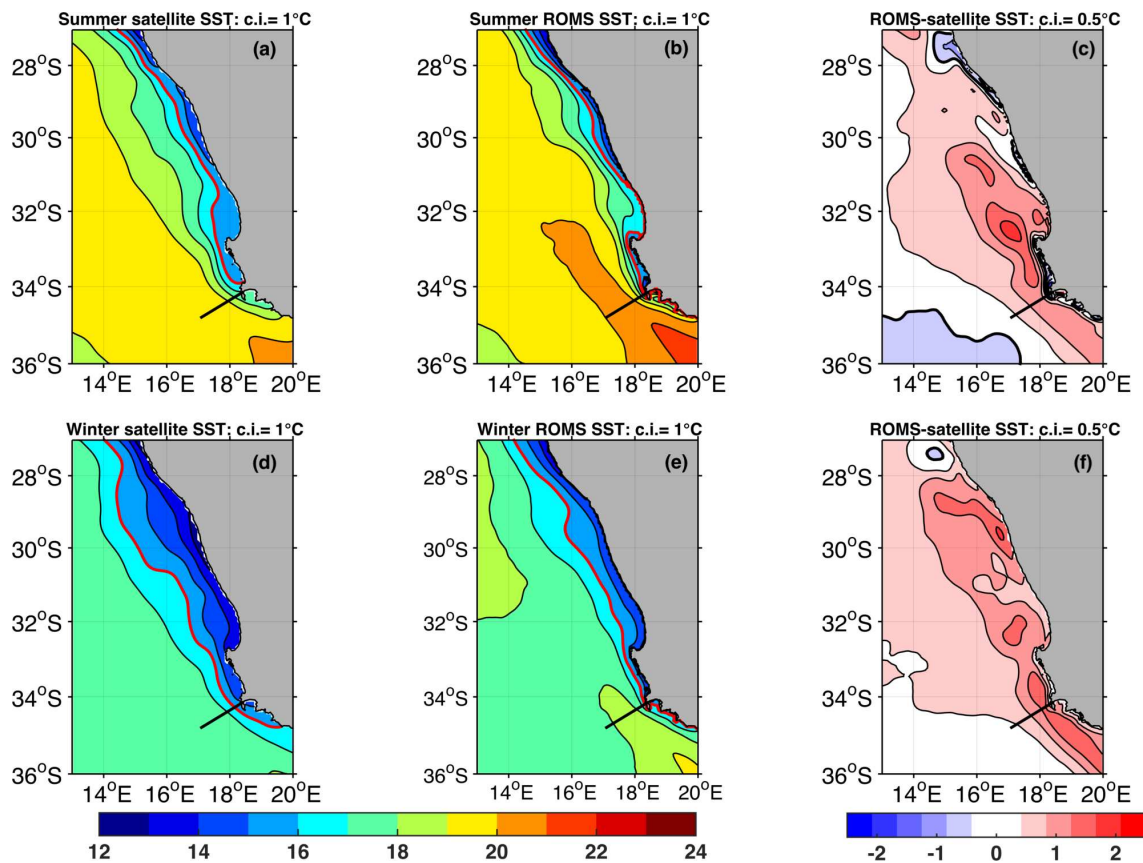


Figure 3.11: Climatology of the summer and winter SST obtained from (a,d) CCI satellite, (b,e) model and (c,f) difference (model - observation) for the 1992-2010 period. The 16°C isotherm is superimposed in red. Positive (negative) values in the difference map are regions where the model overestimate (underestimate) observed SST.

857 near the coast, and maximum temperatures of 16-17°C are found offshore (Fig. 3.11d,e).
858 Some notable differences are also observed. The cold coastal water bounded by the 16°C
859 isotherm (red line) appears to meander further offshore in the observations than that in
860 the model. The model is $\sim 1^\circ\text{C}$ warmer in the offshore region, and the warm bias increases
861 to 1-2°C in the coastal region (Fig. 3.11c,f). Its origin stems from the warm Agulhas
862 Current water leakage overestimation into the South Atlantic indicated by the warm SST
863 tongue coming from the southern boundary. This introduces a warm bias in our simulation
864 extending as far north as Cape Columbine, and this was similarly found in *Veitch et al.*
865 (2006).

866 **Zonal SST gradient**

867 Since the model overestimates temperatures both in the offshore region and coastal region,
868 we believe that the cross-shore SST gradient could still be correctly represented. The cross-
869 shore temperature gradient (ΔSST_{x_i}) can be used as a proxy for the density gradient that
870 drives the alongshore equatorward jet (*Veitch et al.*, 2017). This SST gradient (ΔSST_{x_i}
871 in $^\circ\text{C km}^{-1}$) is here calculated as a zonal difference between adjacent model grid points
872 or satellite data pixels as follows:

$$\Delta\text{SST}_{x_i} = \frac{\text{SST}_{(i+1)} - \text{SST}_{(i-1)}}{X_{(i+1)} - X_{(i-1)}} \quad (3.3)$$

873 where ΔX is the zonal distance in km between the neighbouring grid points. The seasonal
874 climatology of this SST gradient, based on the model and satellite data (CCI-SST), is
875 plotted on cross-shore distance versus time Hovmuller diagram at 34.14°S (Fig. 3.12). The
876 SST gradient has a marked seasonal cycle, with maximum gradient during the upwelling
877 season (October-March). The model is able to capture its seasonal pattern: it intensifies
878 in summer ($-0.01^\circ\text{C km}^{-1}$) and extends further offshore during the late upwelling season
879 (January to March). There is a good correspondence between the positions of the Benguela
880 Jet and the SST gradient as shown in *Veitch et al.* (2017). This equatorward geostrophic
881 jet is observed throughout the year between the 200 m and 500 m isobaths. The jet's
882 velocities can be greater than 0.45 m s^{-1} during the early upwelling season (October-
883 November). In the late upwelling season, the jet also moves offshore. This suggests that

the jet may be driven by the cross-shore SST gradient.

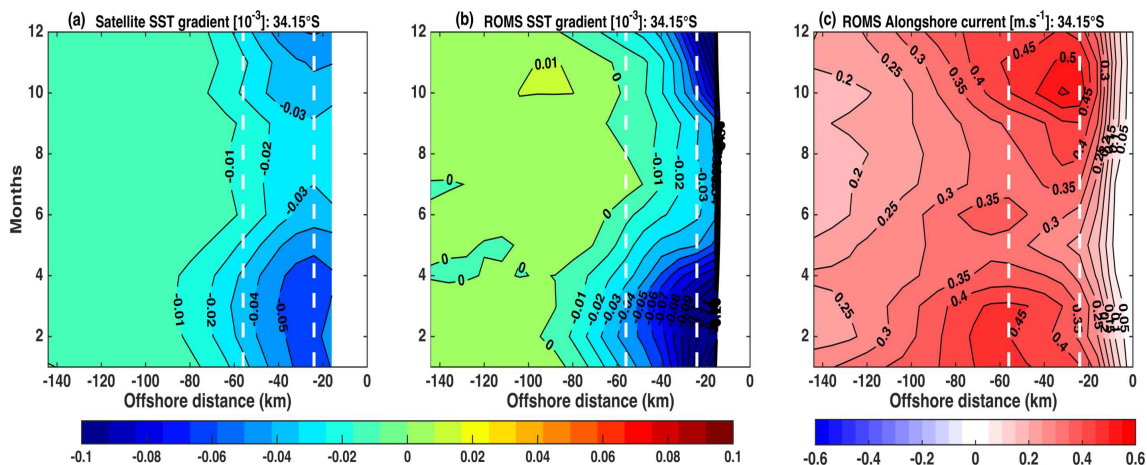


Figure 3.12: Space-time plot of climatological mean cross-shore SST gradient at 33.15°S for: (a) CCI-satellite, (b) model, both with an interval of 0.01°C . (c) : model alongshore current with a contour interval of 0.05 m s^{-1} . The 200 m and 500 m isobaths are shown as white dashed lines.

884

885 3.2.2 Hydrography: Vertical currents and temperature

886 The cross-shore section of the simulated and observed (WOA2013) alongshore geostrophic
 887 velocities taken at a transect off Grotto Bay (33.5°S), extending 300 km offshore is pre-
 888 sented in figure 3.13. The geostrophic velocities are calculated using the thermal wind
 889 relation using the 1000 m depth as a reference level from which to perform the vertical
 890 integration. Figure 3.13a,b shows a well-defined, broad and strong northward coastal jet
 891 corresponding to the Benguela jet that extends to the depth of 300 m. Its core is located
 892 about 120 km offshore, and its intensity weakens with depth.

893 A subsurface poleward undercurrent can be observed on the shelf break in agreement
 894 with current meter measurements of *Nelson* (1989). Poleward velocities can reach up to
 895 0.1 m s^{-1} during summer. A thermal warm bias reaching up to $1\text{-}2^{\circ}\text{C}$ can be observed in
 896 the model. However, this bias is homogeneous in its across-shore ($\simeq 200 \text{ km}$) and vertical
 897 ($\simeq 100 \text{ m}$) extension (Fig. 3.13c). Therefore, the surface layers across-shore density gradi-
 898 ents are expected to reasonably well reproduced in the simulation, and, consequently, the
 899 modelled alongshore geostrophic currents derived from the hydrography also match the
 900 observations.

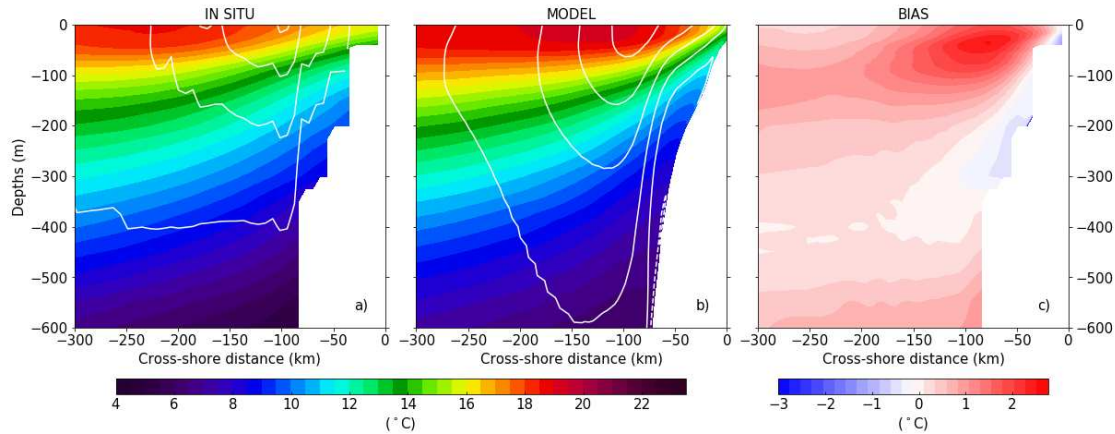


Figure 3.13: Temperature transect off Grotto Bay (solid black line on Fig. 3.10) with the along-shore component of the geostrophic current superimposed (white contours). a) World Ocean Atlas annual average (WOA2013); b) Same as in a) for the model; c) Temperature bias ($T_{\text{model}} - T_{\text{WOA2013}}$). Geostrophic velocities is built with a reference level at 1000 m depth. Contours are plotted every with 0.05 m s^{-1} in the range $[-0.05, 0.25] \text{ m s}^{-1}$. Negative values are dashed.

901 3.2.3 The near-shelf thermal stratification

902 The St Helena Bay Monitoring Line (SHBML) consists of 12 stations located in St Helena
 903 bay just offshore Elands Bay (32.30°S , 18.31°E) (Figure 3.14). Conductivity-Temperature-
 904 Depth (CTD) profiles go from the surface to a maximum depth of 1466 m, and stations
 905 have been monitored monthly by the Department of Environmental Affairs from April
 906 2000 to December 2011 (*Lamont et al.*, 2015). Seasonal climatology of these data provide
 907 a unique dataset to estimate the thermal structure of the model in the nearshore region.

908 Differences between model-derived and in-situ based temperature for the SHBML con-
 909 firm that there is a positive warm bias ($>3^{\circ}\text{C}$) in the upper 100 m in the offshore region
 910 (Fig. 3.15). Nevertheless, the modelled across-shore temperature variations are very simi-
 911 lar to the observed ones. Temperatures inside the bay are much cooler than in the offshore
 912 region, and the thermocline is shifted upward consistent with the presence of coastal up-
 913 welling.

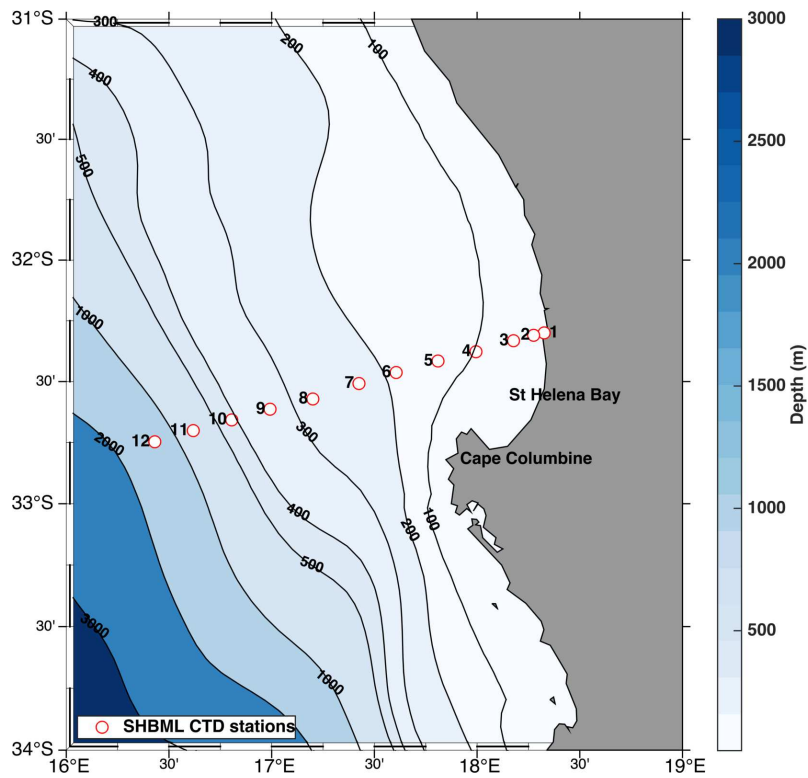


Figure 3.14: Model derived bathymetry (black lines in metres) in blue with 100 m contour interval for shallower region than 500 m and 500 m interval for region deeper than 500 m. The red markers show the locations of the 12 CTD stations of the St Helena Bay Monitoring Line (SHBML). Stations 1-6 are located within the bay region (coastal region), stations 8-10 located over the mid-shelf region and stations 11-12 further offshore in the open ocean.

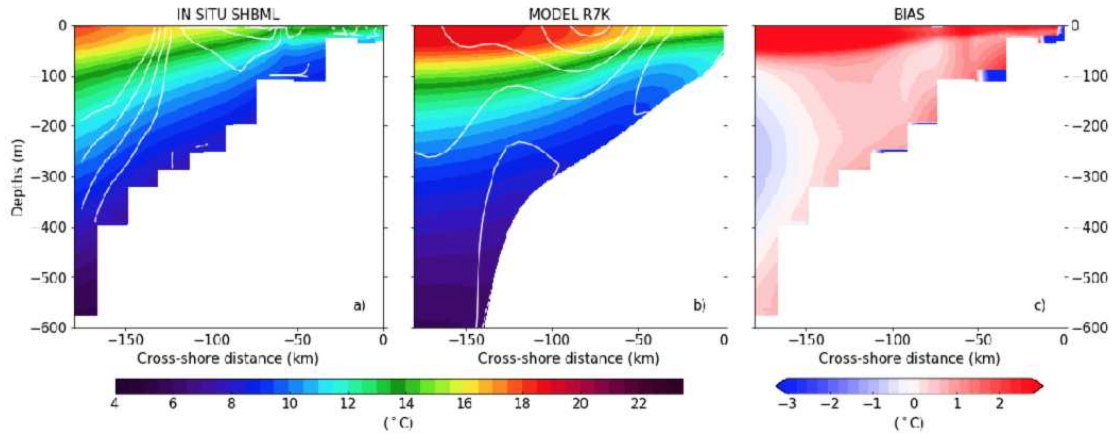


Figure 3.15: Temperature transect along the SHBML CTD stations (location of station is shown in Fig. 3.14) with the alongshore component of the geostrophic current superimposed (white contours). a) SHBML annual average ; b) Same as in a) for the model; c) Temperature bias ($T_{model} - T_{SHBML}$). Geostrophic velocities is built with a reference level at 1000 m depth. Contours are plotted every with 0.05 m s^{-1} in the range $[-0.05 \text{ } 0.25] \text{ m s}^{-1}$.

914 Summary of the model evaluation

915 The oceanic circulation of the SBUS is studied with a mesoscale-resolving interannual
 916 (1989-2010) simulation and evaluated with available satellite and in-situ data. Despite
 917 the warm bias, the results from the model evaluation show that the model is capable of
 918 reproducing many of the physical features typical of a coastal upwelling system:

- 919 • A surface northward density-driven current located on the shelf in between the 200 m
 920 and 500 m isobath, namely the Benguela Jet;
- 921 • A poleward undercurrent confined to the slope-edge in a 250-1000 m depth range.
- 922 • The presence of intense mesoscale field just off the shelf.

923 It is important to note that the evaluation of the coastal circulation is a challenge
 924 due to a limitation of available observational data. When available, the observational
 925 data have a coarse resolution, especially within the 50 km coastal band, and this is not
 926 adequate to resolve coastal processes. Consequently, some of the model bias found in
 927 the coastal region could be explained by the low resolution of the satellite measurements.
 928 Additionally, misrepresentation of surface heat fluxes, wind-stress and vertical mixing in
 929 the coastal area could partially explain some the warm coastal bias.

930 Chapter 4

931 Lagrangian pathways in the 932 southern Benguela upwelling 933 system

934 The content of this chapter is extracted from a paper entitled *Lagrangian pathways in the*
935 *southern Benguela* published in the Journal of Marine Systems in 2019 (*Ragoasha et al.*,
936 2019):

937 Ragoasha, N., Herbette, S., Cambon, G., Veitch, J., Reason, C., and Roy, C., 2019.
938 Lagrangian pathways in the southern Benguela upwelling system. *Journal of Marine*
939 *Systems*, 195, pp.50-66. Doi: <https://doi.org/10.1016/j.jmarsys.2019.03.008>.

940 4.1 Introduction

941 The effect of ocean currents on fish eggs and larvae during their journey from spawning
942 to nursery grounds in the Southern Benguela upwelling system is poorly understood. The
943 survival and successful transport of fish eggs and larvae, at early-stage, results from com-
944 plex biological and physical processes. This study focuses on the physical processes that
945 impact the connectivity between the western edge of the Agulhas Bank and the west coast
946 upwelling. A set of Lagrangian experiments are conducted with particles being released
947 within the top 100 m of the water column along an across-shore transect off Cape Point

948 (34° S). The Lagrangian experiments aim at addressing the following: (1) the identifica-
949 tion of the dominant pathways that ensure this connectivity; (2) their characteristics and
950 robustness; (3) the ocean dynamics that link to the alongshore transport success seasonal
951 variability. Results are based on the 7.5 km horizontal resolution reference simulation.

952

953 Results presented within this chapter are based on the reference mesoscale eddy re-
954 solving interannual (1989-2010) simulation of the region, with a horizontal resolution of
955 7.5 km. Unlike the climatologically forced model of *Veitch et al.* (2017), our ocean model
956 is forced by 6-hourly surface forcing (wind, heat and freshwater fluxes) over the 1989-2010
957 period. The use of a realistic, 22 years-long, eddy-resolving simulation that gives robust-
958 ness to the seasonal cycle that we computed and to our conclusions regarding the role of
959 the Benguela Jet, Ekman drift and mesoscale eddies in the advection of particles between
960 Cape Point and Cape Columbine.

961 A first step towards elucidating the role of the jet consists of identifying the different
962 physical processes that control the seasonal cycle of this northward alongshore connectivity
963 between Cape Point and St Helena Bay, just north of Cape Columbine. The Lagrangian
964 experiments aim at addressing the following: (1) the identification of the dominant path-
965 ways that ensure this connectivity; (2) their characteristics and robustness; (3) the ocean
966 dynamics that link to the alongshore transport success seasonal variability.

967 The Lagrangian experiments designed for the purpose of this study, are described in
968 section 4.2. Section 4.3 then characterises the seasonal cycle of the alongshore connectiv-
969 ity in terms of duration of travel and initial positions. The Lagrangian pathways taken
970 by particles are identified in section 4.4, while section 4.5 investigates the link between
971 the observed Lagrangian dispersal patterns and the across-shore structure of the ocean
972 circulation, mainly the alongshore currents and the mesoscale turbulent eddy field. Fi-
973 nally, a summary and discussion of our results highlight the complexity of the across-shore
974 structure of alongshore jet currents in the region.

975 **4.2 Lagrangian experimental set-up**

976 In our experiment, particles are released over the Sardine and Anchovy Recruitment Pro-
 977 gram (SARP) line, which is a cross-shore transect situated off Cape Peninsula. The section
 978 extends from $(34.15^{\circ}\text{S}, 17.55^{\circ}\text{E})$ and $(34^{\circ}\text{S}, 18.4^{\circ}\text{E})$ and is located between the coastal land
 979 and the offshore region defined by the 3000 m isobath. For each vertical profile, particles
 980 are randomly distributed over the first 100 m depth with a density of 1.125 particles per
 981 metre. This transect is considered as a transition zone separating the spawning area and
 982 the recruitment area for small pelagic fish (*Garavelli et al., 2012*).

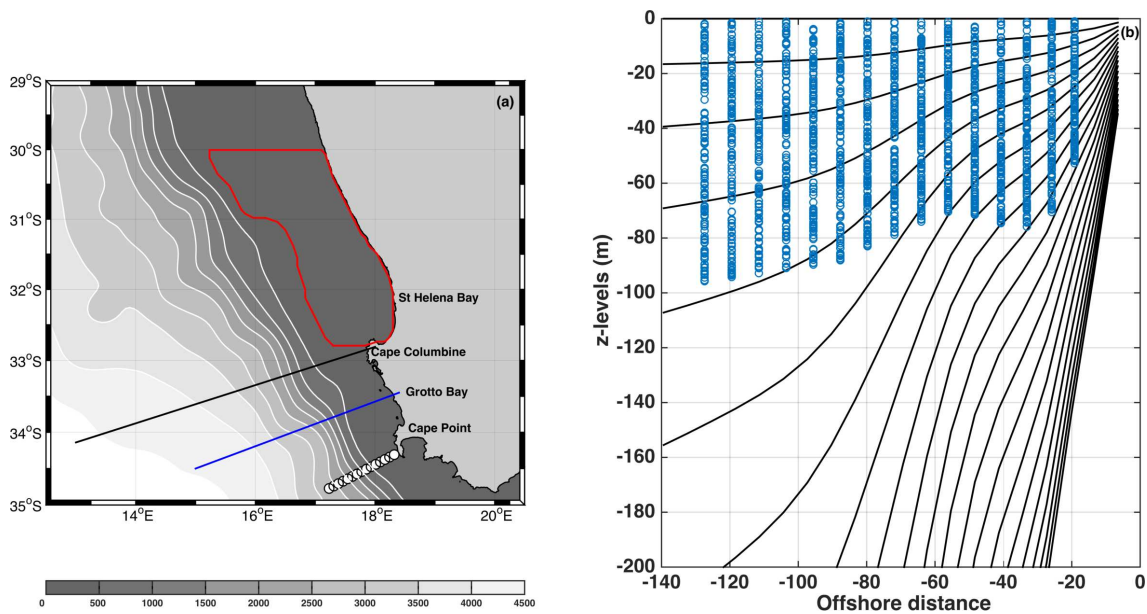


Figure 4.1: (a): Set-up of the Lagrangian experiments: particles are released over the SARP line (white circles), a cross-shore transect off Cape Point. They are counted as successfully transported particles if they reach, within 30 days, the St Helena Bay target area delineated by the red polygon. The bathymetry of the "BENGR15" model (in grey with white contours every 500 m) is also superimposed as well as two other cross-shore transects that particles will cross while being advected northward: the Grotto Bay line at 33.5°S (thick blue) and the Cape Columbine line at 33°S (thick black). (b): Initial distribution of particles over the SARP line for a given release event. σ layers are superimposed (black solid).

983 **4.2.1 The Lagrangian particle-tracking model**

984 In the present study, ROMS outputs of velocity fields are coupled to a Lagrangian particle-
 985 tracking model. The Lagrangian experiments are performed off-line based on an algorithm

986 called Pyticles developed by *Gula et al. (2014)*. Pyticles is a Python/Fortran hybrid paral-
 987 lelized code. The 4th order RungeKutta (RK4) scheme is used to integrate time between
 988 particle positions with the ROMS simulated velocity fields interpolated to the particle
 989 locations at each integration time step. RK4 method uses the interpolated velocity infor-
 990 mation at four increments between time steps t_n and $t_n + 1$ (*van Sebille et al., 2018*). This
 991 interpolation is necessary to account for the time interval differences between consecutive
 992 stored velocity field and the time step (Δt) required for particle displacement. The only
 993 dispersion process considered is the advective flow according to the hydrodynamic model
 994 with no additional diffusive terms implemented in Pyticles. Parameters such as the num-
 995 ber of particles released in time and in space as well as the values for the forcing frequency
 996 and time step can be defined. Particles can be released with a continuous injection or
 997 once-off release.

998 4.2.2 Statistical Lagrangian diagnostics

999 • Transport success (TS) diagnostics

1000 For each release event, transport success (TS_e) is given by the ratio of the number of
 1001 particles that have reached St Helena Bay (red polygon) target area (Fig. 4.1a) over
 1002 the total number of particles released (N_e) (*Huggett et al., 2003*). This polygon extends
 1003 between 32.8°S and 30°S and from the coast to 300 m isobath. Each particle was counted
 1004 only once during its first entry into the polygon at any time within their 30 days life span.

$$TS_e = \frac{n_e}{N_e}, \quad (4.1)$$

1005 where n_e is the number of particles entering St Helena Bay during event e . N_e , the number
 1006 of particles released during event e is kept constant during the Lagrangian experiment.
 1007 TS_e can be seen as a proxy for the SARP line/St Helena Bay connectivity.

1008 The monthly transport success TS_m is computed by averaging over the simulation all
 1009 events specific to each individual month:

$$TS_{m\{m=1,\dots,12\}} = \sum_{e(m)} TS_{e(m)} \times \frac{1}{N_{e(m)}}, \quad (4.2)$$

1010 were $e(m)$ is the event index chosen to correspond to all the release events occurred during
 1011 month m .

1012 For every monthly transport success, we estimate its standard deviation (σ_{TS_m}). This
 1013 could be used as an estimation for the amplitude of the interannual variability.

$$\sigma_{TS_m \{m=1, \dots, 12\}} = \sqrt{\sum_{e(m)} [TS_{e(m)} - TS_m]^2} \quad (4.3)$$

1014 • **Diagnostics of particle dispersion patterns**

1015 In order to get a general view of the routes taken by the released particles, maps of
 1016 particle distribution are plotted according to their age (given in days after their release).
 1017 The density of particles aged t_l within a given grid cell centred at x_i, y_j is calculated as
 1018 follows:

$$d(x_i, y_j, t_l) = \frac{n(x_i, y_j, t_l)_e}{n_p} = d_{ijl}, \quad (4.4)$$

1019 where $n(x_i, y_j, t_l)_e$ is the number of particles aged t_l found within the grid cell centred at
 1020 x_i, y_j , and n_p the total number of particles released. Here, grid cells correspond to the
 1021 horizontal grid of the ROMS model. The above definition insures a consistent normalisa-
 1022 tion of the particle distribution. The density would be one if all particles at the SARP
 1023 line would be advected within the grid cell when they are aged t_l . Note that this density
 1024 is calculated for some specific released events, here grouped monthly.

1025 Maps of the density distribution can be understood as plume of particle dispersion. The
 1026 geometric centre of the plume represents the mean position of particles aged t_l and is
 1027 computed as follows:

$$X_c, Y_c(t_l) = \sum_p X_p(t_l), Y_p(t_l) \times \frac{1}{n_p} = \sum d_{ijl}(X_i, Y_j) \quad (4.5)$$

1028 where X_p, Y_p are the position of a Lagrangian tagged p at an age t_l . n_p is the number of
 1029 particles released, and X_i, Y_j are the ROMS grid coordinates.

4.3 Characteristics of the alongshore connectivity

The following section intends to characterise the seasonal cycle of the alongshore connectivity between the SARP line and St Helena Bay in terms of: i) transport success, ii) time taken by successful particles to travel northward, iii) initial vertical and across-shore position of successful particles.

Fig. 4.2 plots the monthly climatology of the alongshore wind stress at the SARP line together with transport success TS_m and its standard deviation. Both monthly alongshore transport success and its standard deviation show a strong seasonal cycle, that does not strictly superimpose with the upwelling. Upwelling favourable winds prevail in austral spring and summer from October to March, whereas transport success peaks in November-December (late spring), at the beginning of the upwelling season. Our results show that, on average, during this period, particles have a 30% chance of reaching St Helena Bay with a standard deviation as high as 10%. The likelihood of particles making it northward decreases during the peak of the upwelling season (January-February), and reaches its lowest value (13%) in autumn. At this time of the year, the variability of the transport success also appears to be the lowest.

The age of arrival into the target area also varies seasonally (Fig. 4.3a). There is a peak in the age distribution around 7 days for particles released between July and October. For the rest of the year, the age distribution flattens and gradually shifts towards higher values during the upwelling season. From February to June, it is homogeneously spread between 10 and 25 days. The cumulative histogram of monthly transport success for 4 groups of particles built according to their age of arrival (group I: age<10 days; group II: 10<age≤15 days; group III: 15<age≤20 days; group IV 20<age≤30 days) reveals that particles arriving within 10 days account for more than 50% of the successful particles between August and November (Fig. 4.3b). Particles having short connection time are found all year round but are a minority in late summer and autumn. This suggests the existence of a "fast route" between the SARP line and St Helena Bay, which could be the dominant pathway in late winter and spring. Nevertheless, other routes could exist. There might even be a "slow route" from January to June, when the majority of particles belongs to group IV.

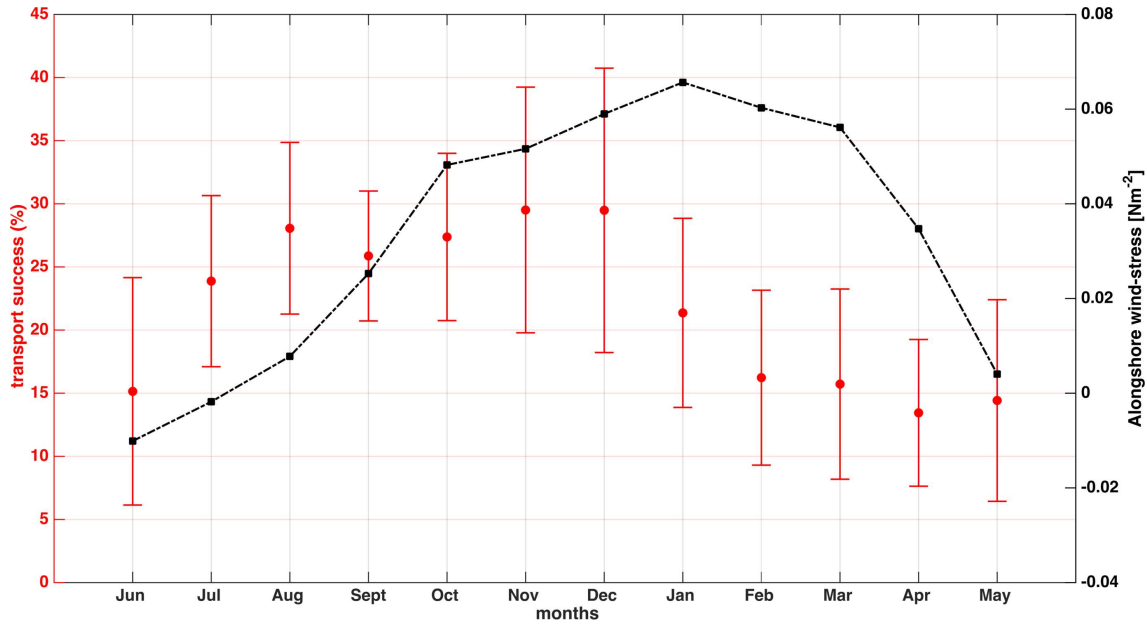


Figure 4.2: Monthly climatology of transport success TS_m (red dots) and alongshore wind-stress at the SARP line (black dots). The standard deviation of TS_m is also superimposed as errorbars. The wind stress is averaged over a box encompassing the Southern Benguela from 12°E to 20°E and 26°S to 36°S .

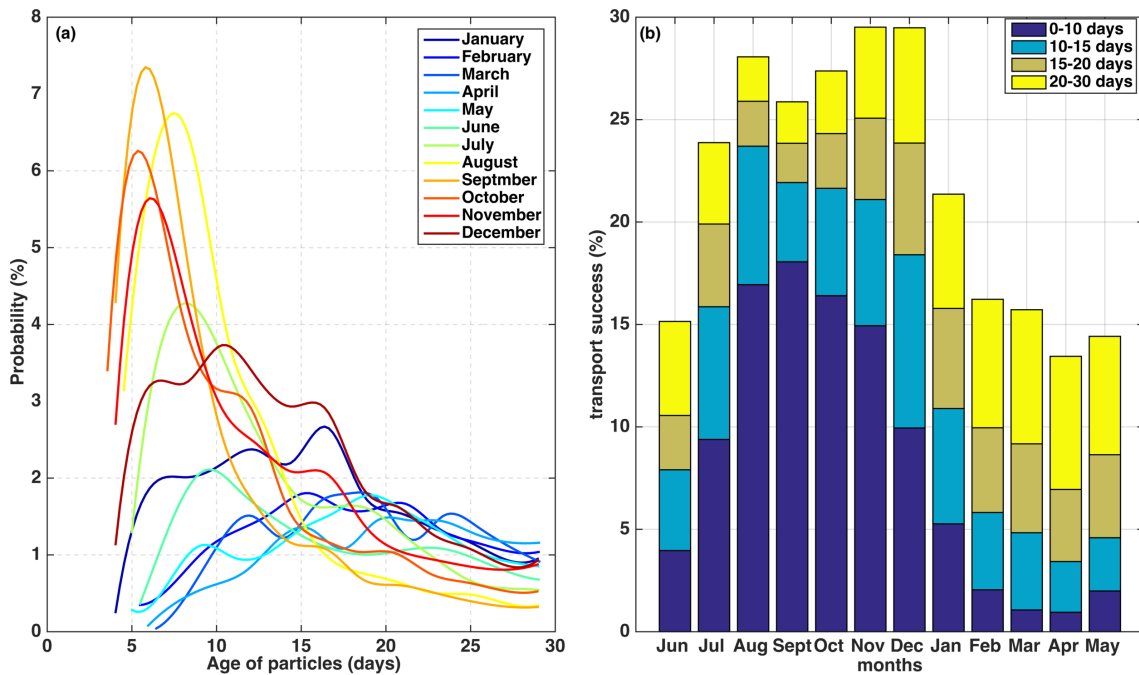


Figure 4.3: (a): Monthly distribution of successful particles by age when entering the target area. (b): Cumulated histogram of monthly climatology transport success with clusters of particles built according to their age of arrival: group I (age < 10 days); group II (10 < age ≤ 15 days); group III: (15 < age ≤ 20 days); group IV: (20 < age ≤ 30 days). Probability have been normalised by the total number of particles released for each month m ($m = 1, \dots, 12$) over the 22 years of the model run.

1060 The initial vertical and across-shore positions of particles are also important parame-
1061 ters affecting transport success. The majority ($\sim 90\%$) of particles arriving in St Helena
1062 Bay are released between 0 m and 70 m depth (Fig. 4.4a). There is a peak in the distri-
1063 butions around 55 m depth, whose intensity varies seasonally and intensifies in between
1064 October-March. Within this depth range, there is a pronounced seasonal contrast. Dur-
1065 ing austral winter, particles released in the top surface layer (0 to 35 m depth) have a
1066 higher probability of reaching St Helena Bay than during the rest of the year ($\sim 3.7\%$
1067 in July versus 1.7% in January). If only particles successfully transported are counted in
1068 the calculation, 50% of those arriving in St Helena Bay in July are those initially released
1069 within the top surface layer. The proportion drops to 25% for those released in January.
1070 During spring and summer, the opposite pattern is observed with particles released in
1071 the subsurface layers having a higher transport success. These results are consistent with
1072 the general assumption that, in eastern boundary upwelling systems (EBUS), upwelling
1073 favourable winds usually enhance the offshore advection of surface particles.

1074 Grouping particles according to the isobath on which they have been released is a way
1075 to investigate how the contrasted across-shore structure of the circulation at the SARP
1076 line impacts transport success. Particles released over the continental shelf (between the
1077 coast and the 300 m isobath) account for more than 50% of the successfully transported
1078 particles, although they only represent one sixth of the total number of particles released
1079 (Fig. 4.4b). Their probability of reaching St Helena Bay gets as high as 75% (not shown)
1080 in late winter and early spring. Further investigations (not shown) indicate that being
1081 released inshore not only increases the particles' probabilities to reach St Helena Bay, but
1082 also decreases their travel durations.

1083 These results suggest the existence of an efficient "fast" inshore route connecting the
1084 western edge of the Agulhas Bank to the west coast upwelling region. However, this
1085 analysis is only based on the final and initial positions of the particles. The identification
1086 of the pathways taken by these particles is done in the following section by considering
1087 the northward progression of particles during their 30 day life span and their average
1088 trajectories.

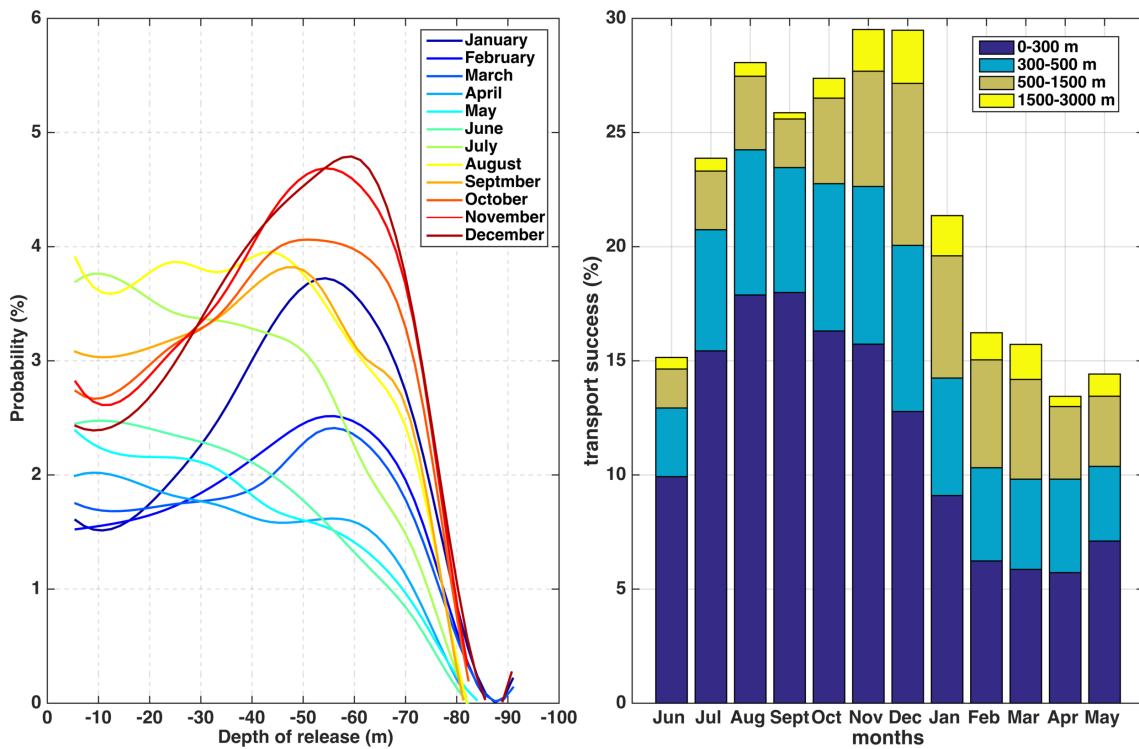


Figure 4.4: (a): Monthly distribution of successful particles by their depth of release. (b): Cumulative histogram of monthly climatology transport success with clusters of particles built according their initial isobath. Probability have been normalised as in Figure 4.3 .

4.4 Alongshore Lagrangian pathways

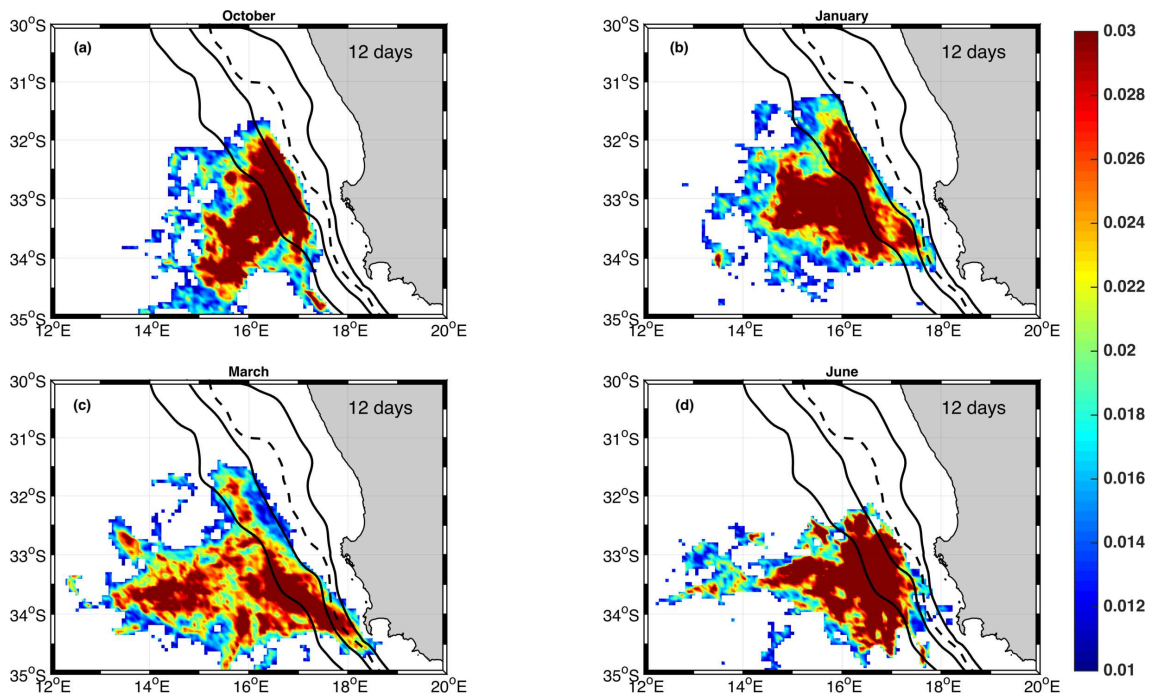


Figure 4.5: Horizontal distribution of unsuccessful particles (plumes) 12 days after their release for the months of: (a) October; (b) January; (c) March; (d) June. The 200 m, 300 m (dashed), 500 m and 2000 m isobaths are superimposed. The colorbar shows the probability of finding a particle aged 12 days within the water column at a specific grid cell.

1090 In order to get a general view of the routes taken by Lagrangian particles, maps of
 1091 particles' spatial distribution (plumes) are plotted according to their age τ (in days) on
 1092 the horizontal grid of the "BENGR15" simulation. The age of particles is zero when they
 1093 are released. For a series of release events, we count all particles aged τ found within the
 1094 water column of the grid cell located at position (x, y) in Cartesian coordinates. After
 1095 dividing this by the total number of particles released during this list of events, we obtain
 1096 the probability $d(x, y, \tau)$ of finding a particle aged τ at location (x, y) . We also build a
 1097 monthly climatology of this probability. For example, the monthly probability of January
 1098 is obtained by considering all release events that occurred in January over 1989-2010.

1099 Maps of the spatial distribution of the particles show that, once released on the SARP
 1100 line, successful and unsuccessful particles take distinct routes. 12 days after their re-
 1101 lease, unsuccessful particles have been largely spread offshore beyond the 500 m isobath
 1102 (Fig. 4.5), whereas successful particles remain trapped on the shelf edge between the 200 m

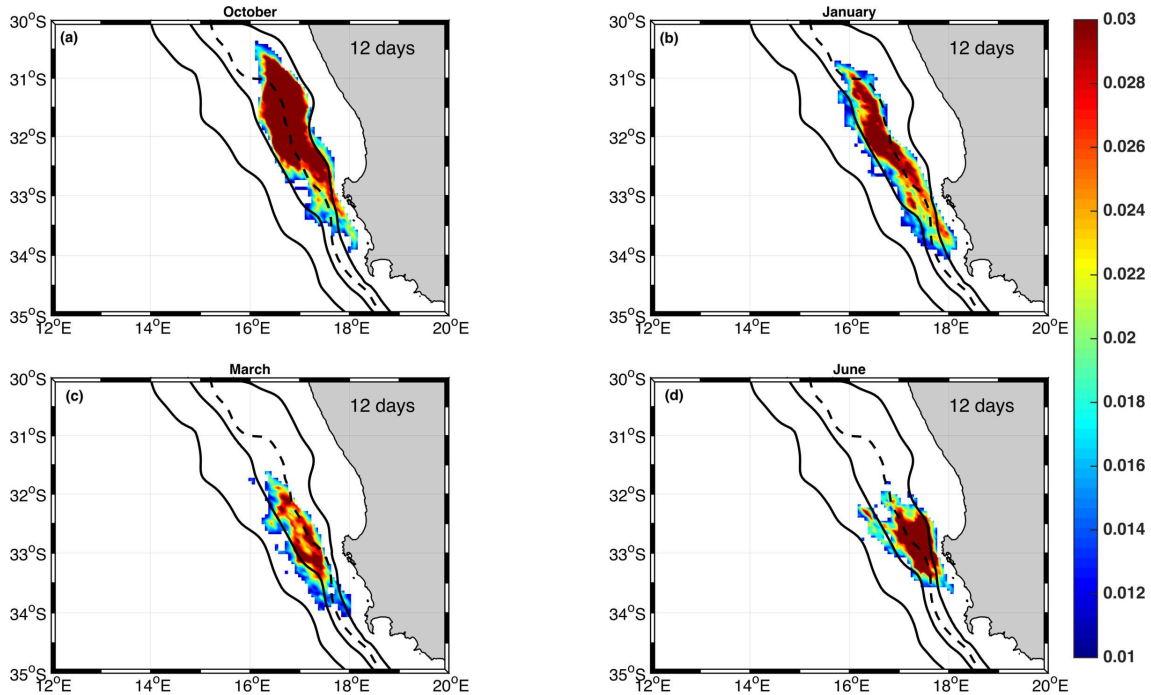


Figure 4.6: Same as Fig. 4.5 for the successful particles. The sum of probabilities over the grid gives the transport success plotted in Fig. 4.2.

1103 and 500 m isobath (Fig. 4.6). The characteristics of the plume change seasonally. In Oc-
 1104 tober, successful particles get dispersed in a compact and elongated plume, that stretches
 1105 quickly towards the equator following the shelf edge. Additionally, most of these particles
 1106 make their journey in less than 10 days after being released inshore of the 300 m isobath.
 1107 From January to June, the northward extension of the plume progressively reduces.
 1108 The fate of unsuccessful particles is to be dispersed offshore, following a northwestward
 1109 orientated cone that extends throughout the coastal transition zone (Fig. 4.5). They ex-
 1110 perience a widespread dispersion into the open ocean as they leave the shelf slope region
 1111 and are advected through the coastal transition zone, an area known for intense mesoscale
 1112 turbulence in EBUS (*Marchesiello et al., 2003; Capet et al., 2008a*). The north-westward
 1113 spreading of unsuccessful particles is maximum in March. The low values of transport
 1114 success found at this time of the year (Fig. 4.2) confirm the general idea that passive
 1115 particles, once advected offshore, struggle to come back inshore. From January to June,
 1116 the plume also shows a V-shape, past Cape Columbine, which suggests a branching of the
 1117 route. One branch seems to follow the slope whereas the other bifurcates offshore, as was
 1118 already pointed out in the work of *Garavelli et al. (2012)* and *Veitch et al. (2017)*.

1119 A monthly climatology of the successful particles' trajectories is presented in figure. 4.7.
1120 The latter is computed after joining together the geometric centre of the plume of suc-
1121 cessful particles every 12 hours from 0 to 30 days. This figure highlights the seasonal
1122 changes of the pathways taken by successful particles during their transit to St Helena
1123 Bay. In January and March, the trajectory of particles is shifted offshore. It starts bend-
1124 ing westward at the latitude of the SARP line, crosses the 500 m isobath, and finally veers
1125 back inshore towards the 500 m isobath at Cape Columbine. The majority of successful
1126 particles make their journey in more than 20 days (Fig. 5.15) and follow an offshore path.
1127 In October, the offshore route to St Helena Bay is no longer active. The majority of
1128 particles make their journey in less than 10 days following an inshore route that almost
1129 overlies the 300 m isobath. In June, the few particles reaching St Helena Bay also follow
1130 an inshore route, but the latter is slower than during October. Our findings on the sea-
1131 sonality of the pathways and their locations can be summarised as follows. The inshore
1132 route is the dominant pathway in late winter and early upwelling season, when offshore
1133 Ekman transport is moderate and particles are able to remain inshore. This inshore route
1134 is characterised by short connection times (< 10 days). During the peak of the upwelling
1135 season, this inshore route persists, but becomes less efficient, as it mainly concerns sub-
1136 surface particles located below the Ekman layer (about 40 m in the Benguela upwelling
1137 system). Connection times also tend to be longer (> 10 days). A slower offshore route,
1138 with connection times in the 20-30 days range, exists for particles either initially released
1139 on the shelf slope (from the 300 m to the 1500 m isobaths) or advected offshore by Ekman
1140 transport during their early age. This offshore route becomes the dominant pathway in
1141 summer.

1142 **4.5 Inferring the role of the ocean circulation on the La-** 1143 **grangian pathways**

1144 **4.5.1 Characteristics of the ocean surface circulation**

1145 The Lagrangian experiments show that a "fast" inshore route and a "slow" offshore route
1146 co-exist and bring particles from Cape Point to St Helena Bay. The following section

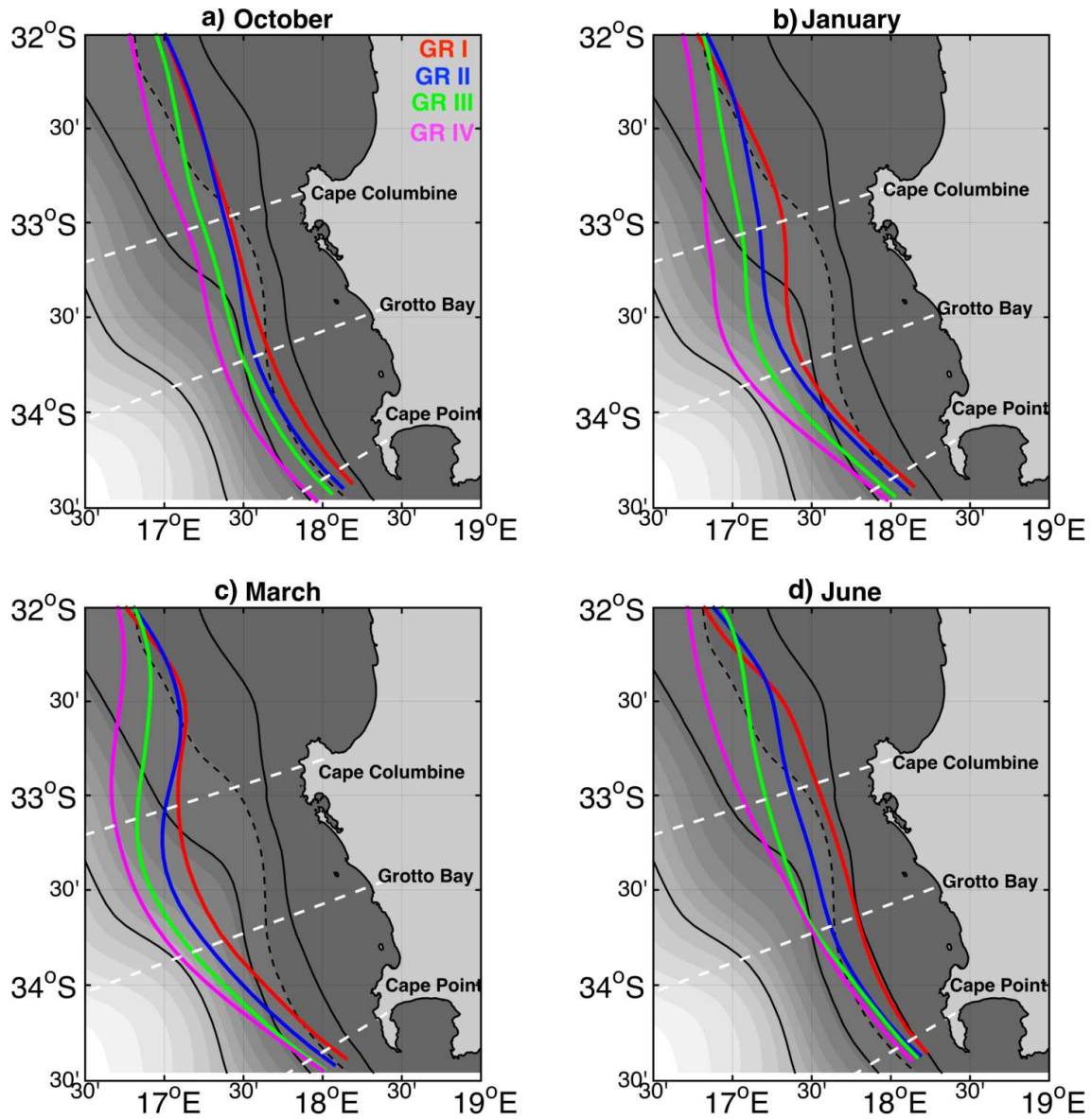


Figure 4.7: Monthly climatology of the trajectory of successful particles by age of arrival into the target area τ (in days). Trajectories are plotted for particles released in: (a) October; (b) January; (c) March; (d) June. colors refer to the clusters of particles built according to their age of arrival τ . Group I: $\tau < 10$ (red); Group II: $10 < \tau \leq 15$ (blue); Group III: $15 < \tau \leq 20$ (green); Group IV: $20 < \tau \leq 30$ (magenta). The 200 m, 300 m (dashed), 500 m and 2000 m isobaths are superimposed (black). Note that the mean of the four trajectories at 12 days corresponds to the geometric centre of the plume plotted in Fig. 4.6. The SARP, Grotto Bay and Cape Columbine lines described in Fig. 4.1 are also superimposed (white dashed).

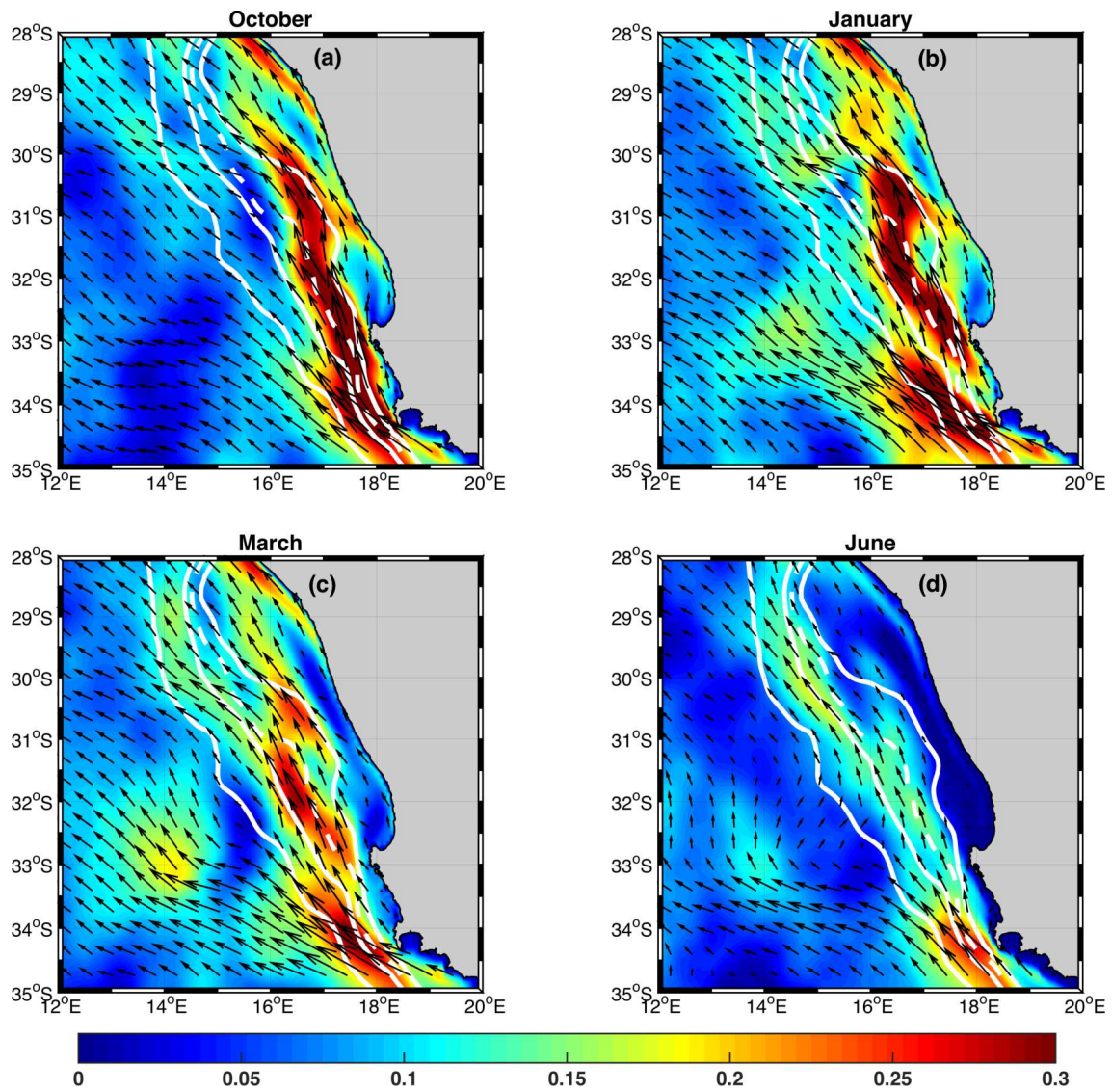


Figure 4.8: Monthly climatology of surface currents for: (a) October; (b) January; (c) March; (d) June. The magnitude of the current is shown in color (m s^{-1}) and current vectors are superimposed, as well as the 200 m, 300 m (dashed), 500 m and 2000 m isobaths (solid white).

1147 investigates how the Eulerian oceanic circulation of the model relates to these identified
1148 pathways.

1149 Maps of monthly climatological surface currents show a nearshore equatorward jet, namely
1150 the Benguela Jet (Fig. 4.8). This jet can be observed all-year round, with a marked sea-
1151 sonal variability. In October, it is intensified and spatially coherent. In January, at
1152 the vicinity of 33.5°S , it separates into two branches: i) an offshore branch that flows
1153 northwestward beyond the 2000 m isobath; ii) a coastal branch that flows alongshore in
1154 between the 200 m and 500 m isobaths and intensifies north of Cape Columbine. During
1155 the late upwelling season and early winter (March-June), the jet is still separated into two
1156 branches, but weakens, except off Cape Point (34°S).

1157 Considering that the Benguela Jet dynamically connects the south and the west coasts
1158 of South Africa (*Garavelli et al.*, 2012; *Veitch et al.*, 2017), we superimpose, along three
1159 across-shore transects, the across-shore profiles of the: i) distribution of the particles;
1160 ii) depth-integrated (top 100 m) alongshore velocity; iii) depth-integrated (top 100 m)
1161 eddy kinetic energy (EKE). The across-shore profiles are plotted for the SARP line along
1162 which particles are released (Fig. 4.9), for the transect off Grotto Bay that intersects parti-
1163 cles' trajectory at midway (Fig. 4.10) and for the transect off *Cape Columbine* (Fig. 4.11)
1164 that intersects particles' trajectory just before they enter the target area. The across-shore
1165 profile of the distribution of particles is obtained by counting particles crossing the tran-
1166 sects. Quantities are normalised by their maximum value found among the 12 monthly
1167 climatology across-shore profiles of all three transects (Fig. 4.7).

1168 Off Grotto Bay and Cape Columbine, the peak in the across-shore distribution of all par-
1169 ticles (blue solid) almost superimposes with the peak in the jet's alongshore intensity (red
1170 solid). As expected, all year round, particles leave the SARP line and propagate northward
1171 embedded within the Benguela Jet. A different pattern is observed when only successful
1172 particles are considered. During their northward progression, successful particles (blue
1173 dashed) always remain on the inshore (cyclonic) side of the jet, following roughly the
1174 300 m isobath. This area generally coincides with low values of surface EKE. The EKE
1175 can be interpreted as a proxy for the intensity of mesoscale oceanic turbulence. One would
1176 expect a particle that goes through an area of strong EKE to be more likely caught within

1177 mesoscale turbulent oceanic features, such as eddies and filaments. An exception to this
1178 pattern is the occurrence of a secondary peak of EKE nearby the peak of successful parti-
1179 cles in October off Grotto Bay (Fig. 4.10a). However, it does not have a negative effect on
1180 transport success because the alongshore jet and/or the topographic slope may be strong
1181 enough to counteract the turbulence dispersive effect in the near-shore region. In January
1182 and March, when the jet shifts offshore, presumably to conserve its potential vorticity as
1183 it gains cyclonic relative vorticity from the wind stress curl (*Castelao and Barth, 2007*),
1184 the peak in all particles also shifts offshore. This results in more particles entering a region
1185 where the surface EKE increases from low values at the coast to high values offshore.

1186

1187 At this stage of the analysis, the ability for particles to remain on the inshore side of
1188 the jet seems to be a necessary ingredient for transport success. Consequently, one would
1189 expect physical processes like wind-induced offshore/inshore Ekman transport during the
1190 upwelling/non-upwelling season to have a negative/positive effect on transport success.
1191 However, the picture is not as simple. This analysis brings to light that the broadening,
1192 the cross-isobath shift and the branching of the jet are other important processes that im-
1193 pact the alongshore connectivity between Cape Point and Cape Columbine, together with
1194 wind-induced Ekman transport. During the upwelling season, in January and March, de-
1195 spite intensified offshore Ekman transport, the broadening of the jet allows most particles
1196 to remain embedded within its core up to Grotto Bay (Fig. 4.10). Off Cape Columbine,
1197 multiple alongshore narrow jets exist all year round (Fig. 4.11). Ekman transport may ei-
1198 ther make particles follow one particular branch (January) or push them out of any branch
1199 (March). In March, the few successful particles are those trapped within the intensified
1200 inshore branch of the jet. In June, the distribution of the particles is disconnected from
1201 the jet (Fig. 4.10d). This can result from the inshore Ekman transport induced by pre-
1202 vailing westerly winter winds. It allows successful particles to remain trapped within the
1203 cyclonic inshore side of the jet up to Grotto Bay. Further north, when the jet divides into
1204 several branches, only the few most inshore particles manage to remain trapped within
1205 the inshore branch. The other ones get ejected from any branch, not even reaching the
1206 second less offshore branch. The across-shore position of particles at the location of the

1207 jet's division seems to be a crucial parameter that determines which branch particles may
1208 follow. Wind-induced Ekman velocities may set up this position.

1209

1210 Finally, our results underline the importance of intrusions of Agulhas waters. Off Cape
1211 Point, where particles are released, there is a strong variability in the across-shore structure
1212 of the alongshore jet (not shown). A significant proportion of particles (30%) is advected
1213 southward just after being released, and then advected back northward through the tran-
1214 sect. During their return northward path, they may cross the transect at a more offshore
1215 position than their initial position, which explains the mismatch (Fig. 4.9) between the
1216 initial distribution of all particles (green solid) and the distribution of all particles crossing
1217 the transect (blue solid).

1218 In summary, we have shown that the efficient and fast inshore route taken by the parti-
1219 cles in late winter and spring corresponds to particles being advected within an intense
1220 northward laminar nearshore jet. In summer and autumn, the route to St Helena Bay
1221 becomes less efficient and slower because of the offshore drift of the jet and its branching at
1222 Cape Columbine. Wind induced offshore Ekman transport contributes to particles being
1223 trapped within the offshore branches of the jet.

1224 **4.5.2 Characteristics of the ocean circulation at depth**

1225 Surface alongshore currents have a complex across-shore structure (multiple jets), and
1226 show pronounced latitudinal contrasts in the intensity and position of the jets. In the
1227 following, we investigate how the vertical structure of the flow, associated with the multiple
1228 jets observed at the surface, may explain the sensitivity of transport success to the initial
1229 depth of particles (Fig. 4.4).

1230 Vertical transects off Cape Point (not shown), Grotto Bay (Fig. 4.12) and Cape
1231 Columbine (not shown) all show the existence of a poleward undercurrent lying over the
1232 shelf slope. This poleward undercurrent seems to persist all year round, but has a strong
1233 seasonal cycle. In winter (June), it is weak ($\sim 5 \text{ cm s}^{-1}$), narrow ($\sim 40 \text{ km}$) and located
1234 deeper than 100 m depth. During the upwelling season, it is strong ($\sim 25 \text{ cm s}^{-1}$), wider
1235 ($\sim 100 \text{ km}$); its core extends in the offshore direction and towards the surface ($\geq 50 \text{ m}$

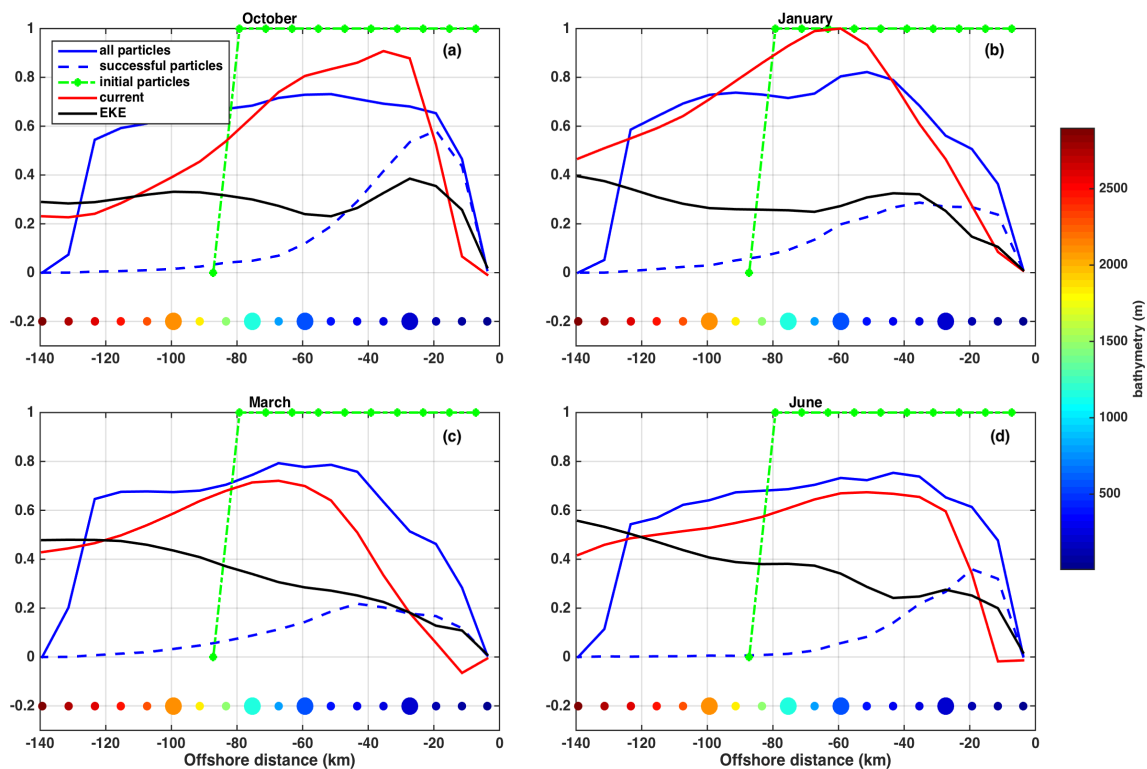


Figure 4.9: Across-shore profile of normalised depth-integrated alongshore current (red), EKE (black), and probability of the occurrence of particles (blue solid) at the SARP line. The histogram of successful particles is also superimposed (blue dashed) as well as the histogram of all particles at any release events (green solid). Profiles are plotted for: (a) October; (b) January; (c) March; (d) June. The depth-integrated alongshore current and EKE are computed over the top 100 m and then normalised by their maximum value found among the 12 monthly climatology across-shore profiles for all three transects. For the EKE, this maximum ($0.07 \text{ m}^2\text{s}^{-2}$) occurred along the Grotto Bay transect in May. For the depth integrated current, this maximum (0.45 m s^{-1}) was found at the SARP line in January. The probability of occurrence of particles is normalised by its value on the SARP line when they are released. The colorbar shows the across-shore bathymetry, the enlarged circles mark the location of the 200 m, 500 m, 1000 m and 2000 m isobaths, respectively.

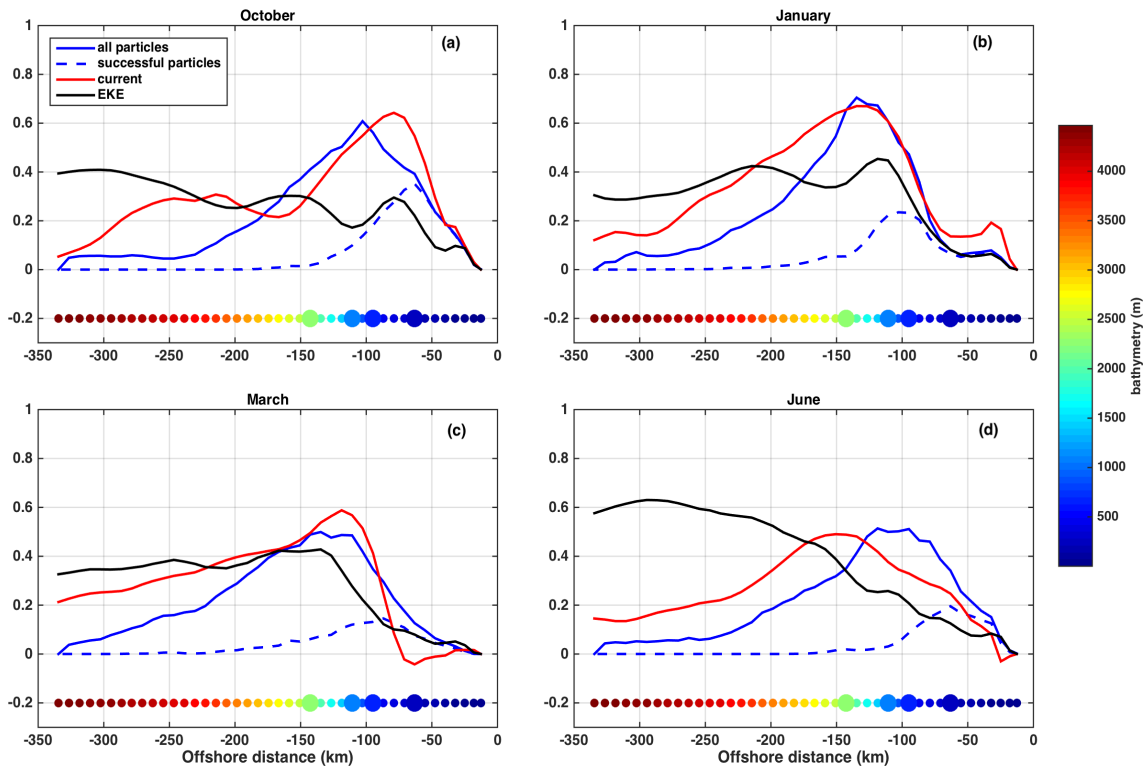


Figure 4.10: Same as Fig. 4.9 but for *Grotto Bay* (33.5°S).

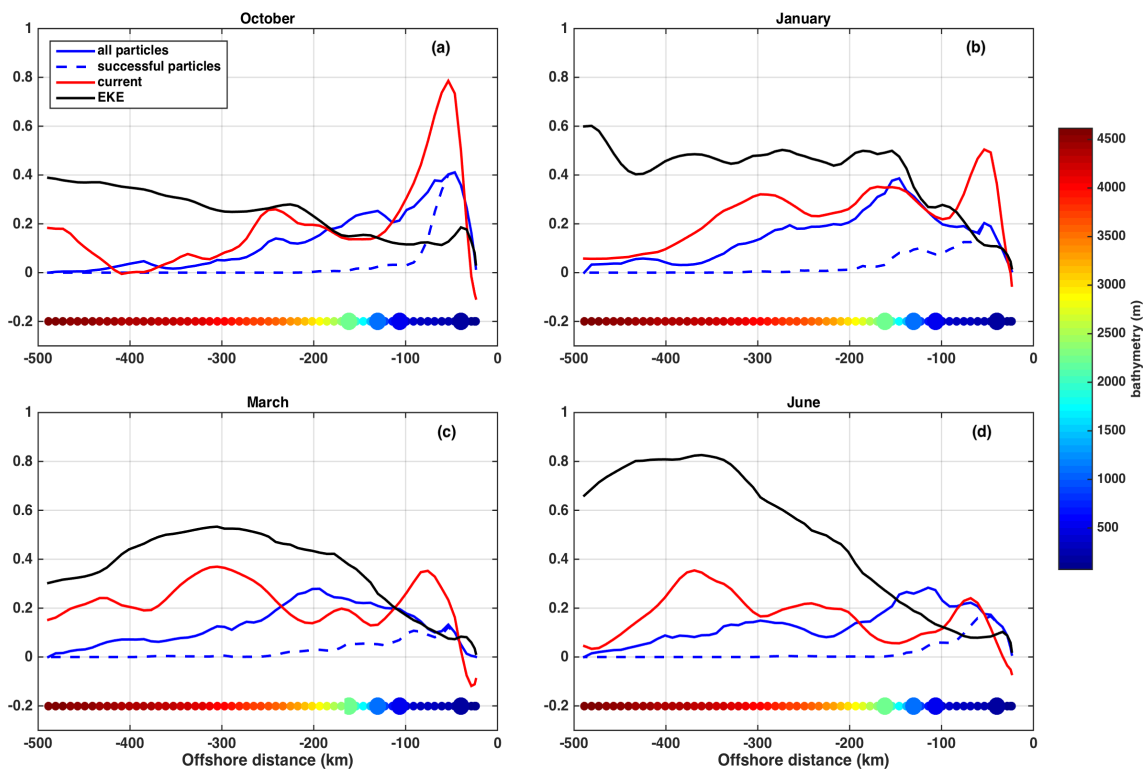


Figure 4.11: Same as Fig. 4.9 but for *Cape Columbine* (33°S).

1236 depth). The strengthening of the poleward undercurrent and its uplifting to the surface
1237 during austral summer is induced by the development of a negative (cyclonic) wind-stress
1238 curl within the 100 km nearshore band in relation to the coastal wind drop-off (Fig. 4.12).

1239 The shoaling and strengthening of the poleward undercurrent impacts the surface
1240 currents. At the SARP line, it mainly induces an offshore shift of the alongshore northward
1241 jet, which leads to the occurrence of a surface counter current in the nearshore area. North
1242 of Cape Point, the strengthening of the poleward undercurrent comes with the vertical
1243 pinching and weakening of the surface current. Off Grotto Bay, in January, the core of the
1244 undercurrent domes so sharply that it almost outcrops the surface 100 km from the coast.
1245 This mechanism is responsible for the bifurcation and separation of the surface current
1246 into multiple jets. While the main branch of the northward surface jet remains located
1247 offshore and extends as deep as 500 m depth, the coastal branch is squeezed between the
1248 surface and 50 m depth and follows the 300 m isobath. This squeezing of the surface jet
1249 should cause particles released below 50 m depth to be embedded within the poleward
1250 undercurrent. Consequently they would fail to reach St Helena Bay. The cumulated
1251 transport success according to the depth of release (not shown) shows the opposite. It
1252 turns out that the successful particles progress northward after getting trapped within the
1253 core of the thick offshore branch of the current.

1254 The processes impacting the particles located in the top 50 m of the water column
1255 can be explored by considering the ageostrophic velocities in the vertical across-shore
1256 transects off the west coast (Fig. 4.13). The latter are computed after subtracting the
1257 geostrophic velocity (derived from the hydrography) from the total velocity. As expected,
1258 ageostrophic currents are found in the surface and bottom boundary layers. When the
1259 upwelling favourable winds prevail in summer, intense offshore velocities are found in the
1260 top 40 m due to the Ekman drift. Surface particles get advected offshore into a region
1261 of intensified mesoscale activity. In EBUS, surface EKE generally increases as one goes
1262 offshore through the coastal transition zone but remains low at depths (*Capet et al.*,
1263 2008b). In the SBUS, intrusions of warm waters from the Agulhas Current in the form
1264 of eddies and filaments sustain high levels of EKE at depths (*Veitch et al.*, 2009; *Veitch*
1265 *and Penven*, 2017). This high EKE pattern is generated by eddies and filaments being

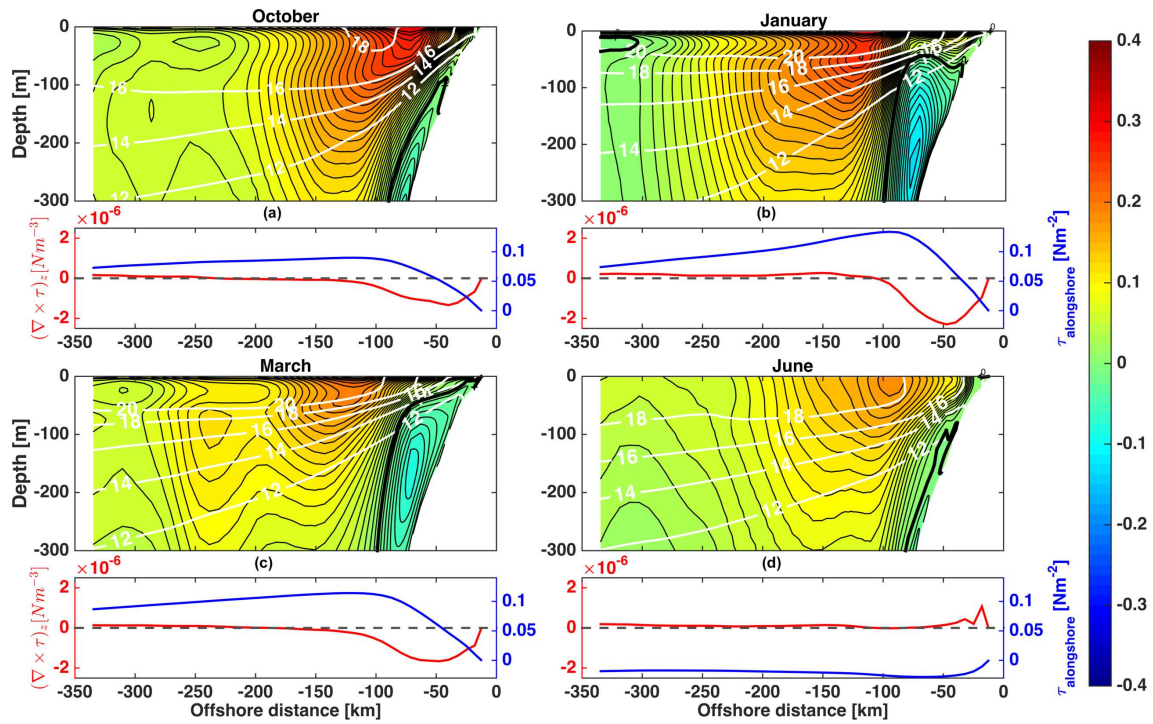


Figure 4.12: (Top): Grotto Bay vertical transect of alongshore velocities (color in m s^{-1}) and isotherms (white contours with contour spacing of 2°C). Contours of the current intensity are also superimposed (black) with a contour pacing of 1 cm s^{-1} and the zero contour in thick. (Bottom): The alongshore wind stress (blue) and z-component of the wind-stress curl (red) are superimposed for each across-shore transect, with the zero line (black dashed). Quantities are the monthly climatologies for: (a) October; (b) January; (c) March; (d) June.

1266 directly injected in the system from the Agulhas retroreflection region and/or being formed
1267 from baroclinic/barotropic instabilities of the jet itself. This is clearly observable off Grotto
1268 Bay in winter, when the absence of upwelling allows these intrusions to progress near the
1269 coast (Fig.4.14). In summer, these intrusions still happen but progress more offshore,
1270 beyond the upwelling density front. However, peaks of surface EKE can still be observed
1271 within the top 100 m of the water column in the first 200 km nearshore band. These peaks
1272 are linked to barotropic/baroclinic instabilities of the different branches of the jet. One
1273 peak coincides with the offshore anticyclonic part of the surface jet, whereas the inshore
1274 peak coincides with the poleward undercurrent.

1275 4.6 Discussion and conclusions

1276 Lagrangian experiments, based on a 22 years (1989-2010) interannual simulation of the
1277 oceanic circulation in the SBUS, are used to describe the alongshore connectivity between
1278 Cape Point and St Helena Bay. *Hutchings* (1992) had previously pointed out the existence
1279 of an alongshore corridor followed by fish eggs and larvae that were spawned on the Agul-
1280 has Bank. A series of Lagrangian studies had then linked this corridor with the existence
1281 of the Benguela Jet, a shelf edge equatorward current (*Blanke et al., 2009; Veitch et al.,*
1282 2017).

1283 This study confirms that successful particles leave the SARP line, *i.e.* the transect off
1284 Cape Point, and progress northward embedded within the Benguela Jet. Nevertheless,
1285 the latter intensifies during the upwelling season, while transport success decreases dur-
1286 ing this period. This brings to light that other dynamical processes impact Lagrangian
1287 transports. Wind-induced across-shore Ekman transport, offshore mesoscale instabilities,
1288 Agulhas intrusions in the form of eddies and filaments combine together with the com-
1289 plex structure of the alongshore jets to transport particles towards St Helena Bay. The
1290 across-shore position of the jet, its intensity, its orientation relative to the coast and its
1291 bifurcation also play a key role and need to be taken into account when quantifying the
1292 alongshore connectivity.

1293 We have shown that there is an efficient fast inshore route during late winter and early
1294 spring that is facilitated when the surface equatorward alongshore jet is in the form of one

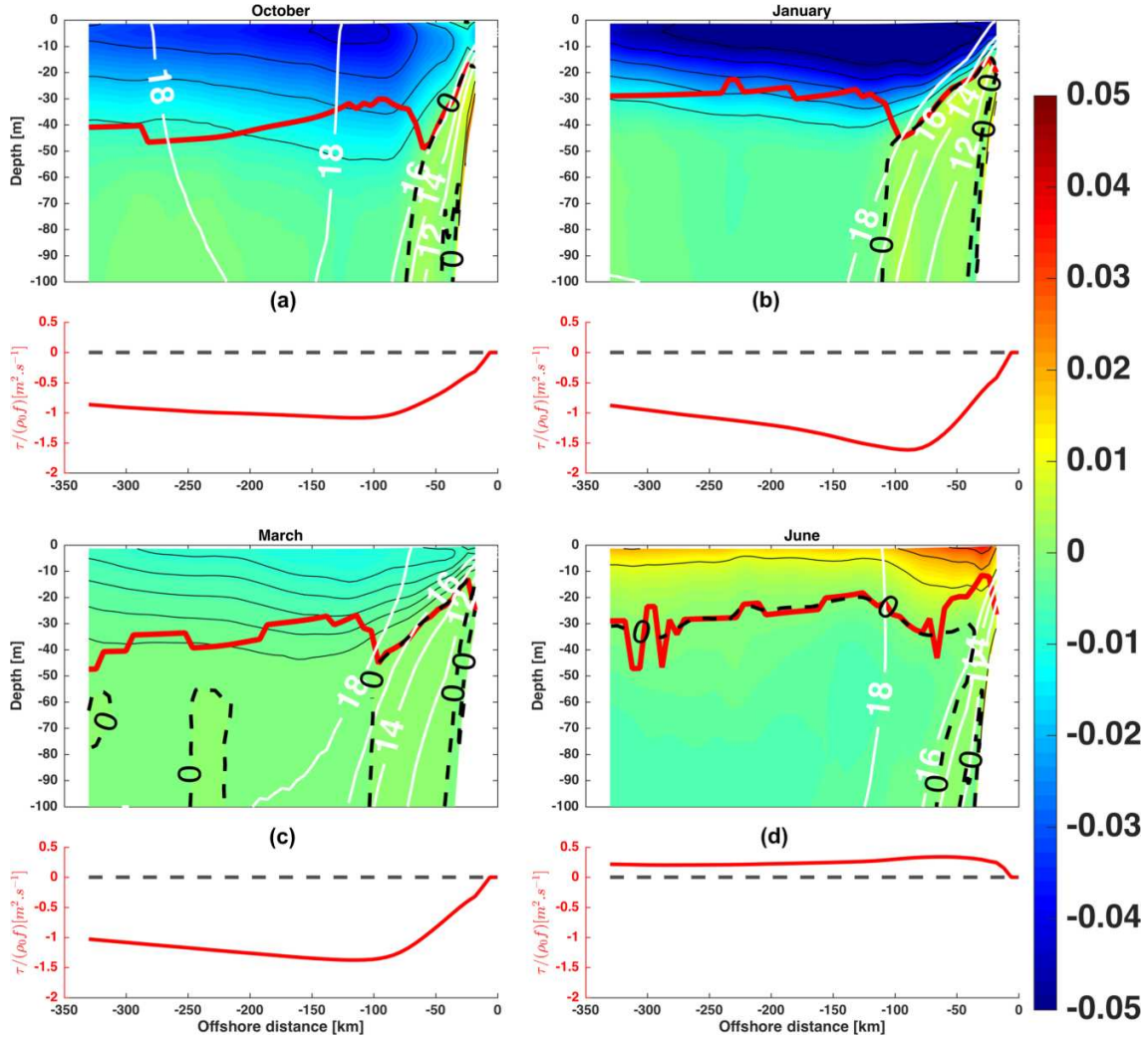


Figure 4.13: (Top): Grotto Bay vertical transect of across-shore ageostrophic velocities (color in m s^{-1}) and isotherms (white contours with contour spacing of 2°C). Contours of the current intensity are superimposed (black) with a contour spacing of 1 cm s^{-1} (zero contour in thick). The depth of the Ekman layer (m), computed after matching the vertical integration of the ageostrophic velocity with the total across-shore Ekman transport ($\tau_{\text{alongshore}}/(\rho f)$) is also superimposed in red. (Bottom): The across-shore Ekman transport along each across-shore transect (red), with the zero line (black dashed). Quantities are the monthly climatologies for: (a) October; (b) January; (c) March; (d) June.

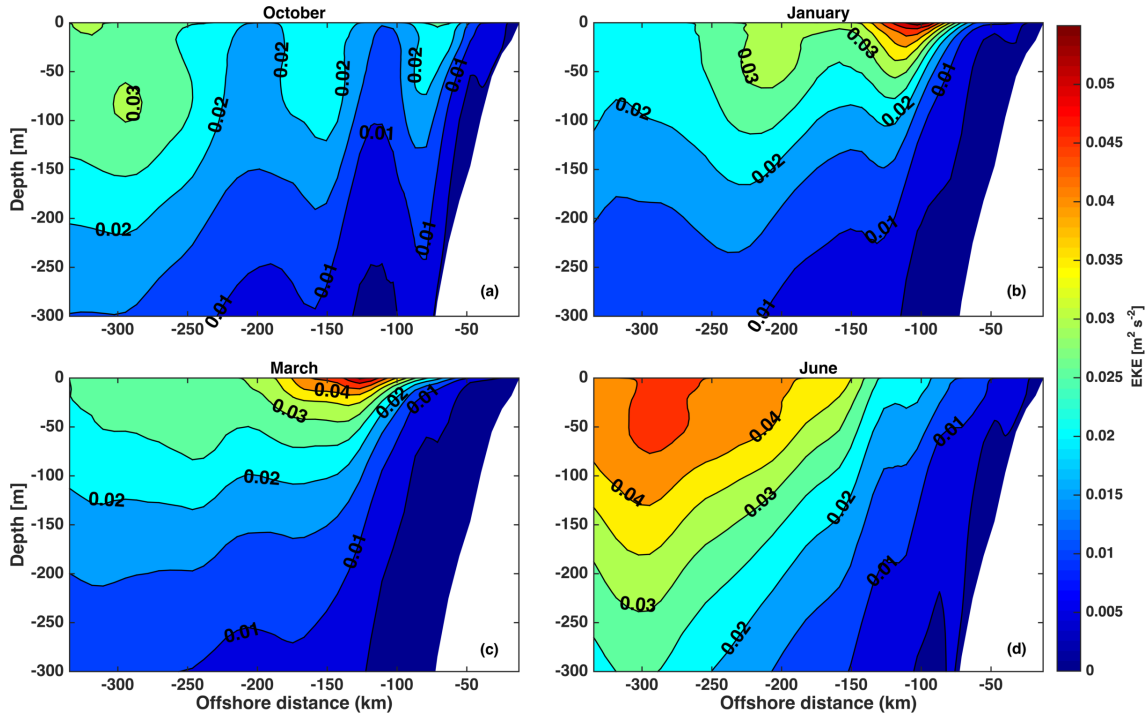


Figure 4.14: Grotto Bay vertical transect of monthly climatology of EKE (color in $\text{m}^2 \text{s}^{-2}$) for: (a) October; (b) January; (c) March; (d) June.

1295 coherent intense nearshore branch. At this time of the year, transport success between
 1296 Cape Point and St Helena Bay is at its highest rate (30%), and mainly involves particles
 1297 initially released within the nearshore band delimited by the 300 m isobath: 90% of these
 1298 latter particles are successfully transported to St Helena Bay in less than 15 days following
 1299 a laminar route with little across-shore dispersion.

1300

1301 When the prevailing upwelling favourable winds start blowing consistently, transport
 1302 success decreases continuously until it reaches its lowest value (15%) at the end of the
 1303 upwelling season (March). Previous studies have attributed this decrease to an offshore
 1304 loss of particles induced by a peak in the offshore wind induced Ekman drift. Our study
 1305 shows that a reduced transport success can also be linked to a change in the structure
 1306 of the equatorward jet. This occurs when the jet is shifted offshore towards a region of
 1307 higher EKE, and particles are then more likely to fall within mesoscale features such as
 1308 eddies or filaments and to get dispersed offshore. At this time of the year, the fast inshore
 1309 successful route still exists, but becomes less likely than other offshore routes. Particles

1310 taking these offshore routes have a longer journey towards the target area (> 20 days).

1311

1312 This study confirms that the offshore deviation of the Benguela Jet is linked to the
1313 poleward undercurrent intensification and shoaling (*Veitch et al.*, 2017). During the up-
1314 welling season, the wind drop-off within the first 100 km coastal band (due to the contrast
1315 in the land/sea drag coefficient) induces injection of negative cyclonic wind stress curl
1316 within the water column. The core of the poleward undercurrent thickens and outcrops to
1317 the surface, inducing the branching of the surface jet and/or a surface counter current in
1318 the nearshore band. This branching directly impacts the transport success, as successful
1319 particles follow the inshore branch. To do so, they must already be located within the
1320 inshore cyclonic side of the jet's core at the latitude of the jet's bifurcation. We have also
1321 shown that wind-induced Ekman surface drift acts on the particles either to keep them
1322 within the inshore side of the current in winter or to push them further offshore in summer.
1323 In summary, our simulations identified an efficient inshore pathway taken by particles in
1324 spring. This period is the optimal window, as the jet is in its intensified state, close to
1325 the shore, and relatively stable. At this time of the year, offshore Ekman transport is still
1326 low enough to allow particles to remain on the inshore side of the jet. The core of the
1327 poleward undercurrent remains located deeper than 100 m depth and therefore does not
1328 trap any particles.

1329 One may question the limitations of this work in terms of realism of the ocean circulation
1330 model, and argue about the robustness of our results if carried out with a higher resolu-
1331 tion model. Increasing the model resolution may certainly lead to a better representation
1332 of the bathymetry, in particular in regions with steep slopes, which may in turn modify
1333 the jet's positions and intensity, and therefore impacts the absolute values of transport
1334 success. However, we believe that the seasonal cycle of transport success and the processes
1335 that lead to it would still hold.

1336 The characteristics of the jet in the 7.5 km interannual simulation, with the bifurcation
1337 and branching into multiple cores, are consistent with observations. The separation of
1338 the jet off Cape Columbine was pointed out in *Fowler and Boyd* (1998). *Shannon and*
1339 *Nelson* (1996) (their fig. 20) found a double-core jet off Cape Point. The two cores were

1340 respectively located over the 100 m and 300 m isobath and separated by the shoaling of
1341 poleward undercurrent. Additionally, the seasonal cycle of transport success is intimately
1342 related to the seasonal cycle of the Benguela Jet itself. The latter is in thermal wind-
1343 balance with the across-shore temperature gradient. Coarser simulations, forced with a
1344 monthly seasonal climatology wind, already had the ability to reproduce the seasonal cycle
1345 of this gradient just because they could reproduce the seasonality of the coastal upwelling
1346 (*Penven et al., 2001; Veitch et al., 2017*). Lagrangian studies based on these simulations
1347 found a seasonal cycle of transport success similar to ours (*Parada et al., 2003; Huggett*
1348 *et al., 2003; Mullon et al., 2005*). The main difference concerned the winter particles that
1349 had a higher chance of reaching St Helena Bay than in our simulations (*Huggett et al.,*
1350 *2003*), probably because inshore Ekman transport kept particles within the inshore jet,
1351 and the latter had less seasonal variability.

1352

1353 Lagrangian experiments based on a higher resolution simulation are expected to pro-
1354 vide new insights on the role of (sub-)mesoscale turbulence on the dispersion of particles.
1355 Future studies should analyse this variability in detail using a higher resolution (≤ 3 km)
1356 nested simulation embedded within the 7.5 km simulation described here. Combining a
1357 better resolution of the innershelf dynamics together with a high-frequency atmospheric
1358 forcing should also lead to a more subtle nearshore dynamics made of thin jets (*Gan and*
1359 *Allen, 2002b,a*) and inertial waves (*Simpson et al., 20021; Lucas et al., 2014*).

1360 This study has helped us identify some of the physical processes important for the dis-
1361 persal of fish larvae in the SBUS. Our results show that being entrained within the fast
1362 inshore route taken by the Benguela Jet and avoiding being carried offshore when the jet
1363 splits into several branches are two major drivers of the connectivity between the spawning
1364 and nursery grounds. They must be considered as two important components of the fish
1365 reproductive strategy in the Southern Benguela. Finally, this study shows that successful
1366 transport from the Agulhas Bank to the west coast upwelling region cannot be attributed
1367 to only a simple wind induced modulation of the jet, as mentioned in previous studies.
1368 The seasonal characteristics of transport success result from the complex interactions of
1369 the Benguela Jet, the poleward undercurrent and the offshore Ekman transport together

1370 with the turbulent off-shelf eddy field.

1371 To conclude, it is important to recall that if the Lagrangian particles used for this study
1372 were to be considered as living larvae, transport success to the nursery grounds by the
1373 oceanic currents is only one of the factors that contribute to their survival. Other physical
1374 properties such as water temperature, vertical stratification, small-scale turbulence also
1375 strongly influence the growth rate of larvae, as well as their survival by affecting the
1376 availability, distribution and composition of the prey and predator field.

1377 Chapter 5

1378 The impact of downscaling on 1379 SBUS circulation and Lagrangian 1380 pathways

1381 5.1 Introduction

1382 Chapter 4 underlined the importance of the Benguela Jet for the northward along-shore
1383 transport of particles from Cape Point to St Helena Bay. It also suggested that the
1384 presence of mesoscale eddies, mostly located offshore, beyond the 500 m isobath, could
1385 have a negative effect on this northward transport success. However, these results relied
1386 on a mesoscale eddy-permitting simulation with a horizontal resolution of 7 km. In the
1387 Southern Benguela Upwelling System (SBUS), the first baroclinic Rossby deformation
1388 radius is of the order of 35 km offshore and becomes as low as a few kilometers on the
1389 shelf. Therefore, many of the fine-scale dynamics that constitute the inertia range of the
1390 energy spectra (*Capet et al.*, 2008b) were not properly resolved in the reference simulation.
1391 The 7 km horizontal resolution was not sufficient to accurately represent nor the mesoscale
1392 eddies generated through the barotropic/baroclinic instabilities of the Benguela Jet (*Veitch
1393 et al.*, 2009), nor the the complex structure of the innershelf circulation in depths less than
1394 100 m.

1395 In Eastern Boundary Upwelling Systems during the upwelling season, surface currents

1396 are predominantly directed equatorward in thermal wind balance with the cross-shore
1397 density front resulting from the inshore upwelling of cold waters. Nevertheless, current
1398 reversals occur when the prevailing equatorward upwelling favourable winds relax. Usually,
1399 episodes of upwelling favourable winds and relaxation occur with a 2 to 4 day period. These
1400 events have been extensively documented for the California mid-shelf region characterised
1401 with depths beyond 100 m (*Send et al.*, 1987; *Winant et al.*, 1987; *Huyer and Kosro*,
1402 1987; *Kosro*, 1987; *Largier et al.*, 1993; *Dever et al.*, 1998). They lead to strong poleward
1403 flows of tens of cm s^{-1} (*Cudaback et al.*, 2005; *Roughan et al.*, 2006). Inner-shelf current
1404 reversals have also been reported along the Iberian coast (*Relvas*, 2002) and in the SBUS
1405 (*Fawcett et al.*, 2008) where they can lead to increases of temperature up to $5^{\circ}\text{C day}^{-1}$
1406 in the very near-shore region in depths less than 50 m depth. The forcing mechanism
1407 for these nearshore poleward flows can either be the presence of a strong cyclonic wind
1408 stress curl due to the coastal wind drop-off or along shore pressure gradients that form in
1409 the presence of headlands, inside bays, during episodes of prevailing upwelling favourable
1410 winds (*Gan and Allen*, 2002a). When the equatorward along-shore wind stress relaxes,
1411 these pressure gradients are no longer balanced, which results in a poleward jet that flows
1412 around the headlands and transport some warm waters (*Gan and Allen*, 2002b; *Melton*
1413 *et al.*, 2009; *Washburn et al.*, 2011).

1414 In this chapter, our goal is to study the differences in the coastal circulation and
1415 mesoscale turbulence that emerge when the model resolution is increased, and to high-
1416 light how a finer representation of the dynamics may affect the alongshore connectivity
1417 between Cape Point (34°S) and St Helena Bay (32°S). Our analysis focuses more specif-
1418 ically on the particles initialised in the nearshore coastal band (depth less than 200 m),
1419 where gains in the representation of the fine-scale coastal circulation are expected to be
1420 most significant when increasing the model resolution. We follow the same approach as in
1421 chapter 4. Lagrangian particles are released every 12 hours along a cross-shore transect
1422 off Cape Point and then followed for 30 days. However, the currents advecting the parti-
1423 cles are now the outputs a 3-km horizontal resolution simulation nested into the reference
1424 configuration (7-km horizontal resolution: "BENGR15"). Like in chapter 4, we first focus
1425 on the physical processes that control the seasonal cycle of the along-shore northward

transport of particles. The analysis of the interannual variability is left to chapter 6. Sec-

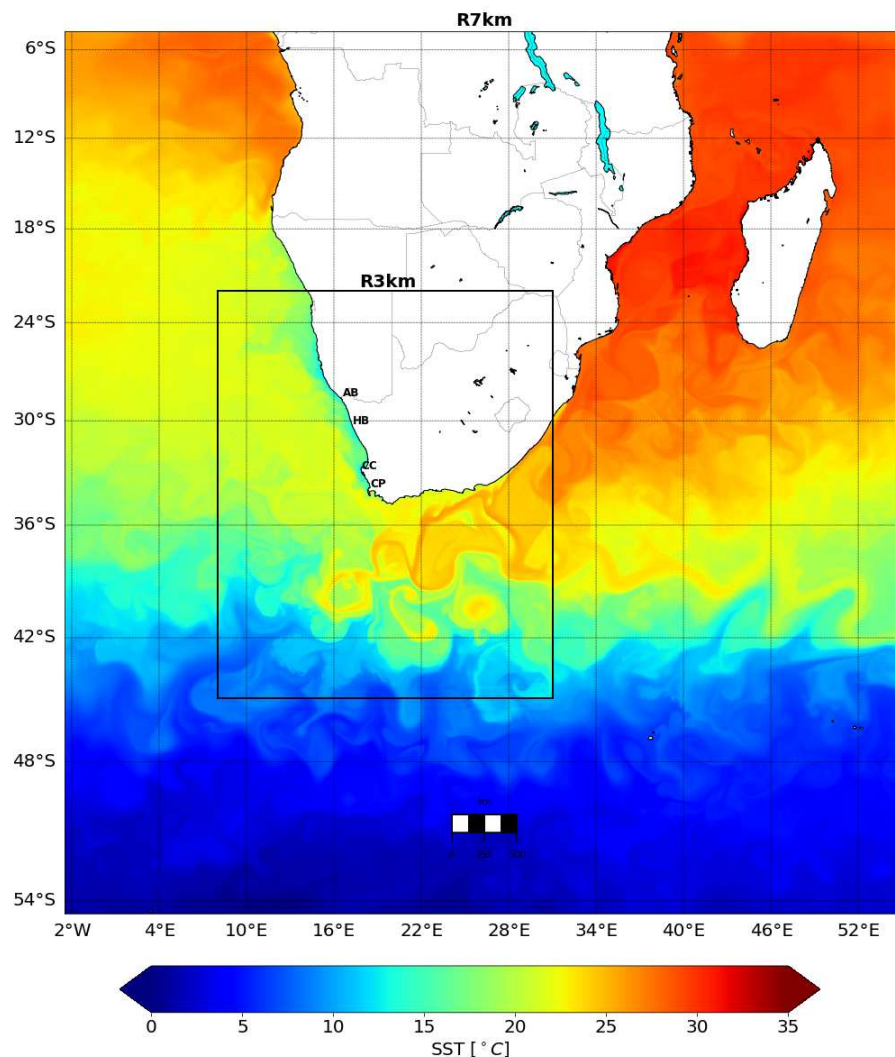


Figure 5.1: Snapshot of SST (Year 2005, 1st of January). The black box delineates the boundaries of the nested 3-km child grid. Within the box, the SST of the 3-km simulation is superimposed upon the one of the 7-km simulation.

1426

1427 tion 5.2 describes the new 3-km horizontal resolution nested simulation. Increasing the
1428 horizontal resolution improves the realism of the ocean circulation in the SBUS. The im-
1429 pact of model resolution on the mean circulation features and its variability are presented
1430 in section 5.3. The impact of model resolution on the Lagrangian pathways are then
1431 presented in section 5.5. Lagrangian experiments of particles advected with currents of
1432 the two simulations are carried out to investigate the impact of downscaling on transport
1433 success.

1434 5.2 A 3-km resolution simulation of the SBUS

1435 5.2.1 The design of an offline nested grid:

1436 A new grid with a horizontal resolutions of 3-km and 75 vertical levels (850 × 691 × 75
1437 grid points) is designed to encompass most of the coastal shelf waters of South Africa and
1438 to include the region of the Agulhas Current retroflexion. This new domain is embedded
1439 within our reference 7-km grid (Fig. 5.1). It extends from 45°S to 22.5°S and from 8°E
1440 to 30°E. Due to computational cost, the number of vertical levels of the 3-km resolution
1441 simulation is reduced to 75 sigma levels, instead of 100 in the 7-km reference simulation
1442 (Fig. 5.3).

1443 Bathymetry

1444 The bathymetry of our new configuration is built following the same methodology as for
1445 our reference simulation (Chapter. 3). The 1-min global dataset provided by the Shuttle
1446 Radar Topography Mission (*Smith and Sandwell, 1997*) is interpolated on the new 3-
1447 km grid, and smoothed out so that the steepness of the topography ($r = \frac{\delta h}{2h}$) does not
1448 exceed 0.2 (*Beckmann and Haidvogel, 1993*). Because the smoothing is dependent on
1449 the model grid resolution, the 3-km nested grid allows for some fine scale features to
1450 be better represented (Fig. 5.2), both along the shelf edge and offshore. In the 7-km
1451 configuration, the Cape Canyon which extends south-west of Cape Columbine was almost
1452 completely smoothed out. In the 3-km configuration, it has a clear signature. Cross-
1453 shore transects off Cape Point also show some serious impact on the shelf slope. Two
1454 shelf breaks can be distinguished in the 3-km simulation at 30 km and 80 km offshore
1455 respectively, whereas in the 7-km configuration, the continental slope plunges smoothly
1456 towards depth (Fig. 5.3). Because along-slope jets are usually constrained by the iso-
1457 contours of constant planetary potential vorticity (f/h contours), these modifications of
1458 the across-shore shelf slope are expected to impact strongly the positions of the along-slope
1459 jets. In addition, isolated seamounts between the Agulhas bank and Agulhas ridge, the
1460 west-southwest/east-northeast orientated ridge that bounds the Cape Basin at southern
1461 latitudes, do not get smoothed out within the 3-km resolution grid. The latter are expected
1462 to play a large role in the leakage of Agulhas waters into the Cape Basin because of they

1463 are on the main trajectory of the Agulhas rings (*Arhan et al., 1999; Schouten et al., 2000*).

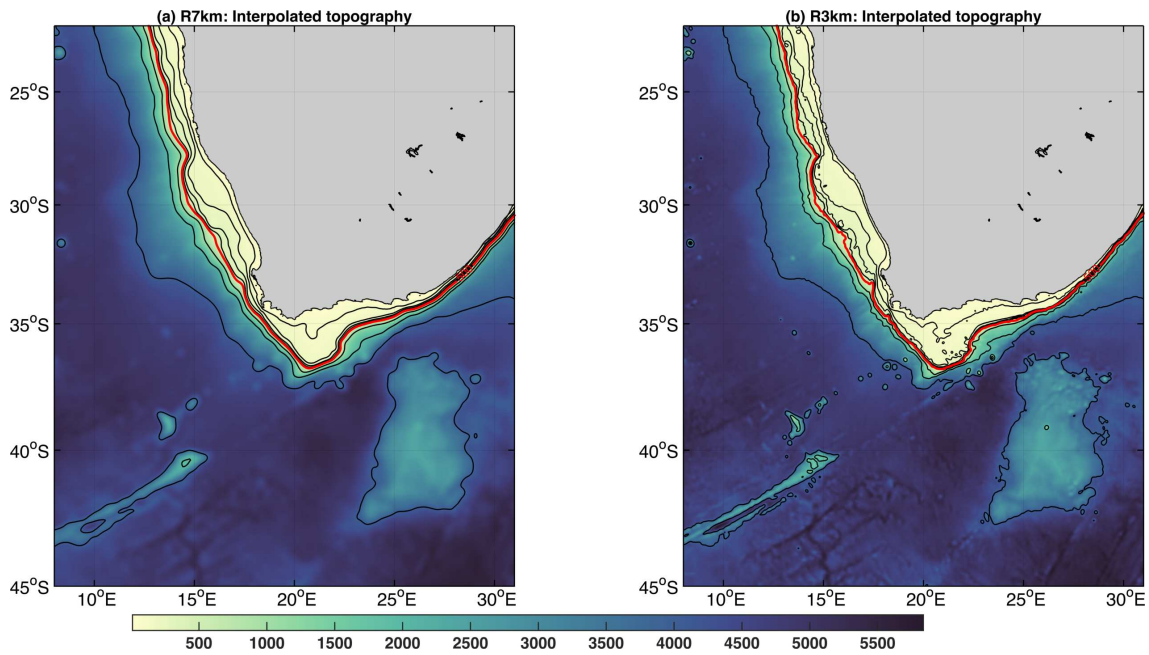


Figure 5.2: Model bathymetry for the reference configuration (R7km) and the nested domain (R3km)

1464

1465 **One way nesting at the lateral boundaries:**

1466 A one-way nested approach is employed using the roms2roms algorithm (*Mason et al.,*
1467 2010). While the lower resolution parent grid supplies boundary conditions to the higher
1468 resolution child domain, there is no feedback from the child to the parent. In the following,
1469 we use a daily-average archive of all prognostic variables (u, v, η, T, S) from the R7km
1470 reference simulation (Chap. 3) to specify the momentum, heat and fresh-water fluxes at
1471 the boundaries of the 3-km resolution simulation.

1472 **Surface forcing**

1473 For consistency, both simulations were identically forced at the surface with the same
1474 6-hourly atmospheric forcing (heat, fresh and momentum fluxes) provided by CFSR at
1475 0.3° horizontal resolution (*Saha, 2010*).

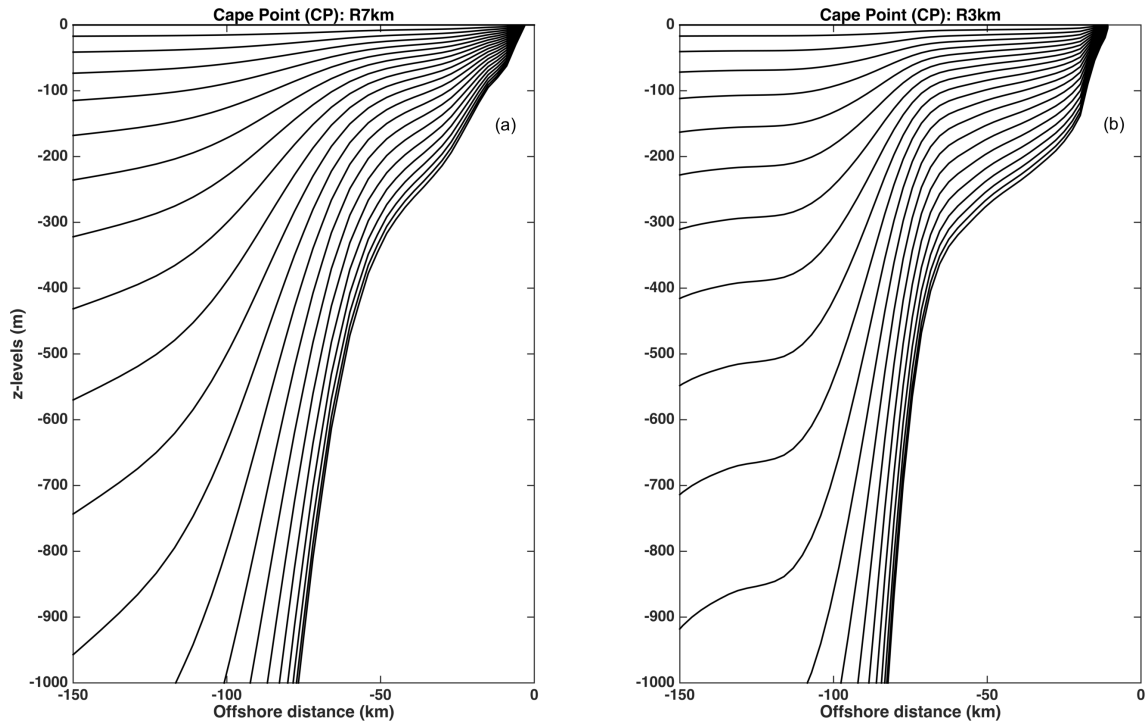


Figure 5.3: Cross-shore transects of vertical levels off Cape Point in the (a) R7km and (b) R3km model, plotted for every fifth level.

1476 **Initial condition and time integration**

1477 To have a coherent set of simulations, the R7km was run again and saved at a daily rate
 1478 instead of 3 days. The 3-km resolution simulation is initialised on 1st January 1990 after
 1479 interpolating the daily average archive of the 7-km simulation onto the new grid. The
 1480 new simulation is run from 1989 to 2011, and daily averages of all variables of interest
 1481 are archived. Our analysis focuses on the 1992-2011 period after discarding the first three
 1482 years to avoid spurious dynamics related to adjustment processes. A three years spin-up
 1483 phase was considered to be long enough to reach a statistical equilibrium. Some of the
 1484 key parameters of the two simulations are summarised in table 5.1. Hereafter, these two
 1485 simulations are referred to as R7km for the reference 7-km resolution simulation, and
 1486 R3km for the new 3-km resolution simulation.

1487 **5.3 Impacts of the horizontal resolution on the circulation**

1488 Before testing the impact of the horizontal resolution on the Lagrangian pathways of
 1489 particles in the SBUS, it is necessary to assess the differences in the representation of

Simulations	R7km	R3km
Domain	54°S-6°S; 2°W-54°E	45°S-22.5°S; 8°E-30°E
Grid points	909x866	850x691
σ -levels	100	75
Output frequency	Daily	Daily
Run period	1989-2011	1992-2011
Surface forcing	CFSR 6h 0.3°	CFSR 6h 0.3°
Lateral forcing	SODA 2.3.2	R7km
Bathymetry	SRTM 30 plus	SRTM 30 plus

Table 5.1: Summary of model configuration for the R7km and R3km simulations.

1490 circulation between the R7km and R3km simulations. This section therefore compares the SST, SSH, surface currents and EKE between the two simulations and observations.

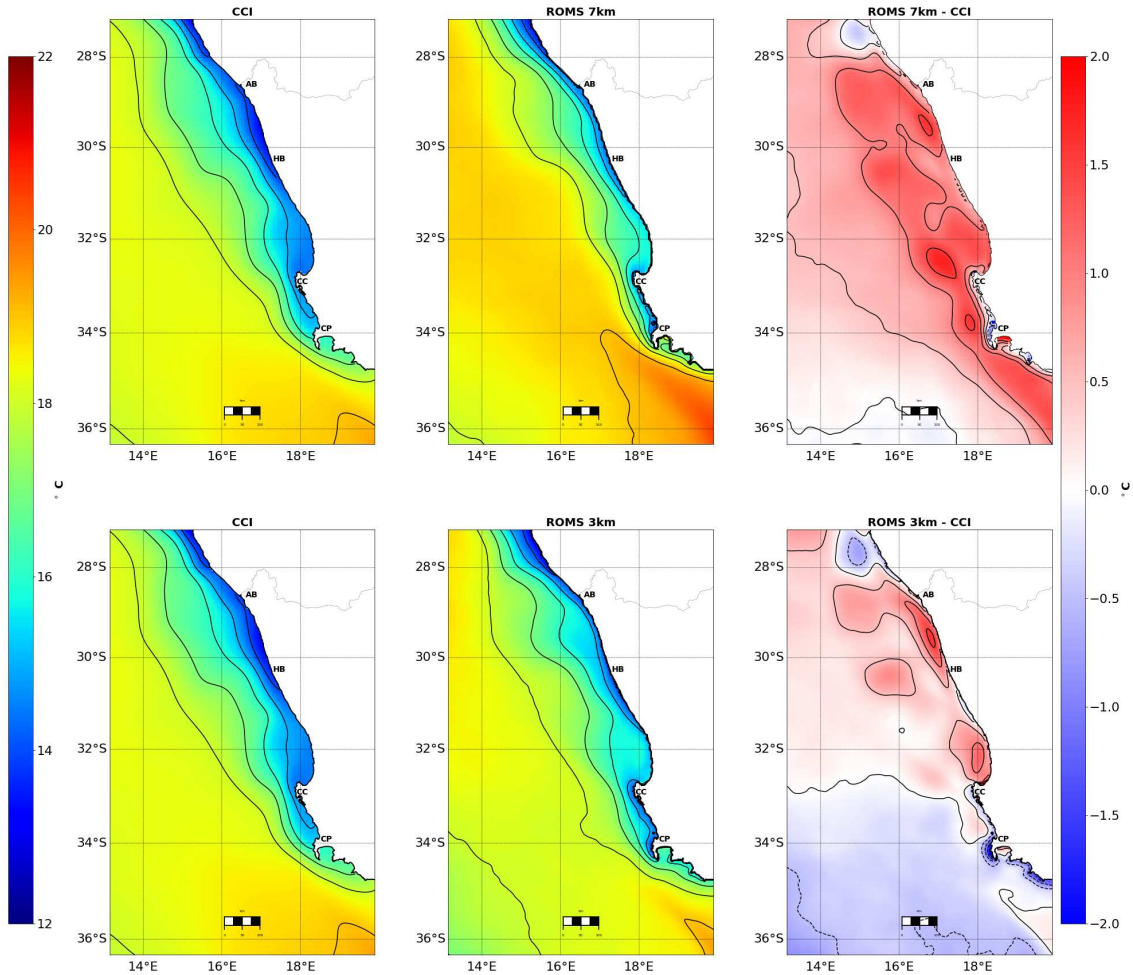


Figure 5.4: Annual average (2005-2012) of merged satellite CCI SST (left), model SST (centre), and SST bias [$SST_{\text{model}} - SST_{\text{CCI}}$] (right). SST contours are superimposed (black solid) with a 1°C contour interval. Top panels are for the R7km simulation whereas bottom panels are for the R3km simulation.

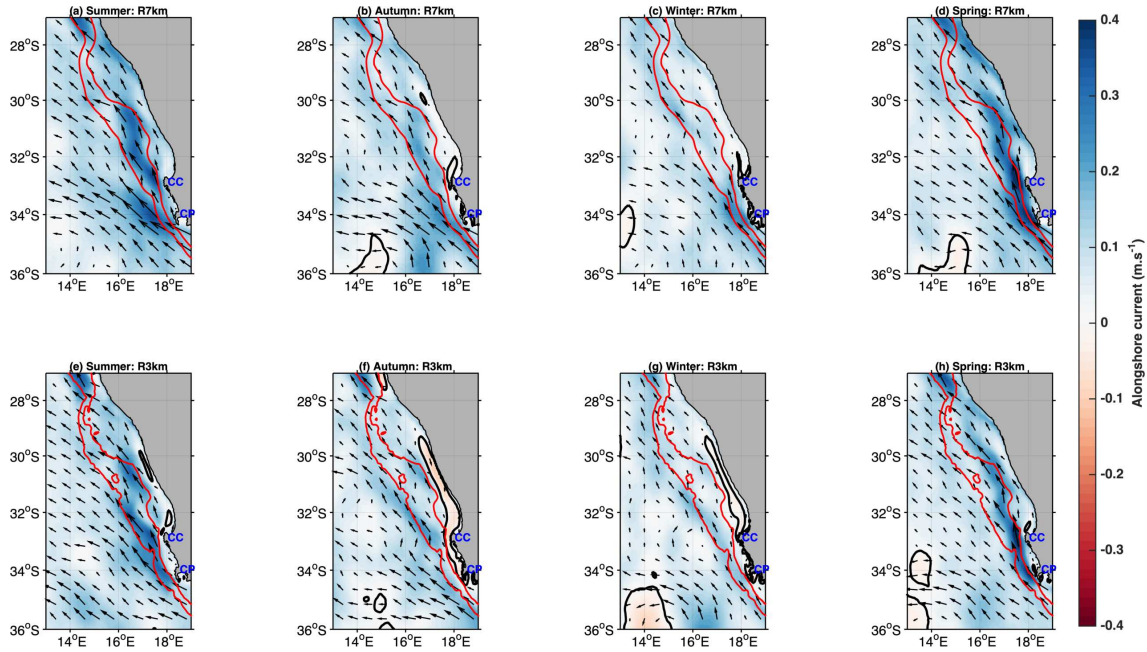


Figure 5.5: Maps of the seasonally averaged alongshore component of surface currents for the R7km simulation (Top panel a-d) and R3km simulation (Bottom panel e-h). The colorbar represents the magnitude of the alongshore component of the velocity vector in m s^{-1} . Current vectors are superimposed, as well as the 200 m and 500 m isobaths (red solid line).

1492 5.3.1 Reduction in the warm bias of the coastal waters

1493 The warm bias of the modelled SST is significantly reduced both in amplitude and spatial
 1494 extension in the finer resolution R3km simulation (Fig. 5.4). It is now mostly confined
 1495 to an innershelf coastal band over depths less than 200 m, and does not exceed 0.5°C
 1496 the offshore domain. Note that this warm bias is also reduced for the coastal sub-surface
 1497 waters located off the continental slope (not shown).

1498 5.3.2 A narrower and smaller Benguela Jet

1499 The simulated mean surface currents look quite similar in the two simulations (Fig. 5.5)
 1500 and compare well with the surface currents derived from altimetry. The Benguela Jet that
 1501 played a key role in the advection of particles from Cape Point to Cape Columbine in
 1502 the R7km simulation is also present in the R3km simulation. However, it is narrower and
 1503 more confined over the shelf-edge. This confinement of the jet over the 200 m - 500 m
 1504 isobaths is even more pronounced in summer, between Cape Point and Cape Columbine.

1505 In the R3km simulation, the jet follows the shelf edge whereas, in the R7km simulation,
1506 it veers offshore and splits into two branches. At the seasonal time scale, the core of the
1507 jet coincides with the location of the maximum across-shore SST gradient in both of our
1508 simulations as observed in the satellite data (Fig. 5.6). This is in accordance with the
1509 findings of *Veitch and Penven (2017)* and shows that the main dynamical ingredients that
1510 control the seasonal cycle of the Benguela Jet are likely to be correctly reproduced in both
1511 simulations. In addition, in both simulations, the jet exhibits a similar seasonal cycle
1512 coherent with the observations. Its intensity peaks at 0.45 m s^{-1} in spring and summer
1513 and weakens to 0.1 m s^{-1} in winter.

1514 Some clear differences between the two simulations can be observed in the across-
1515 shore and vertical structure of this equatorward jet (Fig. 5.7). In the R7km simulation,
1516 the Benguela Jet is characterised by higher speeds than in the R3km simulation. Its core
1517 extends further offshore (200 km in R7km versus 100 km in R3km) and deeper (-300 m in
1518 R7km versus -100 m in R3km). Note that this confinement of the Benguela Jet closer to
1519 the surface is accompanied by an the uplift of the isotherms, which reduces the sub-surface
1520 warm bias to almost zero below -30 m .

1521 **5.3.3 The presence of poleward jets**

1522 Maps of surface currents reveal the existence of a nearshore poleward flow in the R3km
1523 simulation only (Fig. 5.5). This poleward surface countercurrent is evident all along the
1524 coast up to 29°S during summer and autumn. Its speed peaks at 0.1 m s^{-1} at the locations
1525 of the upwelling cells. Vertical transects off Cape Point show that this surface counter
1526 current results from the surfacing of a subsurface poleward jet (Fig. 5.7). The core of this
1527 poleward flow is located in between -50 m and -100 m depth (within 50 km offshore),
1528 which distinguishes it from the all year round persistent poleward undercurrent located
1529 on the shelf slope around -500 m depth. The physical mechanisms driving this poleward
1530 nearshore countercurrent were not studied in this thesis. Additional analysis would be
1531 required to disentangle the role of the cyclonic wind stress curl associated with the coastal
1532 wind drop-off, bottom friction, and modified bottom topography. Across-shore transects
1533 of along-shore currents (Fig. 5.7) show that this sub-surface poleward flow is also present in

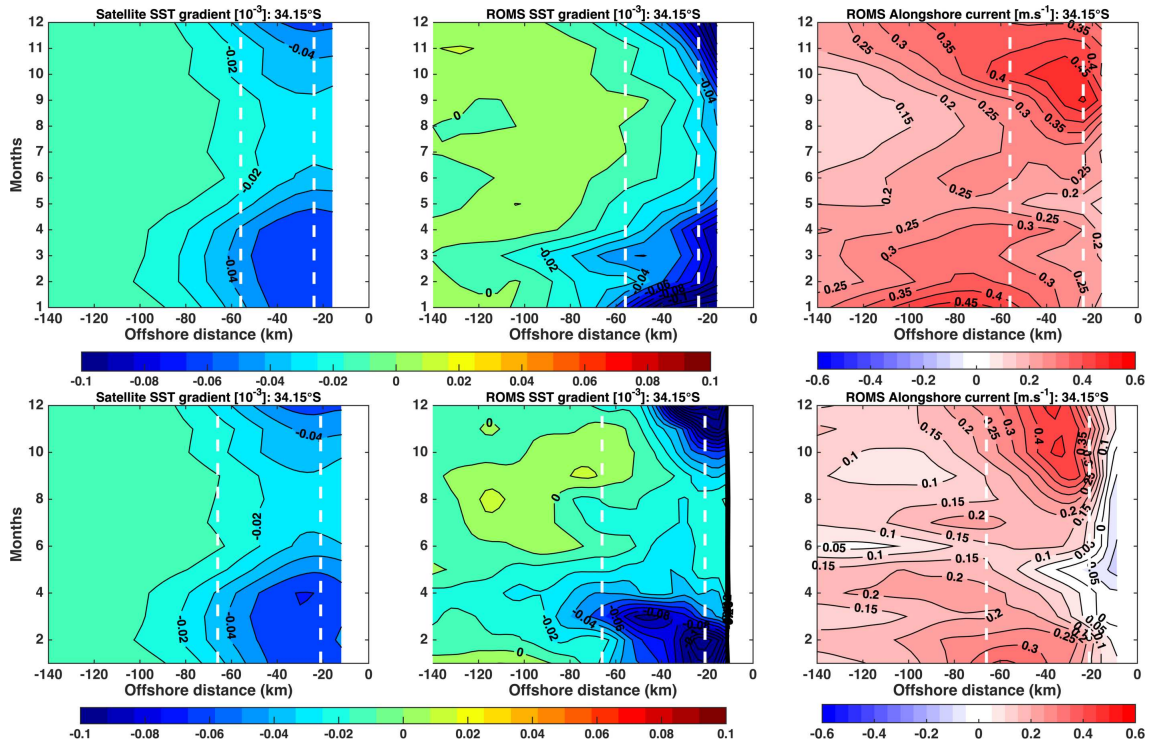


Figure 5.6: Hovmoller plot of: satellite (left), model (middle) cross-shore climatological mean SST gradient off Cape Point (34.15°S), and model monthly climatological mean of surface alongshore current (right panel) for the 7-km (top) and 3-km (bottom) simulations. The contour interval for the SST gradient and model velocity are $0.01^{\circ}\text{C km}^{-1}$ and 0.05 m s^{-1} respectively. The location of the 200 m and 500 m isobaths are shown as white dashed lines.

1534 the lower resolution R7km simulation. However, it is weaker than in the R3km simulation.

Its core is also narrower and never shallows enough to reach the surface.

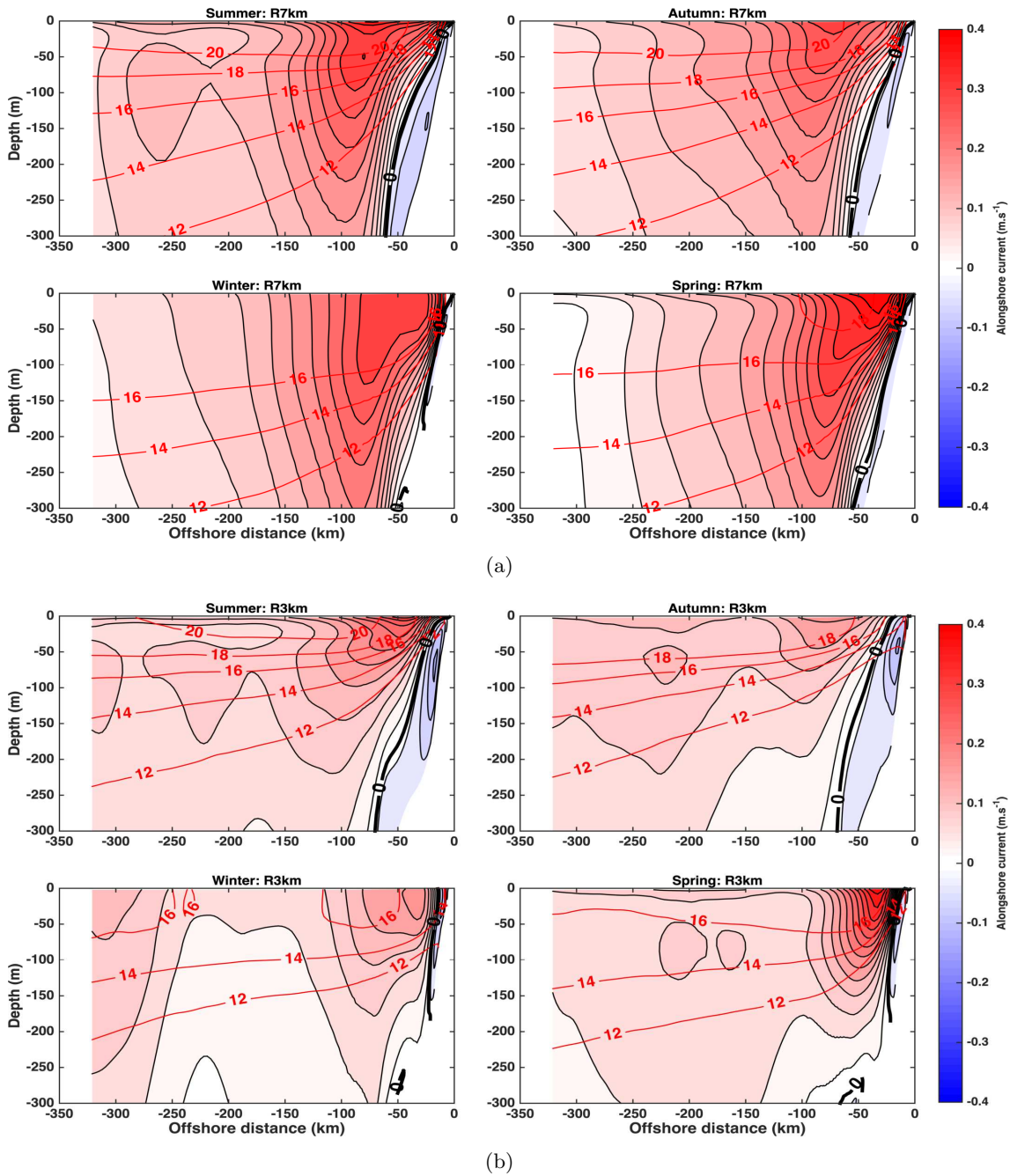


Figure 5.7: Vertical cross-shore transects off Cape Point of the seasonally averaged alongshore component of the model current vector (colorbar in m s^{-1}) from the R7km (a) and R3km (b) simulation. Velocity/temperature iso-contours (black thin/red solid) are superimposed with a $0.1 \text{ m s}^{-1}/2^\circ\text{C}$ contour spacing.

1535

1536 **5.3.4 Impacts of the horizontal resolution on the mesoscale turbulence**
 1537 **in the SBUS**

1538 **Two different mechanisms for the offshore and inshore oceanic turbulence**

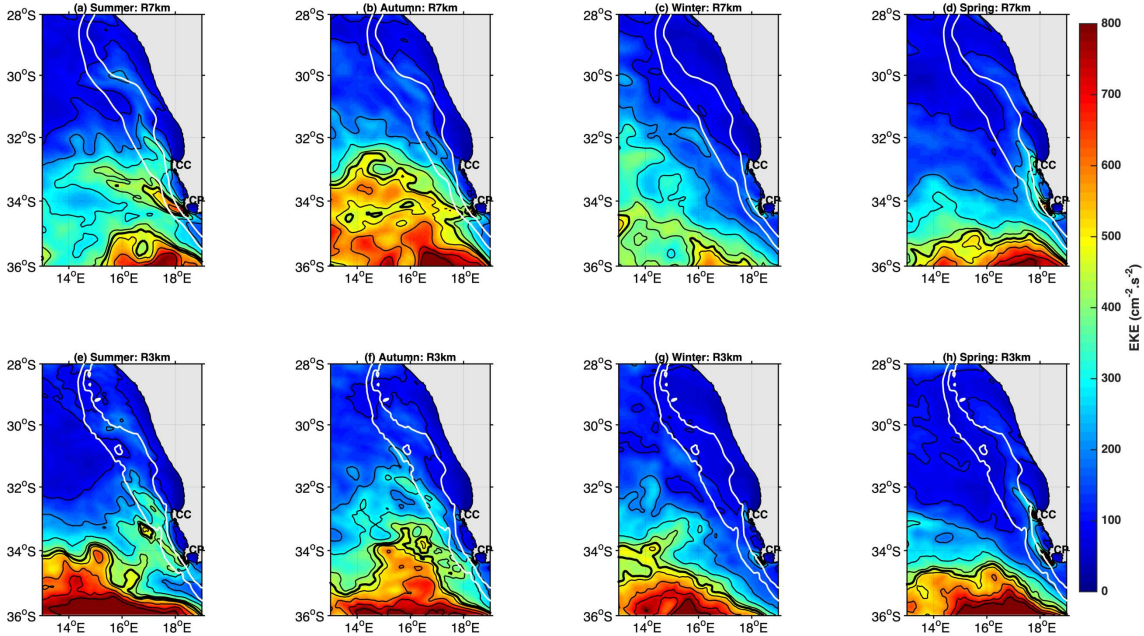


Figure 5.8: Comparison of seasonally averaged surface Eddy Kinetic Energy (EKE) for the: R7km simulation (Top panel a-d) and R3km simulation (Bottom panel e-h). The $450 \text{ cm}^2 \text{ s}^{-2}$ contour is shown in black bold line to delimit the extension of the high values of EKE typically found in region affected by the Agulhas Current retroflexion. The 200 m and 500 m isobaths are shown as white lines.

1539 Despite some differences, the large scale patterns of surface EKE (Fig. 5.8) look similar
 1540 in both simulations. The offshore region is characterised by large EKE values, a testament
 1541 to the passage of Agulhas rings and eddies. In the vicinity of the shelf-edge, some pockets
 1542 of large EKE values can still be observed in the R3km simulation. Nevertheless, they
 1543 tend to be more localised to the lee of Cape Point and Cape Columbine and mainly occur
 1544 during summer and autumn. These large EKE values observed over the shelf-edge are
 1545 likely to be generated by a different mechanism, *i.e.* the barotropic/baroclinic instabilities
 1546 of the across-shore density front reinforced during the upwelling season. Therefore, the
 1547 phenomenology of the mesoscale turbulence is likely to be different between the offshore
 1548 and nearshore region. Snapshots of surface relative vorticity¹ ξ (Fig. 5.9) and temperature
 1549 (not shown) show that offshore eddies are much larger than nearshore eddies. The former

¹ $\xi = \delta_x v - \delta_y u$

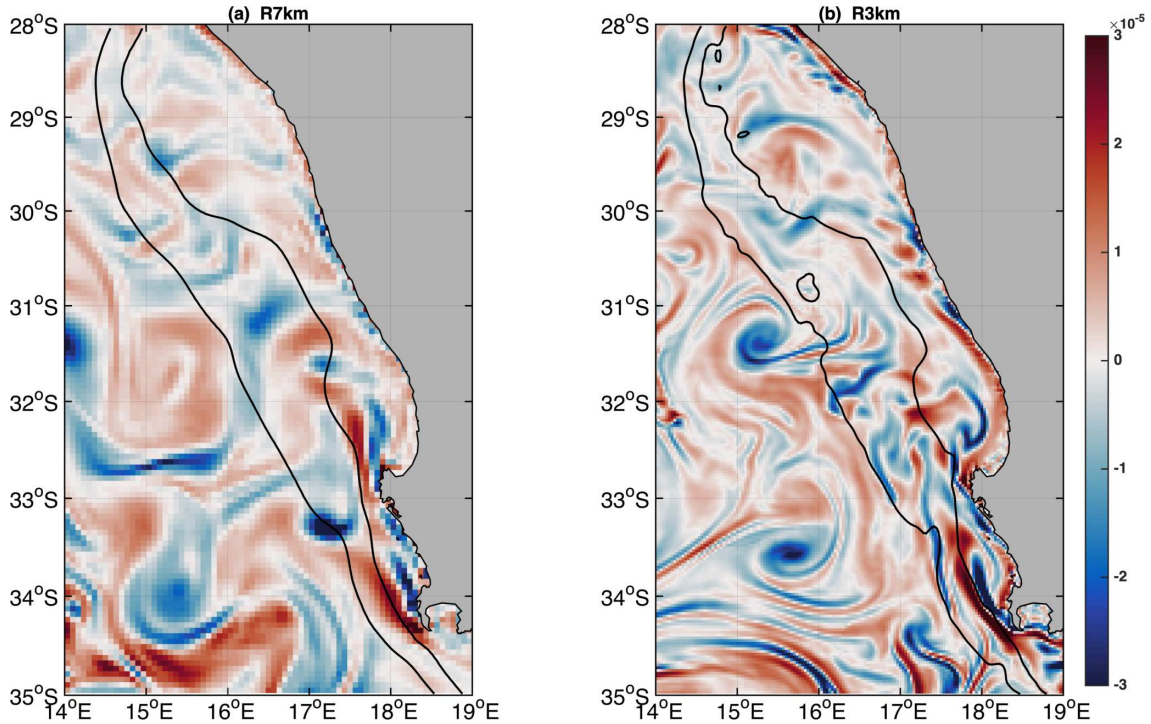
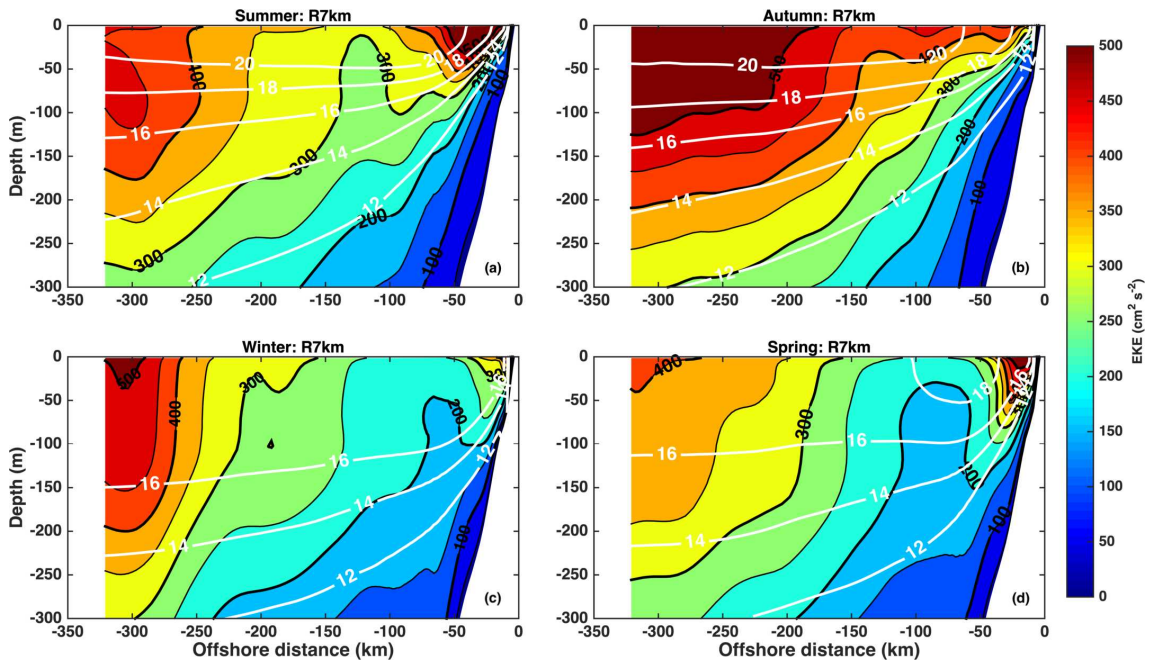


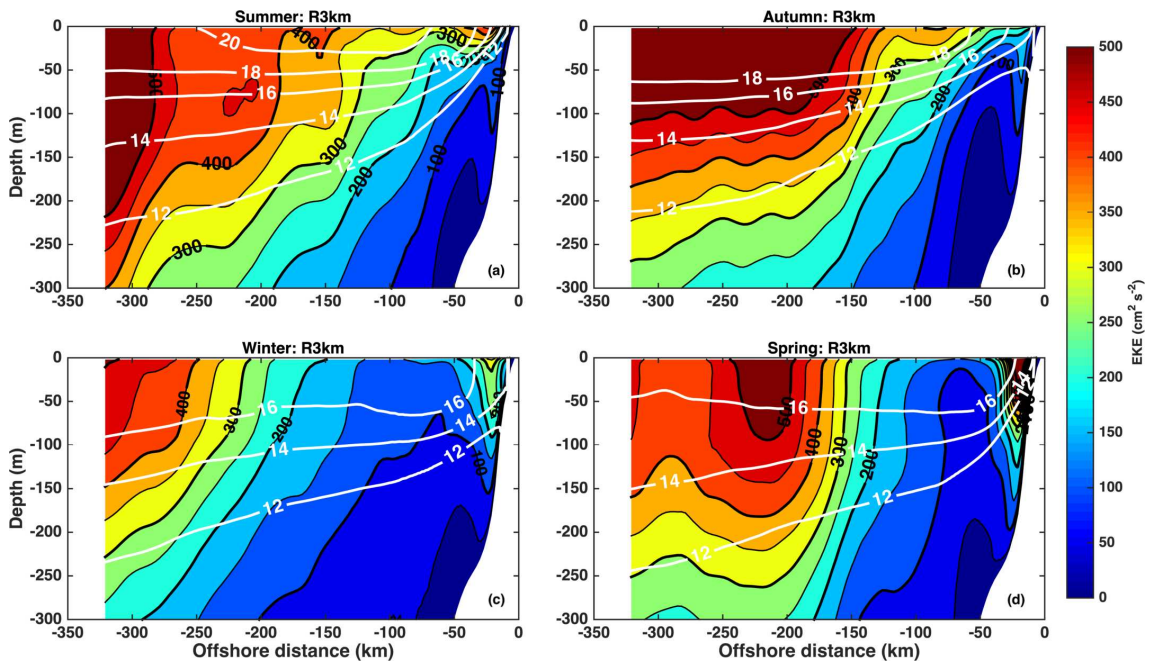
Figure 5.9: Snapshots of surface relative vorticity anomaly in the (a) R7km and (b) R3km simulation on the 1st of October 2010.

1550 are a mixed of large ($R_v \simeq 100$ km) anticyclonic eddies associated with a warm signature
 1551 in SST and smaller ($R_v \simeq 50$ km) cyclonic eddies (Fig. 5.9), both kinds originating pre-
 1552 dominantly from the Agulhas retroflection. On the other hand, the more inshore located
 1553 mesoscale eddies originate from the shelf-edge.

1554 Vertical across-shore transects of EKE also reveal some striking differences between the
 1555 vertical extension of inshore and offshore eddies. The offshore EKE patch extends much
 1556 deeper than the inshore EKE patch as shown by the $300 \text{ cm}^2 \text{ s}^{-2}$ EKE contour (Fig. 5.10).
 1557 In the R3km simulation, the latter contour is found as deep as -300 m offshore whereas as it
 1558 is confined to the first 150 m of the water column on the shelf edge. In addition, the position
 1559 of the inshore patch coincides with the very localised region situated at the interface
 1560 between the core of the Benguela Jet and the core of nearshore sub-surface poleward
 1561 current. This area is likely to be favourable for the occurrence of barotropic/baroclinic
 1562 instabilities. In the following, we choose the $450 \text{ cm}^2 \text{ s}^{-2}$ surface EKE contour (Fig. 5.8) to
 1563 delineate the region directly influenced by the very energetic turbulence originating from
 1564 the Agulhas Current retroflection (Loveday *et al.*, 2014) from the inshore region impacted
 1565 by the instabilities of the along-shore coastal jets. The two regions appear more clearly



(a)



(b)

Figure 5.10: Same as Figure 5.7 but for EKE. Units in $cm^{-2}.s^{-2}$ and the contour interval is $50 cm^2 s^{-2}$. Temperature contours are also superimposed in white with contour interval of $2^{\circ}C$.

1566 separated in the higher resolution simulation: the isoline gets much closer to the shelf
1567 in the R7km simulation than in the R3km simulation, in particular during summer and
1568 autumn when large EKE values are also found nearshore.

1569 **The presence of finer and more intense mesoscale eddies**

1570 As expected, the intensity of the mesoscale eddies increases when the resolution is increased
1571 from 7-km to 3-km, especially in the offshore turbulent region, referred to in the literature
1572 as the 'Cape Cauldron' (*Boebel et al., 2003*) (Fig. 5.8b). This intensification is visible both
1573 in the mean EKE values observed offshore and the absolute values of relative vorticity
1574 found within these eddies. It could be attributed to the enhancement of the dynamical
1575 potential vorticity barrier that surrounds mesoscale eddies and prevents them from being
1576 deformed and/or destroyed by other neighbouring eddies. Consequently, eddies become
1577 more coherent and live longer (*Lévy et al., 2010, 2012*).

1578 As expected, spatial spectrum of vorticity and EKE (not shown) show that the slope
1579 of spectrum in the inertia range decreases as the spatial resolution increases, with cut-
1580 off wave number shifting towards larger values (20 km in R3km versus 50 km in R7km)
1581 confirming the presence of finer scale circulation features.

1582 **5.3.5 An unexpected impact on the Agulhas leakage**

1583 Increasing the model resolution from 7-km to 3-km has led to a better representation of the
1584 cross-shelf temperature gradient and has reduced the warm bias between simulated and
1585 observed temperature both for surface and sub-surface waters (Fig. 5.4). The following
1586 section intends to show that a reduction of the volume of Agulhas waters leaking into the
1587 Cape Basin in the higher resolution simulation could explain the presence of colder water
1588 offshore.

1589 Quantifying the Agulhas leakage into the Cape Basin is challenging due to its turbulent
1590 nature. The latter occurs predominately through the shedding of Agulhas rings at a rate
1591 of 5-6 per year (*Schouten et al., 2000; van Aken et al., 2003; Doglioli et al., 2006*). In
1592 fact, each eddy transports large volumes of Indian Ocean waters, typically 0.5-1.5 Sv per
1593 eddy. Numerous attempts have been made to estimate this leakage using both Lagrangian

1594 (*Donners and Drijfhout, 2004; Doglioli et al., 2006; Speich et al., 2006; Richardson, 2007;*
 1595 *van Sebille et al., 2010; Durgadoo et al., 2013; Rhüs et al., 2013*) and Eulerian (*Gordon,*
 1596 *1986; Gordon et al., 1987; Garzoli and Goni, 2000; Boebel et al., 2003; Reason et al.,*
 1597 *2003; Treguier et al., 2003; van Sebille et al., 2009, 2010; Loveday et al., 2014*) methods.
 1598 Estimations vary from 10 Sv to 41 Sv, but mostly fall in the range of 10-20 Sv with a large
 uncertainty.

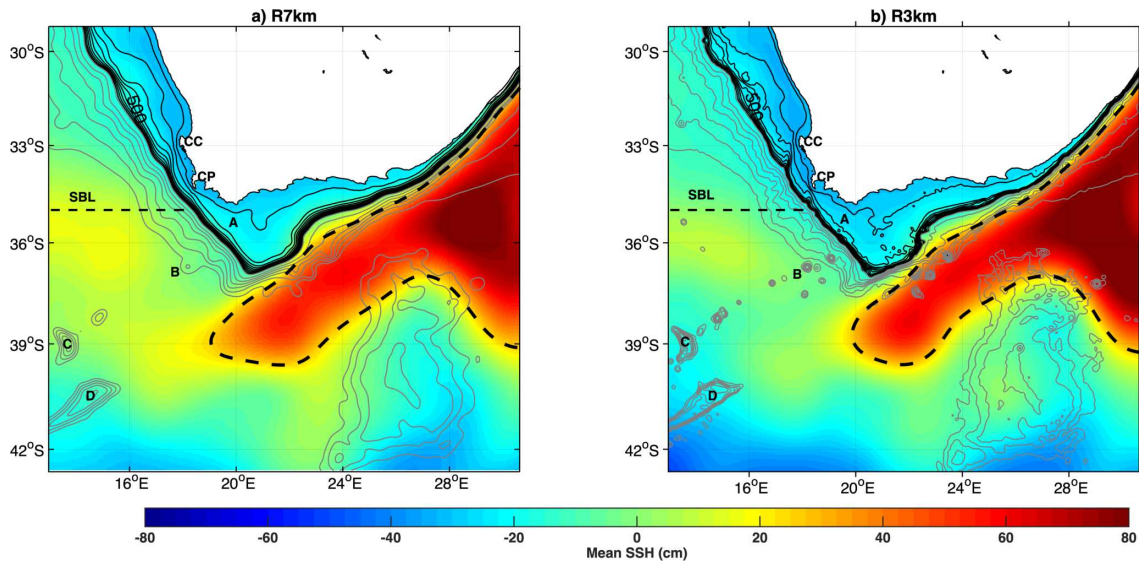


Figure 5.11: Mean annual SSH (color, in metres) over the 2005-2012 period for a) R7km and b) R3km simulation. The bold dashed line show the model retroreflection path of the Agulhas Current derived from the annual mean of the same period. The dashed black transect shows the SBUS southern boundary line (SBL) over which the meridional heat fluxes are estimated. CC stands for Cape Columbine and CP for Cape Point, the southernmost tip of the Cape Peninsula. the Black contours represent bathymetry shallower than 1000 m at 100 m depth intervals, with the grey lines representing bathymetry deeper than 1000 m at 500 m depth intervals. Letters label the main bathymetric features namely: **A**-Agulhas Bank, **B**-Protea and Argentina Seamounts, **C**-Erica Seamount and **D**-Agulhas Ridge.

1599

1600 Here, the Agulhas Current leakage into the SBUS is estimated by calculating the
 1601 volume transport using the monthly climatological mean meridional velocities across a
 1602 zonal transect taken at 35°S hereafter referred to as the southern boundary line (SBL).
 1603 The velocities are integrated over the top 1000 m of the water column between 12°E and
 1604 18°E. The estimated value of the meridional transport is relatively higher in the 7-km
 1605 reference simulation where it is 18.3 Sv and 16.3 Sv in the 3-km resolution simulation.
 1606 Both estimates fall within the mostly observed 10-20 Sv range (*Gordon, 1986; Gordon*

1607 *et al.*, 1987; *Garzoli and Goni*, 2000; *Boebel et al.*, 2003; *Reason et al.*, 2003; *Treguier*
 1608 *et al.*, 2003; *van Seville et al.*, 2009, 2010; *Loveday et al.*, 2014).

1609 The heat transport is also estimated across the SBL transect and Fig 5.12 shows the
 1610 vertically integrated meridional heat transport Q estimated as follows:

$$Q(lon, lat) = \int_{z=-1000 \text{ m}}^{z=0} \rho_o C_p \bar{v}(lon, lat) \bar{\theta}(lon, lat) \partial z \partial x' \quad (5.1)$$

1611 where ρ_o is the potential density of seawater, C_p the specific heat capacity, $\bar{v}(lon, lat, z)$ the
 1612 annual mean meridional velocity and $\bar{\theta}(x, y, z, t)$ the annual mean potential temperature.

At the SBL, the across-shore dependency of heat transport is very different in the two

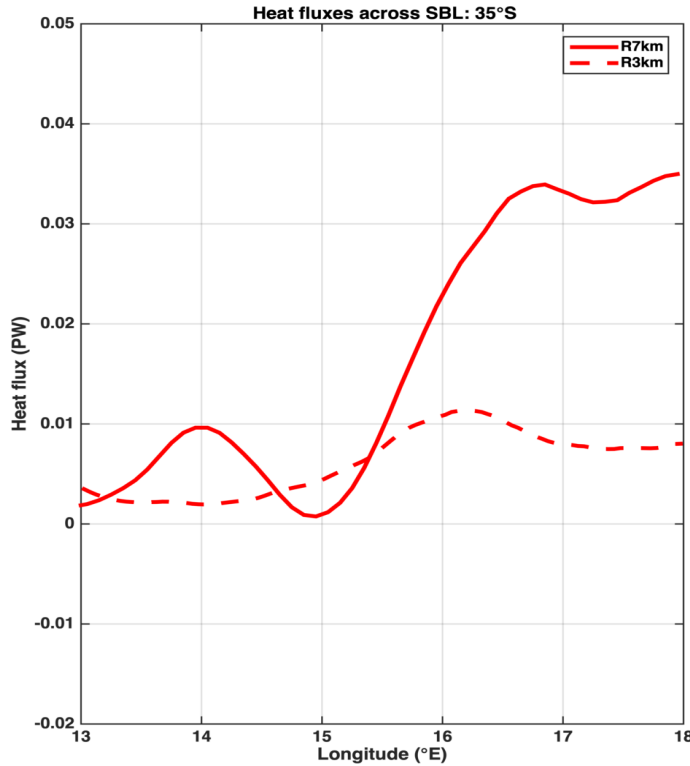


Figure 5.12: Mean annual depth-integrated (upper 1000 m depth) heat fluxes (PW) across SBL (35°S) transect for the R7km (solid red line) and R3km (dashed red line) simulations. 1 PW=10¹⁵W.

1613

1614 simulations (Fig. 5.12). In the reference simulation, it is much more intensified in the
 1615 eastern part of the transect between 16°E and 18°E. Using altimetry data, *Dencausse et al.*
 1616 (2010c) identified three routes followed by the Agulhas rings through the Agulhas corridor
 1617 (Fig. 5.11): i) a northern route between the Agulhas Bank (**A**) and the Erica seamount
 1618 **C**; ii) a central route between the Erica Seamount (**C**) and the Agulhas Ridge (**D**); iii)

1619 a southern route for eddies that initially track south and then pass over the Agulhas
1620 Ridge. Our estimated transports at the SBL suggest that there is an overestimation of
1621 the transport of Agulhas waters into the Cape basin through the northern route in our
1622 reference simulation.

1623 The high fluxes near 14°E and 17°E in the R7km (Fig. 5.12) could be due to a lack
1624 of topographical shielding by a group of isolated seamounts extending south-west of the
1625 Agulhas Bank (Fig. 5.11). In fact, increasing the resolution results in a much better
1626 representation of some important topographic features such as the Erica Seamount and a
1627 group of seamounts (**B**) south-west of the Agulhas Bank (**A**). These isolated seamounts
1628 disseminated along the Agulhas Rings corridor in the R3km simulation may block the
1629 northwestward propagation of Agulhas rings, and in turn reduce the inflow of warm waters
1630 into the Cape Basin (*Arhan et al.*, 1999).

1631 In addition, the position of the Agulhas rings retroflexion is located further east in
1632 the R3km simulation than in the R7km simulation (Fig. 5.11), in accordance with satellite
1633 sea surface height observations showing that a more westward (eastward) position of the
1634 Agulhas retroflexion results in an enhanced (reduced) Agulhas leakage (*van Sebille et al.*,
1635 2009). This improvement of the position of the Agulhas retroflexion could also be linked
1636 to the presence of a more realistic bathymetry in the high resolution simulation (*Speich*
1637 *et al.*, 2006).

1638 **5.4 Coastal circulation in the R3km simulation: summary**

1639 Increasing the ocean model horizontal resolution from 7 km (R7km) to 3 km (R3km)
1640 results in a net decrease of the SBUS warm SST bias in the simulation. The reduction
1641 in the offshore warm SST bias in the R3km could be explained by the reduced heat flux
1642 into the Cape basin due to a reduced Agulhas leakage of warm Indian ocean waters. The
1643 increased resolution also allows finer details of the coastal topography to be represented,
1644 and this leads to a coastal jet that is narrower and more confined towards the shelf-edge.
1645 Moreover, an increase in inner shelf poleward flow is evident in the higher resolution
1646 simulation. Further improvements of the mesoscale eddy structures increased the EKE
1647 associated with the passage of Agulhas rings and eddies. However, this high offshore EKE

1648 activity is separated from the inshore region in the R3km. There is also a difference in
 1649 the magnitude of inshore EKE associated with the shear instabilities of the Benguela Jet
 1650 and poleward current.

1651 The changes in the mean circulation are expected to have an impact on the alongshore and
 1652 cross-shore transport of Lagrangian particles, and thus transport success. The robustness
 1653 of seasonal characteristics of transport success of particles advected by velocity fields from
 1654 the R7km simulation (Chapter 4) is tested in the next section. The transport success in
 1655 the two simulations is compared.

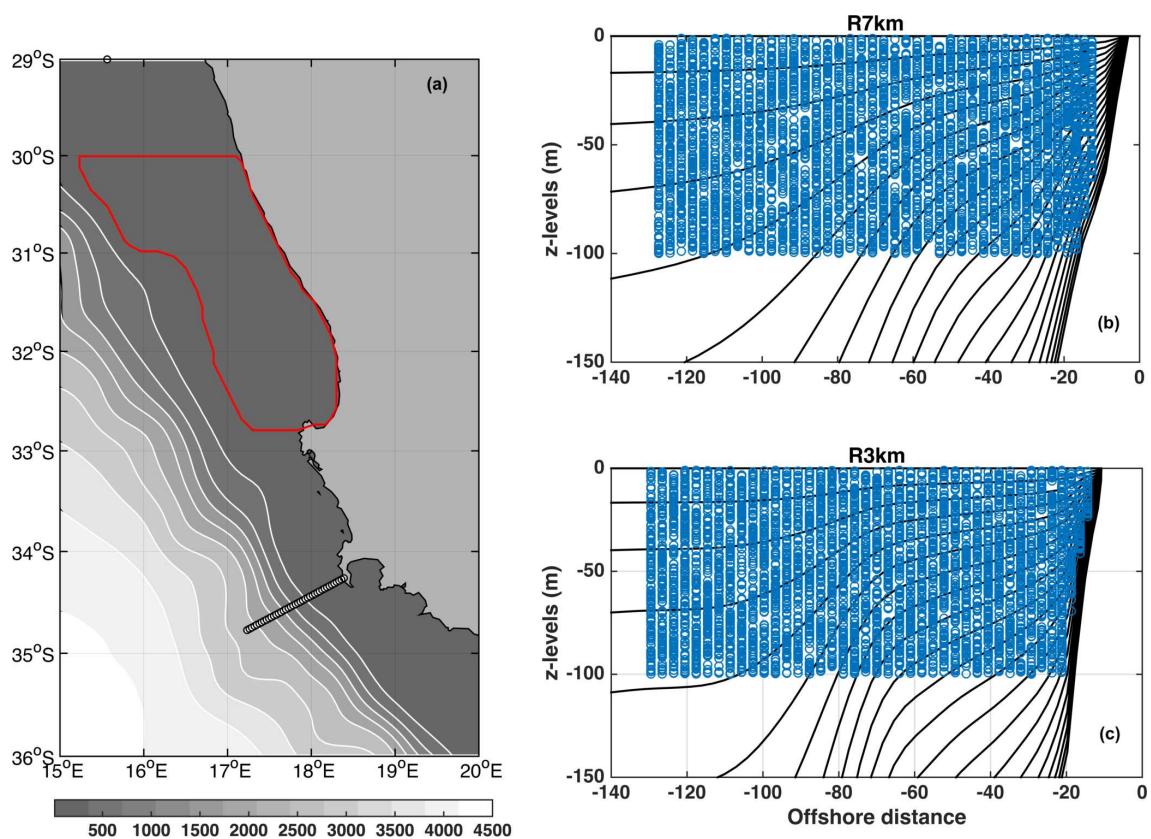


Figure 5.13: a): Set-up of the Lagrangian experiments: particles are released every 12 hours over the SARP line (white circles), a cross-shore transect off Cape Point. They are counted as successfully transported particles if they reach, within 30 days, the St Helena Bay target area delineated by the red polygon. The bathymetry of the R7km model (in grey with white contours every 500 m) is also superimposed. (b,c): Initial distribution of particles over the SARP line for a given release event for R7km and R3km, respectively. σ layers are superimposed (black solid).

1656 **5.5 Sensitivity of the Lagrangian transport success to the** 1657 **downscaling**

1658 Given the importance of the oceanic circulation on the alongshore and cross-shore trans-
1659 port of Lagrangian particles, the impact of downscaling on the Lagrangian pathways is
1660 investigated in this section. Here, we study how the changes of the underlying Eulerian
1661 fields impact the Lagrangian trajectories between Cape Point and St Helena Bay. To be
1662 coherent with the previous chapter, the focus will be on the seasonal pattern.

1663 **5.5.1 The Lagrangian setup**

1664 The same methodology as in Chapter 4 is used. The particles were released every 12 hours
1665 from 1992 to 2011 at a transect off Cape Point (SARP line) over 108 vertical profiles dis-
1666 tributed every 3 km in horizontal distance (Fig. 5.13a). For each vertical profile, particles
1667 are randomly distributed over the first 100 m depth with a density of 1.125 particles per
1668 metre. This resulted in a total of 4976 for the R7km simulation (Fig. 5.13b) and 4840 for
1669 the R3km simulation (Fig. 5.13c). The lower number of particles in R3km is attributed to
1670 the shape of the shelf which is wider and shallower than in R7km and thus less particles
1671 are released over the inner shelf region. The particles are advected for 30 days by daily
1672 velocity fields provided by the R7km and R3km simulations. Transport success is given
1673 by the ratio of the number of particles reaching St Helena Bay target area over the total
1674 number released. This area is defined as a polygon that extends from the coast to the
1675 300 m isobath and from 32.8°S to 30°S (Fig. 5.13a).

1676 **5.5.2 Comparison of the alongshore connectivity between R7km and** 1677 **R3km**

1678 Transport success in both simulations exhibits the same seasonal cycle, which is highest
1679 from August to December (Fig. 5.14a,b). Transport success is generally higher in the
1680 R3km with the exception of May. The largest differences in transport success between
1681 the two simulations occurs during the peak season (September-November). The lowest
1682 transport success in the R7km occurred in June whereas it occurred in May in the R3km.
1683 In the R7km, the standard deviation is lowest during spring (Fig. 5.14b) and in autumn

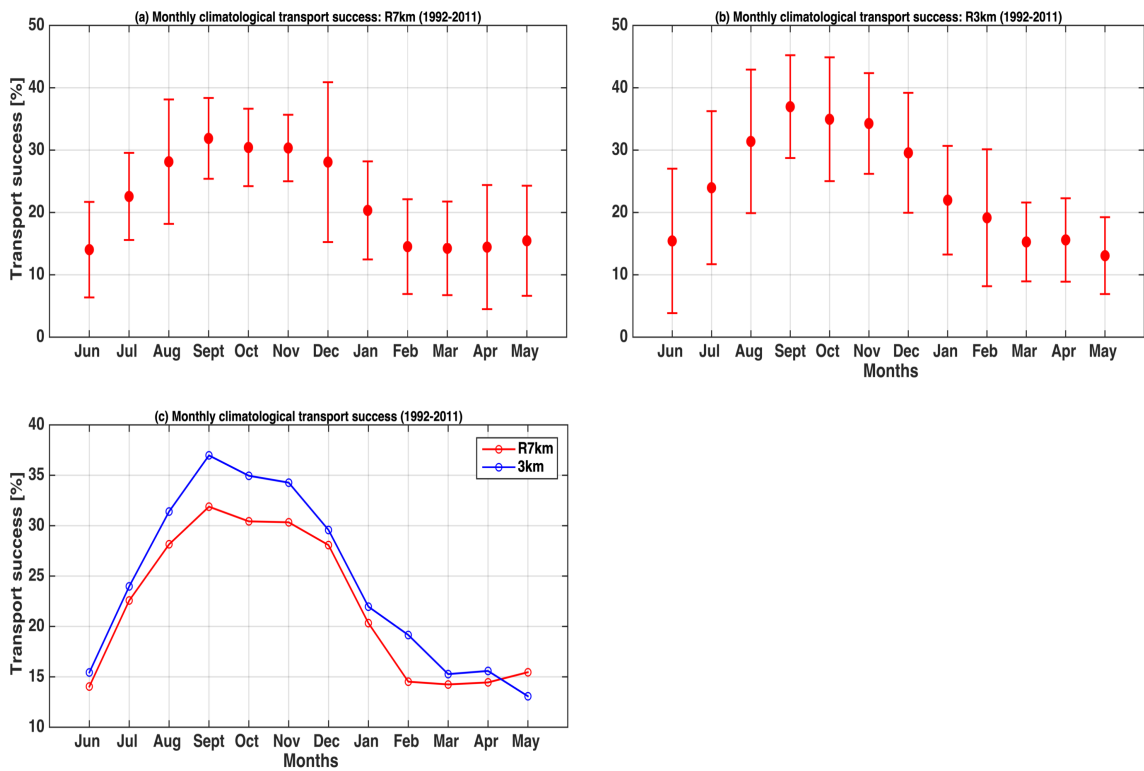


Figure 5.14: Monthly climatology of transport success TS_m (red dots) and standard deviation of TS_m superimposed as error bars for a) R7km and b) R3km. c) Monthly climatology of transport success for R7km (red line) and R3km (black line).

1684 (March-May) in the R3km (Fig. 5.14c).

1685 Figure 5.15 compares the seasonal age of arrival in the target area. There is a peak
1686 in the age distribution between five to ten days for particles released between July and
1687 December in the R7km (Fig. 5.15a), while it peaked during September-December in the
1688 R3km (Fig. 5.15b). For the R3km, the oldest particles (20-30 days) have better probability
1689 of transport success from the late upwelling season to winter (January to July).

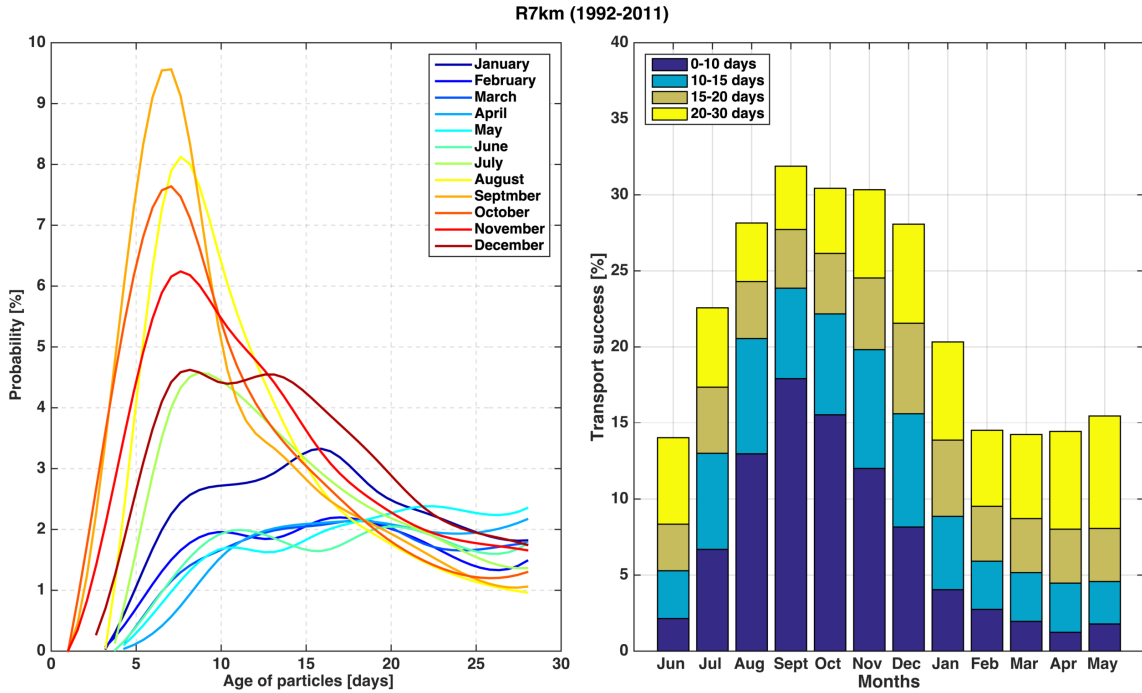
1690 Particles are then grouped according to the isobath on which they have been released,
1691 and this was done by grouping them as inshore and offshore particles. The 500 m isobath
1692 is used as the border between the inshore and offshore region. For both simulations,
1693 Figure 5.16a reveals that most of the successful particles originate from the inshore region,
1694 and their seasonal pattern is similar to that of the total transport success. In the R3km,
1695 this proportion is much smaller during autumn and early winter (March-July).

1696 To evaluate the influence of the coastal poleward current on transport success, the inshore
1697 transport success was further divided into groups of inner shelf (<200 m isobath) and
1698 mid-shelf (>200 m-500 m isobaths). Their seasonal probability distribution (black and
1699 red lines) is superimposed on Fig. 5.16a). There are less successful inner shelf particles
1700 in the R3km than R7km throughout the year particularly during winter and spring. This
1701 result suggests that it is the mid-shelf particles rather than the inner shelf particles that
1702 contribute significantly to the inshore transport success in the R3km simulation during
1703 this period.

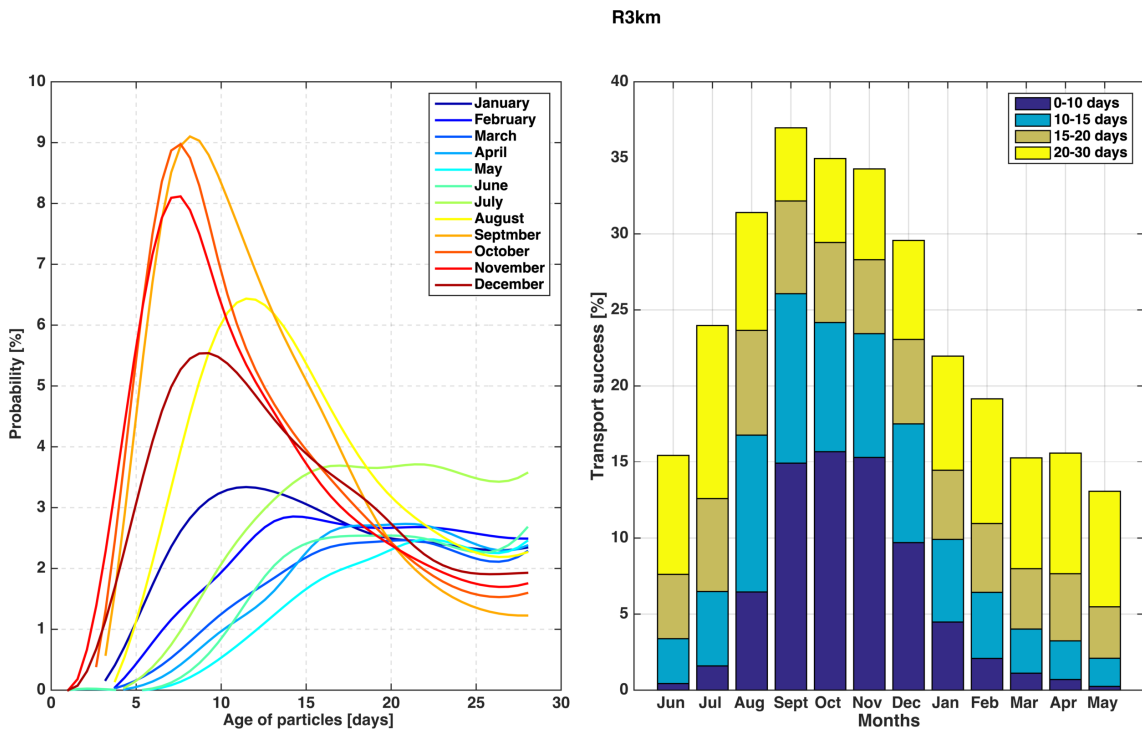
1704 The opposite is observed in the offshore region, where the particles originating from this
1705 region have a relatively low chance of reaching St Helena Bay throughout the year, the
1706 probability being relatively lower in the R7km than in the R3km (Fig. 5.16b). Moreover,
1707 the seasonal cycle of the offshore transport success is different to the inshore particles as
1708 it peaks in November to February.

1709 The most distinguishable impact of increasing the model resolution on the alongshore
1710 connectivity is that transport success increases when increasing the model resolution but
1711 only between August and February for inshore particles. The contribution of this inshore
1712 transport success mainly comes from the particles from the mid-shelf region under the
1713 influence of the Benguela Jet. In R7km, there is a high inner shelf transport success

1714 during autumn and winter. There is an increase in transport success of offshore particles
1715 in the R3km which resulted in the high total transport success from March to July despite
1716 the low inshore transport success during this period. This could explain why there is
1717 an increase in the proportion of old particles (20-30 days) because offshore particles are
1718 expected to take longer than those travelling over the shelf in the Benguela Jet. The effect
1719 of the model resolution on the patterns of Lagrangian trajectories is presented next.



(a)



(b)

Figure 5.15: Left: Monthly distribution of successful particles by age when entering the target area. Right: Cumulative histogram of monthly climatology transport success with clusters of particles built according to their age of arrival: group I (age<10 days); group II (10<age≤15 days); group III: (15<age≤20 days); group IV: (20<age≤30 days). For R7km (a) and R3km (b). Probability have been normalised by the total number of particles released for each month m ($m = 1, \dots, 12$) over the 20 years of the model run.

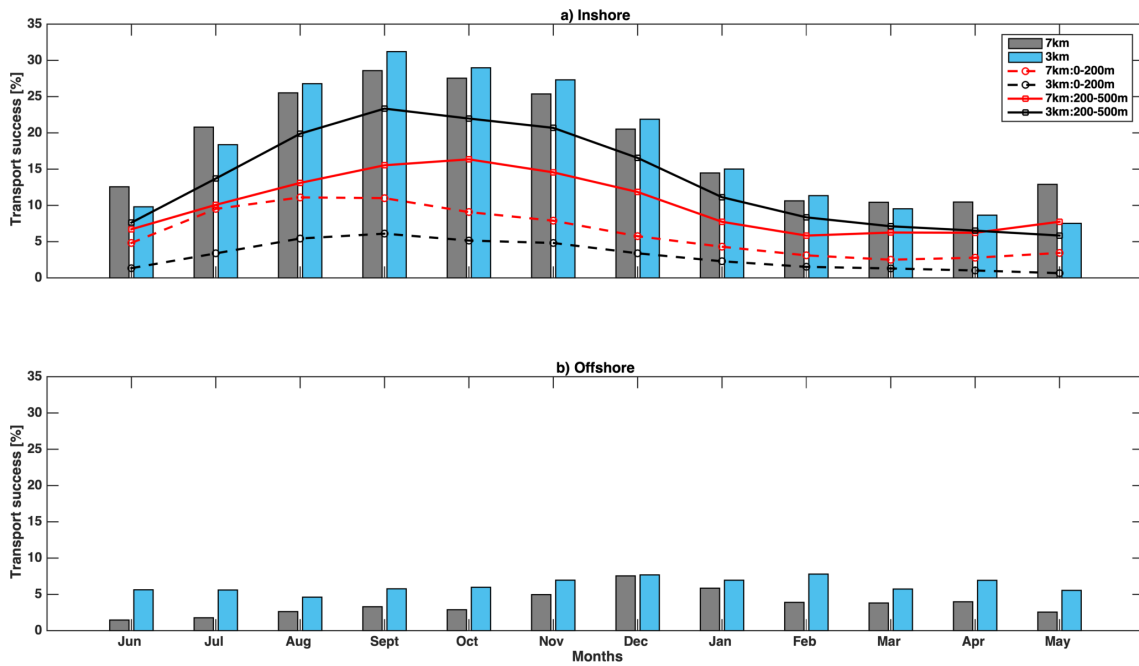


Figure 5.16: Monthly cumulative contribution of the (a): inshore particles and (b): offshore particles to transport success for the R7km (grey bars) and R3km (cyan bars). The monthly transport success of the particles originating over the inner shelf (0-200 m isobath) is also superimposed on (a) for the R7km (red dashed line) and R3km (black dashed line). The red and black lines respectively represent the transport success for particles from the mid-shelf (200-500 m isobaths) in the R7km and R3km.

1720 **5.6 Horizontal dispersion**

1721 The sensitivity of particle horizontal dispersion to the differences in the structure of the
1722 alongshore currents and in the mesoscale variability between the two simulations is anal-
1723 ysed. To characterise the horizontal dispersion of particles, the maps of particles' spatial
1724 distribution (plumes) are plotted according to their age τ (in days) on the horizontal grid
1725 of the R7km and R3km simulations.

1726 All particles aged τ found within the water column of the grid cell located at position
1727 (x, y) in Cartesian coordinates are counted. Particle density is calculated by dividing
1728 the total number of particles within each grid cell by the grid cell surface area. The
1729 seasonality of the dispersion patterns is considered by averaging the monthly climatology
1730 according to the four austral seasons: summer (December-February), autumn (March-
1731 May), winter (June-August) and spring (September-November). To study the differences
1732 in the evolution of the plume, maps of particles' spatial distribution are shown at 6 days
1733 and at 12 days after their release for unsuccessful particles (Fig. 5.17) and successful
1734 particles (Fig. 5.18).

1735 **Unsuccessful particles**

1736 The north-westward wide spread of the highest plume density ($>4.5 m^{-2}$) of unsuccessful
1737 particles into the open ocean in summer and autumn is greatest in the R7km (Fig. 5.17,
1738 panel a-b) and greater than in the R3km (Fig. 5.17, panel e-f). At 6 days, this plume
1739 of particles had travelled further westward across 16°E in the R7km (Fig. 5.17a, panel
1740 a-b) than in the R3km (Fig. 5.17a, panel e-f), although it appears more clustered during
1741 the other seasons (winter and spring). The size difference in the plume between summer-
1742 autumn (bigger) and winter-spring (smaller) is more apparent in R3km than in the R7km.
1743 A reference contour (blue line) of $3.5 m^{-2}$ is also superimposed on the distribution maps
1744 to estimate the spreading of particles in the two simulations. Figure 5.17b shows that
1745 the north-westward spreading that is enhanced at 12 days is minimised with increasing
1746 resolution. It is also apparent that increased resolution leads to the plume exhibiting
1747 more variability characterised by thinner filamentary structures suggesting that smaller
1748 structures are resolved by the R3km simulation. Closer to the coast, a poleward transport

1749 of particles can also be noticed in both simulations with the largest transport occurring in
1750 summer and autumn. There are more particles advected south in the R3km simulation,
1751 and this explains why transport success of the inner shelf particles (Fig. 5.16a) is much
1752 smaller in the R3km simulation than in R7km. Interestingly, the high resolution also
1753 allows for the effective transport of high density plume into False Bay (east of Cape
1754 Point) throughout the year except in spring, this indicates that the poleward flow enters
1755 the bay rather than passing offshore of it.

1756 **Successful particles**

1757 In both simulations, the highest density of successful particles shaped like a alongshore
1758 filament is confined between the 200 m and 2000 m isobaths (Fig. 5.18). At 6 days, the
1759 plume is located between Cape Point and Cape Columbine (Fig. 5.18a). In winter and
1760 spring, the plume travels along shelf with some of the particles located in the inner shelf
1761 region and less particles offshore of the 2000 m isobath. On the other hand, there are
1762 less particles located in the inner shelf in autumn and summer, likely due to the increased
1763 southward transport seen in Figure 5.17a (panel a-b and e-f); more so in the R3km sim-
1764 ulation (Fig. 5.17a, panel e-f). Also, there is a plume of $<3.5 m^{-2}$ particles travelling
1765 offshore of the 2000 m isobath at 6 days that moves back towards the shelf at 12 days
1766 during summer. This implies that successful particles take a more offshore route during
1767 summer during the first 6 days hence the long travel time. At 12 days, the plume has
1768 been advected north past Cape Columbine (Fig. 5.18b), and it has propagated northward
1769 much faster in the R7km in autumn and winter (Fig. 5.18b, panel b-c) related to a much
1770 stronger Benguela Jet than in the R3km (Fig. 5.18b, panel f-g).

1771

1772 In conclusion, successful particles tend to travel between 200 m and 500 m isobaths but
1773 the north-westward dispersion at early ages during summer and autumn results in some
1774 successful particles travelling offshore. Hence, the particles are characterised by a longer
1775 connection time (20-30 days). However, there are relatively more offshore particles that
1776 become successful in the R3km because they do not get advected long distance offshore
1777 into the turbulent region. In the R3km, there is also a reduced transport success of inshore

1778 particles in summer and autumn due to the southward transport on the shelf.

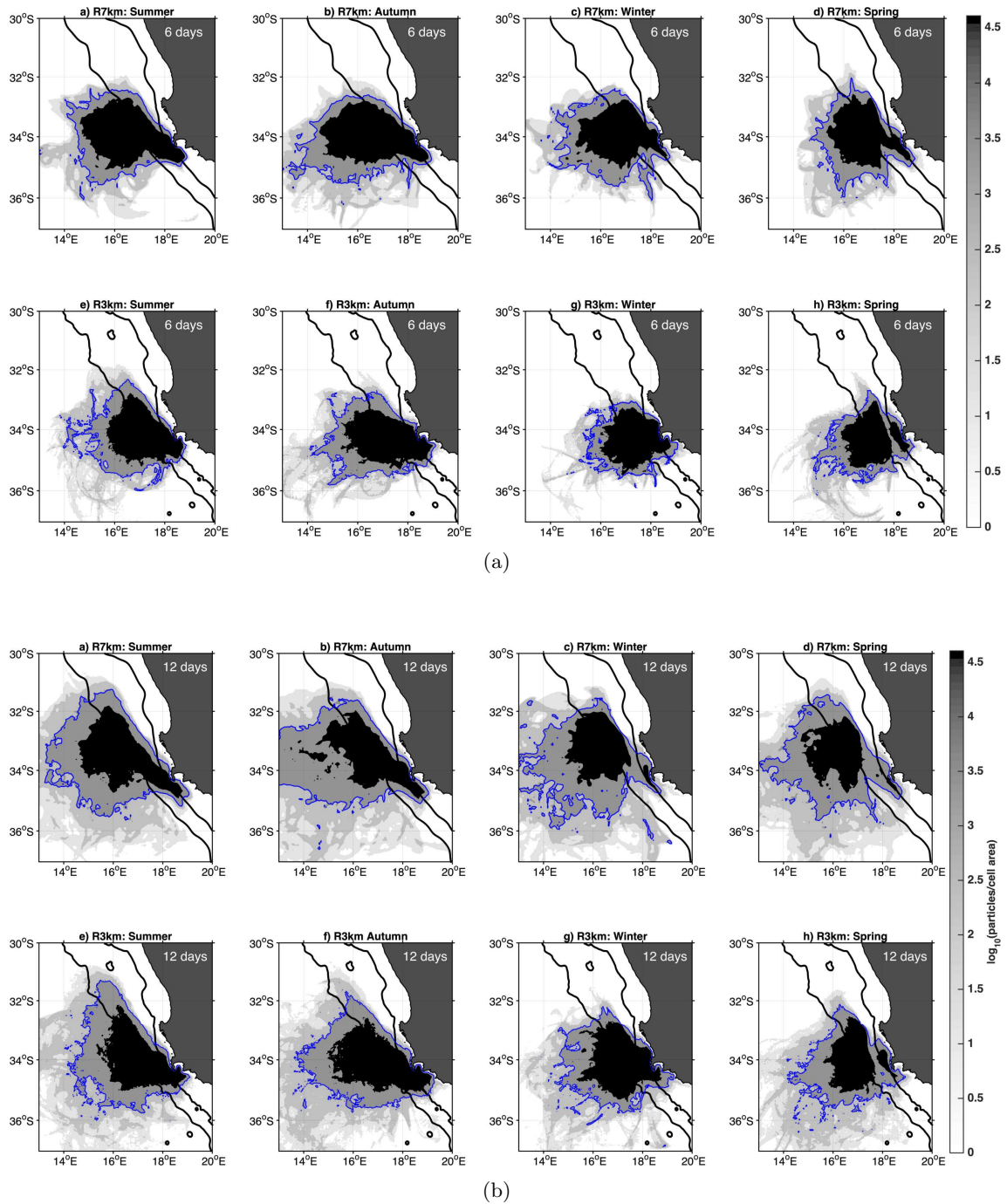


Figure 5.17: Seasonal plume dispersion pattern of unsuccessful particles at a) 6 days and b) 12 days after their release for 7 km (panel a-d) and 3 km (panel e-h). The density (colorbar) is obtained by dividing the sum of particles located in a defined grid box by their respective grid box area. The density of particles is shown in log scale to capture the large dispersion of particles in the higher resolution and a reference contour (blue line) of 3.5 m^{-2} is superimposed. The 200 m and 2000 m isobaths are also superimposed.

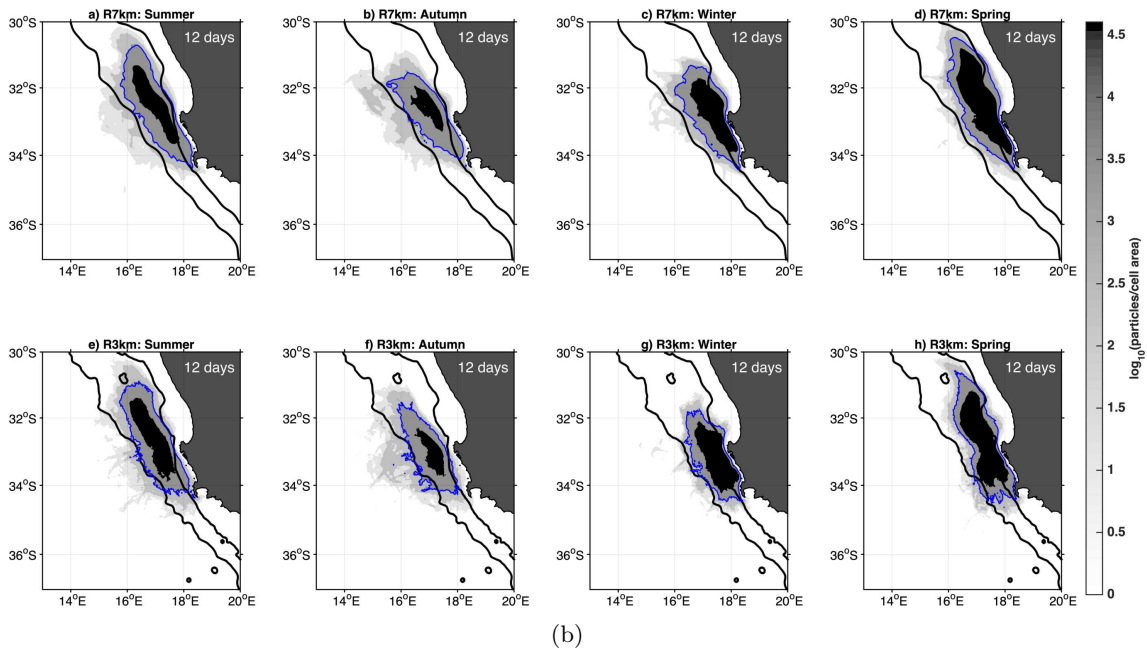
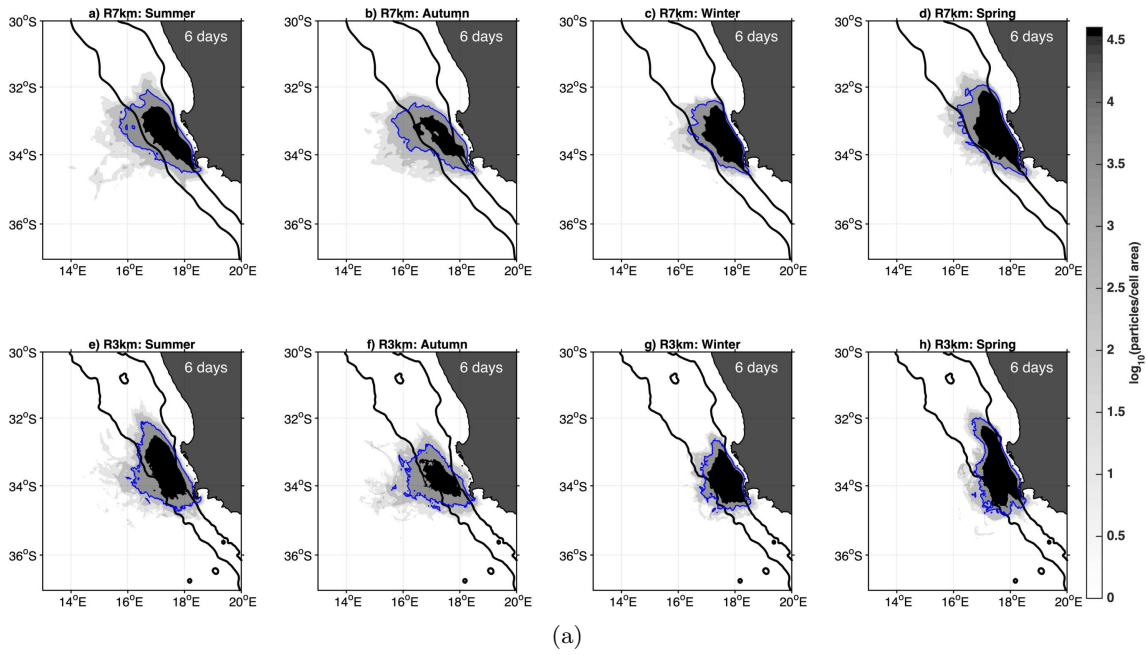


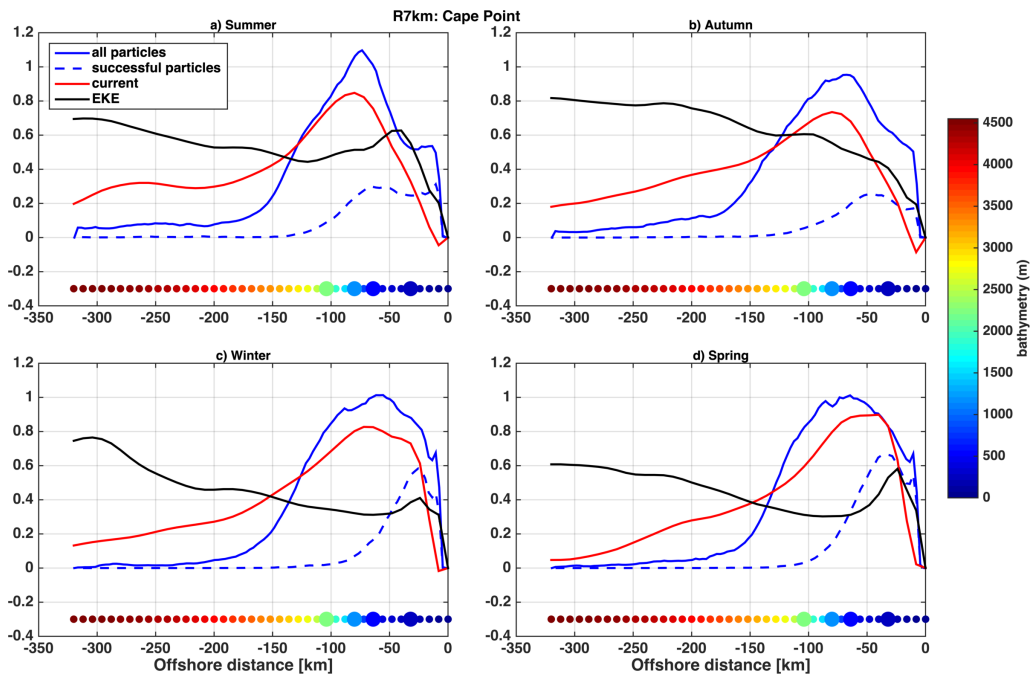
Figure 5.18: Same as in Figure 5.17 but for successful particles.

1779 5.6.1 Lagrangian pathways and Eulerian oceanic circulation

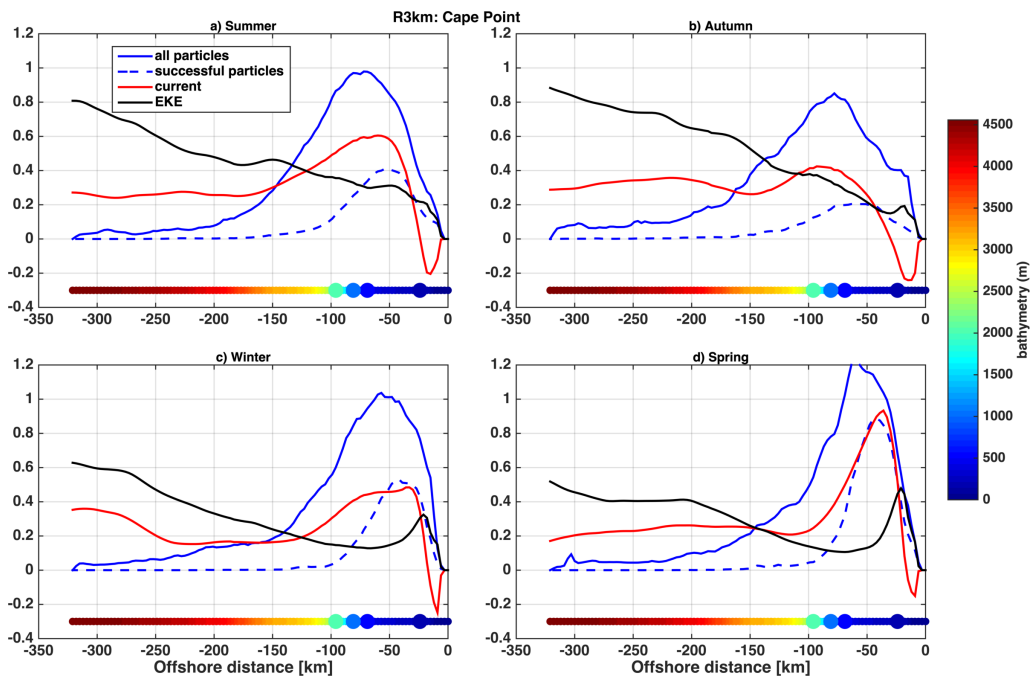
1780 The differences in inshore and offshore transport success in the two simulations is an
1781 indication of a difference in the ocean circulation influencing their transport. To relate
1782 the seasonality of the different Lagrangian transports to the contrasting ocean circulation
1783 features previously described, the cross-shore profiles of normalised: i) particles' distri-
1784 bution; ii) depth-integrated (top 100 m) alongshore velocity; iii) depth-integrated (top
1785 100 m) EKE at the Cape Point transect are presented in Figure 5.19. Generally, the peak
1786 in the cross-shore distribution of all particles coincides with the peak of the jet in both
1787 simulations (Fig. 5.19). The probability of all particles that travel within the 50 km off-
1788 shore distance is significantly smaller in the R3km simulation, and this due to the poleward
1789 current. On the other hand, successful particles tend to travel in the inshore side of the
1790 jet.

1791 In the two simulations, the differences in inshore transport success is related to both the
1792 intensity and structure of the Benguela Jet and the inner shelf poleward flow. In summer
1793 and spring, inshore particles in the R3km leave Cape Point in a jet that is closer to the
1794 coast and confined between the 200 m and 500 m isobaths thus resulting in high inshore
1795 transport success. In winter and autumn, there are less inshore particles in the R3km due
1796 to low inner shelf transport success attributable to the nearshore poleward flow at Cape
1797 Point driving a southward transport.

1798 Offshore particles were found to mainly be influenced by eddy activity at the shelf-edge.
1799 However, in the R7km, the proximity of high EKE signal to the shelf-edge exerts an en-
1800 hanced turbulence dispersive effect (Fig. 5.8a). Thus, there is less transport success for
1801 the offshore particles. Furthermore, more particles are exposed to this high EKE in sum-
1802 mer and autumn in the R7km (Fig. 5.19a,b) because of the north-westward transport of
1803 particles off Cape Point and hence a longer travel time during this period. Particles are
1804 also transported in the offshore direction in autumn in the R3km, however this had a
1805 small impact on transport success since the offshore EKE is relatively weaker than in the
1806 R7km run. Hence, the offshore particles in the R3km have higher probability of transport
1807 success during this period. The increase in transport success for the offshore particles in
1808 the R3km contributed significantly to the total transport success.



(a)



(b)

Figure 5.19: Seasonal cross-shore profile of normalised depth-integrated alongshore current (red), EKE (black), and probability of particles' occurrence (blue solid) at Cape Point for (a) R7km and (b) R3km. The histogram of successful particles is also superimposed (blue dashed). The enlarged circles mark the location of the 200 m, 500 m, 1000 m and 2000 m isobaths, respectively.

1809 Finally, the increase in the long travelling time observed in the R3km could be explained
1810 by the significant offshore transport success and the fact that particles travel in a weaker
1811 jet during summer, autumn and winter in particular.

1812 **5.7 Summary and Conclusion**

1813 This chapter investigated how increasing the model resolution from 7-km to 3-km impacts
1814 the the coastal circulation and therefore the alongshore connectivity between Cape Point
1815 and St Helena Bay. Despite some changes in the circulation, the seasonal cycle of the
1816 alongshore connectivity remains more or less the same. This is a highly unexpected re-
1817 sult considering that increasing the resolution has significantly affected the across-shelf
1818 structure of the alongshore jets observed over the mid-shelf as follows:

- 1819 • The Benguela Jet is narrower and less intense in the higher resolution simulation. It
1820 is also more confined towards the shelf edge and the surface.
- 1821 • A poleward surface and sub-surface flow is enhanced inshore of the Benguela Jet.

1822 Both of these impacts are expected to decrease the transport success of particles from
1823 Cape Point to St Helena Bay. In fact, the coastal poleward current contributes to the
1824 southward transport of inshore particles. Furthermore a weaker Benguela Jet results in a
1825 slower northward advection of particles. However, these negative effects are compensated
1826 by a much lower offshore loss of particles in the higher resolution simulation due to a
1827 more coherent Benguela Jet confined over the shelf-edge. The latter follows the 200-500 m
1828 isobaths, and its northwest-ward veering is reduced in the higher resolution simulation.

1829 These changes in the oceanic circulation may be linked to:

- 1830 • a better representation of the fine scale dynamics when the resolution is being in-
1831 creased. Some fine scale structures smoothed out by the numerical viscosity at a
1832 resolution of 7-km may be more persistent with a resolution of 3-km. These effects
1833 might lead to a better representation of the along-shore jets and their instabilities.
1834 They might also play a role on the growth of the very near-shore poleward surface
1835 and sub-surface flow.

- 1836 • a better representation of the coastal topography, in particular around capes and
1837 canyons. This would also lead to the enhancement of the across-shore shelf slope,
1838 which in turn could contribute to confine the along-shore jets over some specific
1839 isobaths.

- 1840 • a reduction of large-scale biases, here the leakage of warm Agulhas waters into the
1841 Cape Basin. This result was largely unexpected. A better representation of the Ag-
1842 ulhas Ridge and the isolated seamounts disseminated in between the eastern flank of
1843 this ridge and the continent seems to have blocked partly the leakage of warm Agul-
1844 has waters into the Cape basin, and pushed the route of turbulent eddies trapping
1845 these waters more offshore. As a result: i) the warm bias observed in the reference
1846 7-km simulation is significantly reduced in the 3-km simulation, which contributes to
1847 change the characteristics of the Benguela Jet; ii) the mesoscale eddy turbulence in
1848 the offshore coastal transition zone is reduced in the 3-km simulation, which reduces
1849 the offshore loss of particles.

1850 However, the main finding of this chapter is that the Benguela Jet is the main driver
1851 of the seasonal alongshore connectivity in the SBUS. Since the Benguela Jet is present
1852 in both simulations, there are little differences in the simulated transport success at the
1853 seasonal time-scale. In fact, the seasonal cycle of the Benguela Jet does not seem to be
1854 affected by the influx of Agulhas waters into the Cape basin that differed between the
1855 two simulations. To confirm this result, two nested 1 km simulations have been run over
1856 the 2005-2011 period. One was nested within the 3-km [R1km.EXP1] (Fig. 5.20a) and
1857 therefore had a reduced Agulhas water influx in comparison to the R7km simulation. The
1858 other was nested within the 7-km simulation [R1km.EXP2] (Fig. 5.20b) and therefore had
1859 an enhanced Agulhas waters influx in comparison to the R3km simulation. As expected,
1860 preliminary results show that increasing the resolution to 1 km leads to a stronger and
1861 more coherent surface and sub-surface nearshore poleward flow. Finer scale eddies can
1862 also be observed in the 1 km resolution simulations. Nevertheless, despite some small
1863 changes in its across-shore position, width and vertical extension, the seasonal cycle of
1864 the Benguela Jet is very similar in the two simulations, and coherent with our findings
1865 from previous lower 3-km and 7-km simulations. Therefore, the characteristics of the

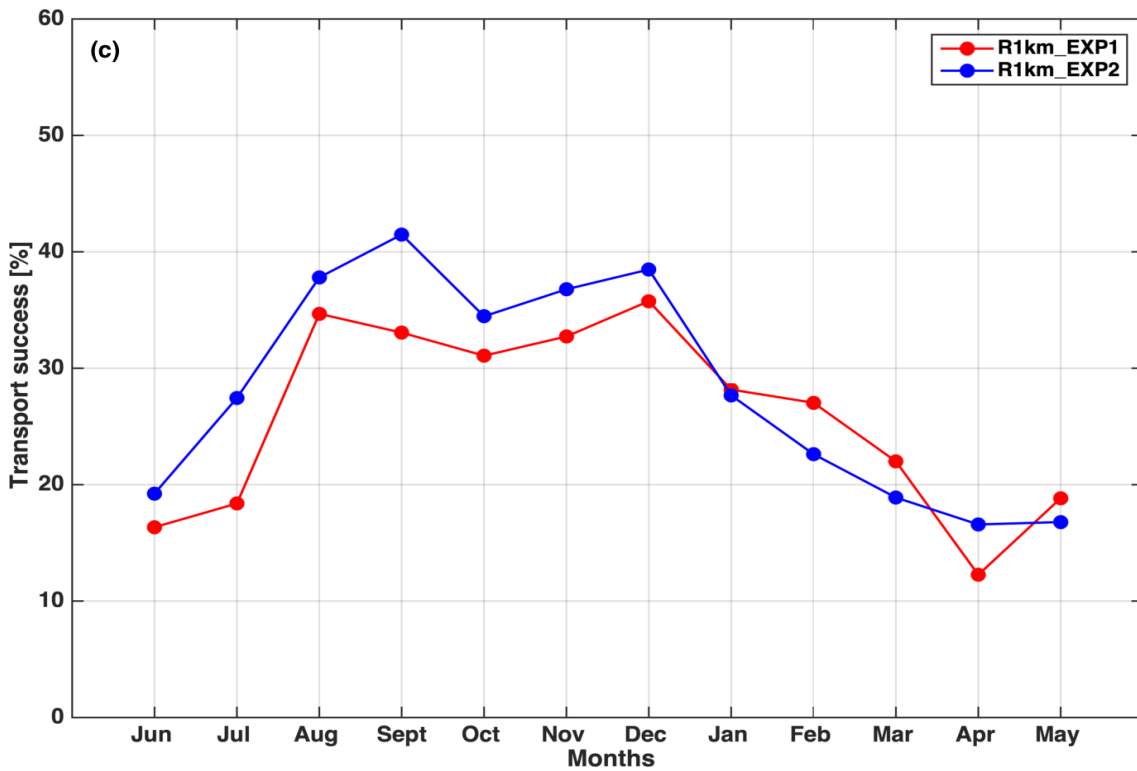
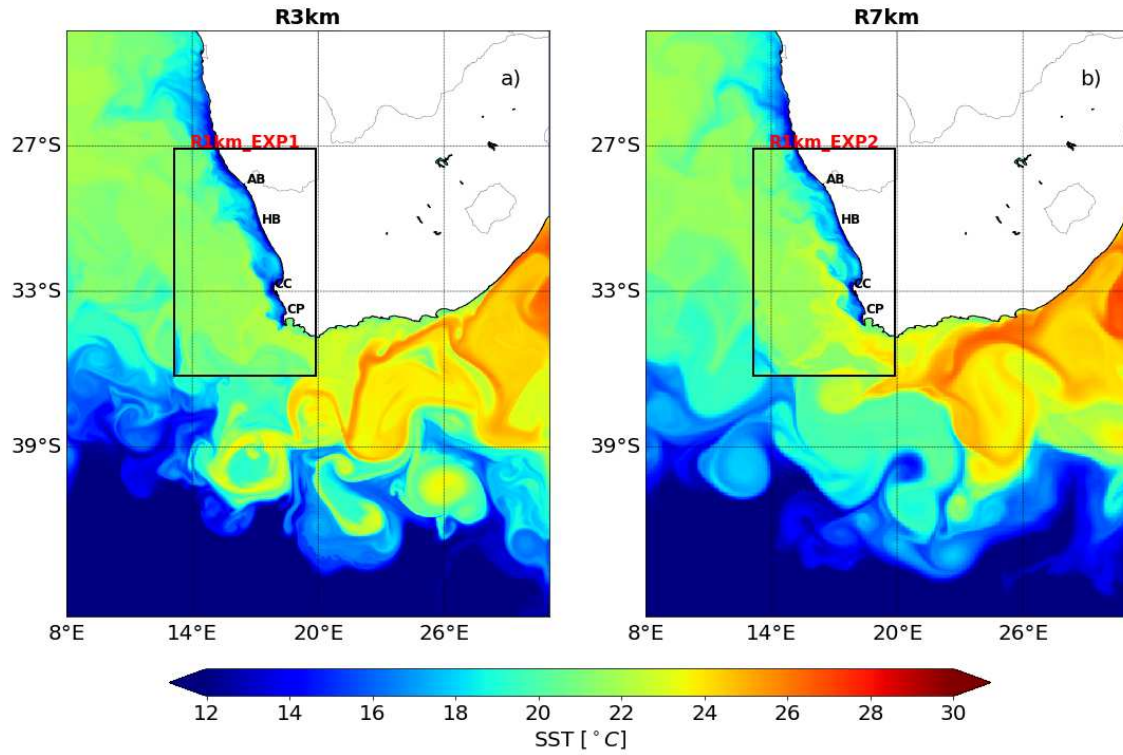


Figure 5.20: Snapshot of SST (Year 2005, 1st of January) for the (a) 7-km and (b) 3-km simulation. The black box delineates the 1-km simulation boundaries and the SST of the 1-km is superimposed within the box. (c) Monthly climatological transport success of R1km_EXP1 (red line) and R1km_EXP2 (blue line) over the 2005-2011 period.

1866 along-shore connectivity and the seasonal cycle of the northward transport success is very
1867 similar (Fig. 5.20c). Nevertheless, differences between the two simulations are expected at
1868 intraseasonal and interannual time-scales.

Chapter 6

Interannual variability of transport success

6.1 Introduction

In Eastern Boundary Upwelling Systems (EBUS), the variability in small pelagic fish populations abundance depends on fish recruitment success which is influenced by a set of both environmental and biological factors (*Bakun, 1996; Painting et al., 1998*). Small pelagic fish spawn in a stable (non-turbulent) environment, and their eggs and larvae must be transported almost as passive tracers towards a region of food availability (*Parrish et al., 1983*). In the southern Benguela upwelling system (SBUS), sardines and anchovies spawn on the Agulhas Bank (Fig. 6.1a) located off the south-west and south coast of South Africa from October to March (*Painting et al., 1998; van der Lingen and Huggett, 2003*). The northward advection of the eggs and larvae towards the food-rich nursery area off the west coast in St Helena Bay (located just to the northeast of Cape Columbine in Fig. 6.1a) is facilitated by a shelf-edge jet named the Benguela jet (*Shelton and Hutchings, 1982; Hutchings et al., 1998*). Those larvae that get successfully transported into the nursery area are considered as recruits.

Abundant catches of anchovy begun in the 1960s after the collapse of sardine fishery (*Hutchings et al., 1998*), and continue to drive an essential socio-economic activity. Small

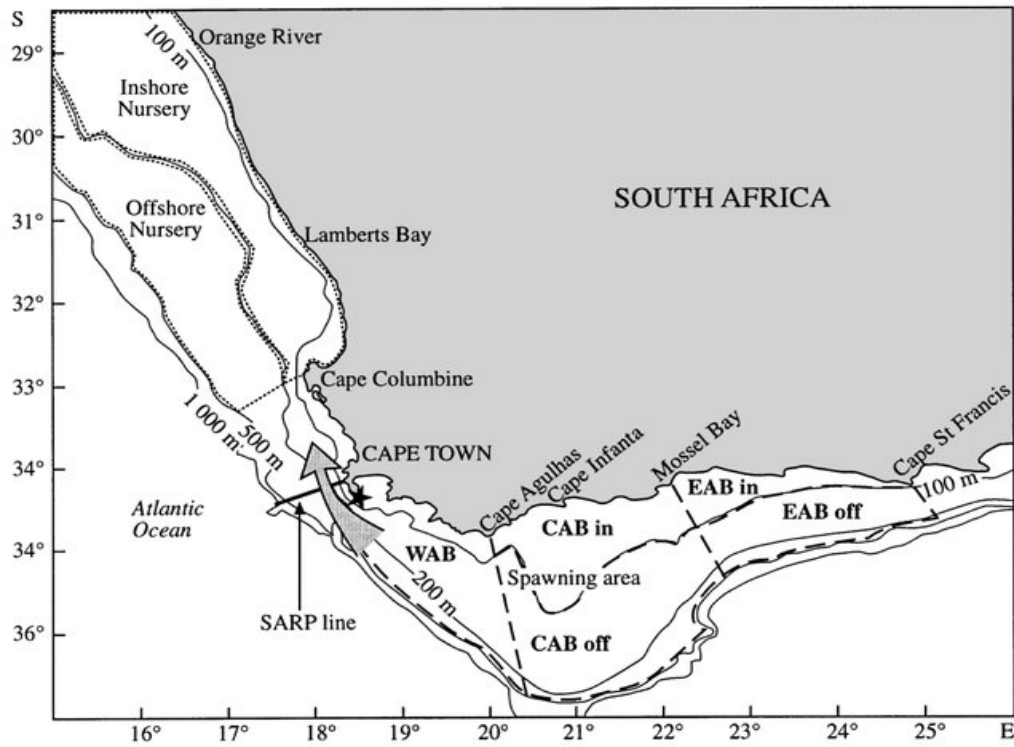
1889 pelagic fish account for 20-25% of total South African fisheries catch value and contribute
1890 less than 1% of the country's total gross domestic product (GDP), but create substan-
1891 tial employment locally (*Brick and Hasson, 2017*). Analysis of repeated hydro-acoustic
1892 surveys of anchovy biomass done since 1984 over the entire area occupied by the an-
1893 chovy adult population (Fig. 6.1b) shows a large interannual variation of recruitment and
1894 spawner stock size (*Huggett et al., 2003*). This variability presents a challenge to fisheries
1895 management and has led to a large number of studies investigating how this variability
1896 is linked to changes in the environmental conditions (*Boyd et al., 1998; Hutchings et al.,*
1897 *1998; Huggett et al., 2003; Mullon et al., 2003; Parada et al., 2003; Lett et al., 2006; Parada*
1898 *et al., 2008; Koné et al., 2013*).

1899

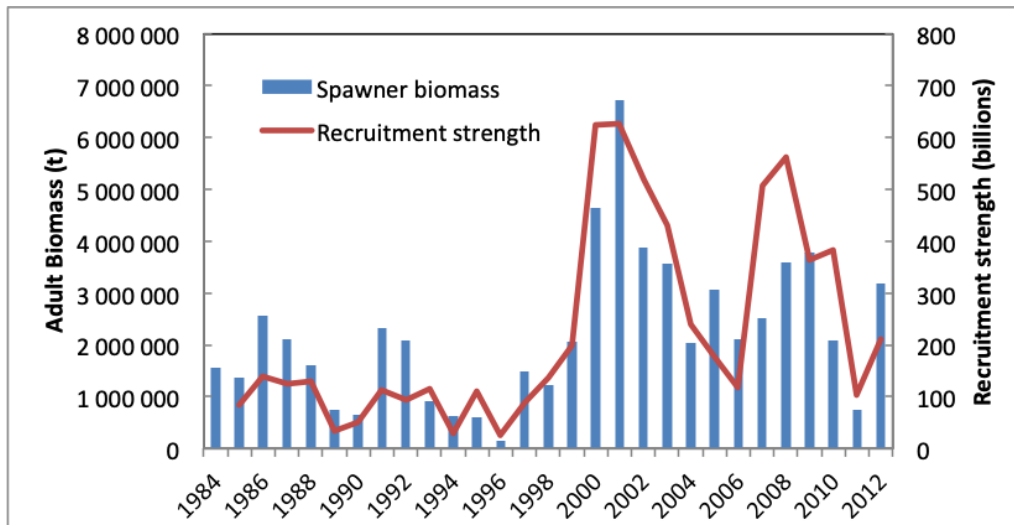
1900 There are multiple variables linked to both the environment (e.g. wind-induced turbu-
1901 lence, temperature, food availability) and biological factors (e.g. egg abundance, spawning
1902 conditions, mortality due to predation, cannibalism and starvation) that can impact re-
1903 cruitment success (*Painting et al., 1998; Shannon et al., 1996; Hutchings et al., 1998*).
1904 This complexity makes it difficult to evaluate the relative impact of these variables (which
1905 may act on different temporal and spatial scales) on recruitment, as they can act simulta-
1906 neously or independently from one year to another (*Mullon et al., 2002*). However, there
1907 has been an attempt in previous studies to understand the relationships between environ-
1908 mental fluctuations such as the wind and transport, enrichment processes and year-class
1909 recruitment strength, but with inconsistent findings (*Bloomer et al., 1994; Boyd et al.,*
1910 *1998; Shannon et al., 1996; van der Sleen et al., 2018*).

1911

1912 Empirical studies have shown a negative relationship between strong south-easterly
1913 winds and recruitment during the transport phase due to increased offshore loss of eggs
1914 and larvae and anchovy (*Boyd et al., 1998*). On the other hand, *van der Sleen et al. (2018)*
1915 found a positive relationship between recruitment and a cumulative upwelling index which
1916 improves when the effect of upwelling is integrated over multiple spawning seasons. They
1917 suggest that high offshore loss that occurs in the summer of a particular year (year_t) could
1918 lead to low recruits and thus low spawners for the following year (year_{t+1}), whereas low



(a)



(b)

Figure 6.1: (a): Map showing anchovy spawning area on the Agulhas Bank (dashed lines) that extends from 18°E to 26°E and the nursery area located off the west coast between Cape Columbine and the Orange River on the west coast (dotted lines). The arrow indicates the approximate location of the shelf-edge jet current (Benguela Jet), the star indicates the location of Cape Point. The SARP monitoring line is represented by the cross-shore transect (black solid line) off Cape Point. The map has been adapted from *Huggett et al. (2003)*. (b): SBUS recruitment strength of anchovies in billions (1985-2012) (black line) and adults biomass (bar graph) in tons (1984-2012). Updated from *van der Lingen and Huggett (2003)*.

1919 offshore loss could be beneficial for population growth during a period of low spawning
1920 activity.

1921

1922 Moreover, the early 2000s was characterized by a large increase in the mean level of
1923 both spawner biomass and recruitment (Fig. 6.1b) (*van der Lingen and Huggett, 2003; Roy*
1924 *et al., 2007*). Figure 6.1b shows that the 1990s were characterised by moderate recruitment
1925 levels and that the 2000-2010 period experienced two periods of exceptionally high recruit-
1926 ment levels separated by low values during in the mid 2000's. Long-term observational
1927 studies of the Benguela upwelling system (BUS) suggest that these changes might be re-
1928 lated to upwelling variability (*Blamey et al., 2012; Lamont et al., 2018*). Previous studies
1929 (*van der Lingen et al., 2002; van der Lingen and Huggett, 2003; Roy et al., 2007*) indicate
1930 that in the early 1990s there was an eastward shift of the spawning region, attributed to
1931 coastal cooling east of Cape Agulhas Bank (Fig. 6.1a). This result is consistent with the
1932 findings of *Blamey et al. (2012)* based on in-situ wind measurements (1960-2010) and *La-*
1933 *mont et al. (2018)* who used reanalysis wind data (1979-2015). From about 1995 to 1999,
1934 the SBUS experienced a dominance of south-easterlies and strong upwelling together with
1935 WAB cooling which led to relatively low anchovy recruitment during this period (*Howard*
1936 *et al., 2007; Blamey et al., 2012*).

1937

1938 Variability in the passive transport of eggs and larvae from the spawning ground to
1939 the nursery on the west coast is also considered one of the critical factors influencing re-
1940 cruitment variability (*Hutchings, 1992; Hutchings et al., 1998; Shannon et al., 1996*). As
1941 a result, several numerical Lagrangian transport studies have focused on oceanic currents
1942 forced by monthly seasonal climatology of wind-stress, freshwater and heat fluxes (*Huggett*
1943 *et al., 2003; Mullon et al., 2003; Parada et al., 2003, 2008*) thus excluding any externally
1944 forced interannual variability. The importance of using a realistic, time-varying surface
1945 forcing was highlighted by *Blanke et al. (2005)* who found that a daily wind product is
1946 more efficient in simulating the observed nearshore interannual variability of SST anoma-
1947 lies (warm and cold events) than a weekly product (*Blanke et al., 2002*).

1948

1949 Oceanic mesoscale eddies can also play a significant role in transporting eggs and larvae
1950 by either retaining or dispersing them (*Hutchings et al.*, 1998). Moreover, eddies and their
1951 associated recirculation patterns can also play an essential role in advecting larvae back
1952 inshore into a retention area (*Penven et al.*, 2000). The impact of eddies on the alongshore
1953 connectivity between the Agulhas Bank and the SBUS shelf was demonstrated by *Blanke*
1954 *et al.* (2009). They found that if a cyclonic eddy (anti-cyclonic) eddy is generated on
1955 the western edge of the Agulhas Bank, the northward transport is reduced (enhanced).
1956 Nevertheless, the effect of the shelf-edge eddies in our domain between Cape Point and St
1957 Helena Bay has not been investigated.

1958

1959 This chapter aims to investigate the interannual variability of transport success be-
1960 tween Cape Point and St Helena Bay and evaluate how it is related to variability of (i)
1961 wind-driven coastal circulation (Ekman transport, Benguela Jet, poleward current) and,
1962 (ii) the presence of mesoscale eddy activity at the coastal transition zone defined as the
1963 area from the continental slope to offshore. Analysis of transport success is based on La-
1964 grangian particles trajectories advected by ocean currents provided by the 3-km resolution
1965 (R3km) interannual simulation described in **Chapter 5**, for the 1992-2011 period. As it
1966 has been discussed in **Chapter 5**, the R3km simulation is skillful at reproducing realistic
1967 Agulhas leakage (section 4.6) and captures well the shelf circulation of the Benguela Jet
1968 and poleward current (section 4.4). Given that this simulation is forced with realistic
1969 rather than climatological surface forcing, it allows us to investigate the role of interan-
1970 nual variability and locally forced episodic events, rather than just seasonal and intrinsic
1971 variability.

1972

1973 The rest of this chapter is structured as follows: section 6.2 presents the interannual
1974 transport success anomalies. Section 6.3 explores the relationship between these transport
1975 success anomalies and the anomalies of local dynamics such as the Benguela Jet, upwelling
1976 (wind field) and the oceanic turbulence (EKE). The contribution of the particles' initial
1977 cross-shore position to transport success anomalies and their relation to the oceanic cir-
1978 culation is presented in section 6.4. Analyses of the cross-shore transport variability of

1979 particles and the description of the role played by mesoscale eddies is found in section 6.6.
1980 Finally, the main findings of this work are summarised in section 6.7.

1981 **6.2 Transport success variability**

1982 The interannual variability of transport success is investigated from 1992 to 2011 using
1983 Lagrangian experiments in which particles were randomly released every 12 hours at a
1984 cross-shore section off Cape Point (34°S) (Fig. 6.2a) over the upper 100 m of the water
1985 column (Fig. 6.2b) and followed for a duration of 30 days. The particles were released
1986 over 108 defined vertical profiles, 3 km apart, with a density of 1.125 particles per meter
1987 which resulted in a total of 4840 particles for each release event. The particles were then
1988 advected by daily 3-D velocity fields (u, v, w) provided by the R3km model simulation
1989 described in chapter 5 (section 5.2.1).

1990 For each release event e , transport success TS_e (Eq. 6.1) is given by the ratio of the number
1991 of particles reaching St Helena Bay (n_e) over the total number of particles released (4840).

$$TS_e = \frac{n_e}{4840}. \quad (6.1)$$

1992 The monthly climatology (\overline{TS}_m) of transport success and its standard deviation σTS_m
1993 (Eq. 6.2a) shown in Fig. 6.3 exhibit a strong seasonal variability. Transport success is
1994 higher during late austral winter to austral spring (August-November) when particles have
1995 >30% probability of reaching St Helena Bay. The lowest transport success occurs during
1996 austral autumn (March-May) reaching as low as 13% in May and the standard deviation
1997 is also lowest at this time of the year. The highest standard deviation of transport success
1998 occurs during the winter season (June-August).

1999 To analyse the interannual variability, anomalies TS'_m were defined by subtracting monthly
2000 climatology \overline{TS}_m (shown in Fig. 6.3) from monthly mean patterns of transport success from
2001 1992 to 2011 (240 months) and normalised by monthly climatological standard deviation
2002 σTS_m (shown in Fig. 6.3) as follows:

$$\sigma TS_{m\{m=1,\dots,12\}} = \sqrt{\sum_{e(m)} (TS_{e(m)} - TS_m)^2} \quad (6.2a)$$

$$TS'_m = \frac{TS_m - \overline{TS_m}}{\sigma TS_m} \quad (6.2b)$$

2003 with $e(m)$, the event index chosen to correspond to all the release events that occur during
 2004 month m over the 1992-2011 period.

2005 The time-series of normalised transport success anomalies is shown in Fig. 6.4 which
 2006 reveals strong interannual variability. There are notable periods of positive and negative
 2007 transport success. Positive anomalies of transport occurred from 1992 to end of mid-1993.
 2008 This period was followed by one of positive transport success from 1998 to mid-2001. The
 2009 period between 2002 and 2004 is more variable, being characterised by as many positive as
 2010 negative events that appear to be random. There was another extended period of negative
 2011 transport success from 2008 to 2011.

2012 Over the entire period, the highest positive transport success occurred in April 1992 with
 2013 a standard anomaly of 3.1 whereas the lowest negative transport success happened during
 2014 September 2002 with a standard anomaly of -2 (Fig. 6.4).

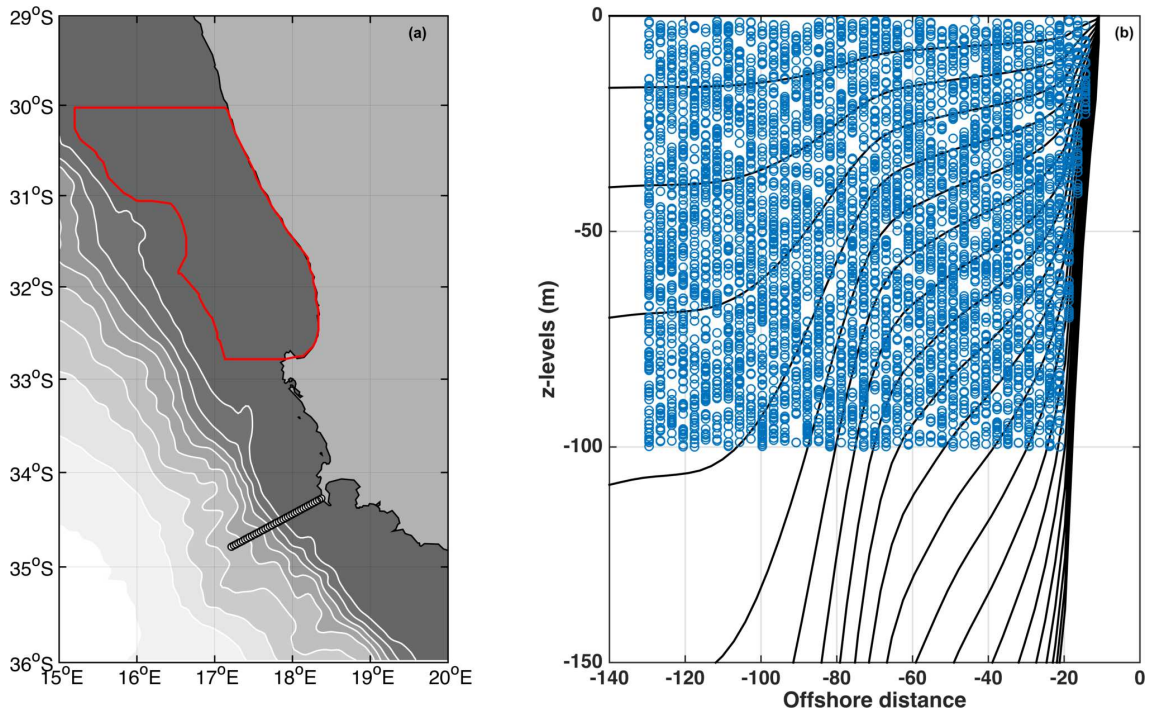


Figure 6.2: (a): Map of the R3km model simulation bathymetry (in grey with a contour interval of 500 m). Particles are released between the surface and -100 m along the across-shore transect off Cape Point (white circles). The target area is defined offshore St Helena Bay and is delineated by the red polygon. A particle is considered successful if it reaches the target area within 30 days (b): One random initial vertical distribution of particles off Cape Point during a release event (blue circles). Terrain following Sigma levels of the R3km model are superimposed in black.

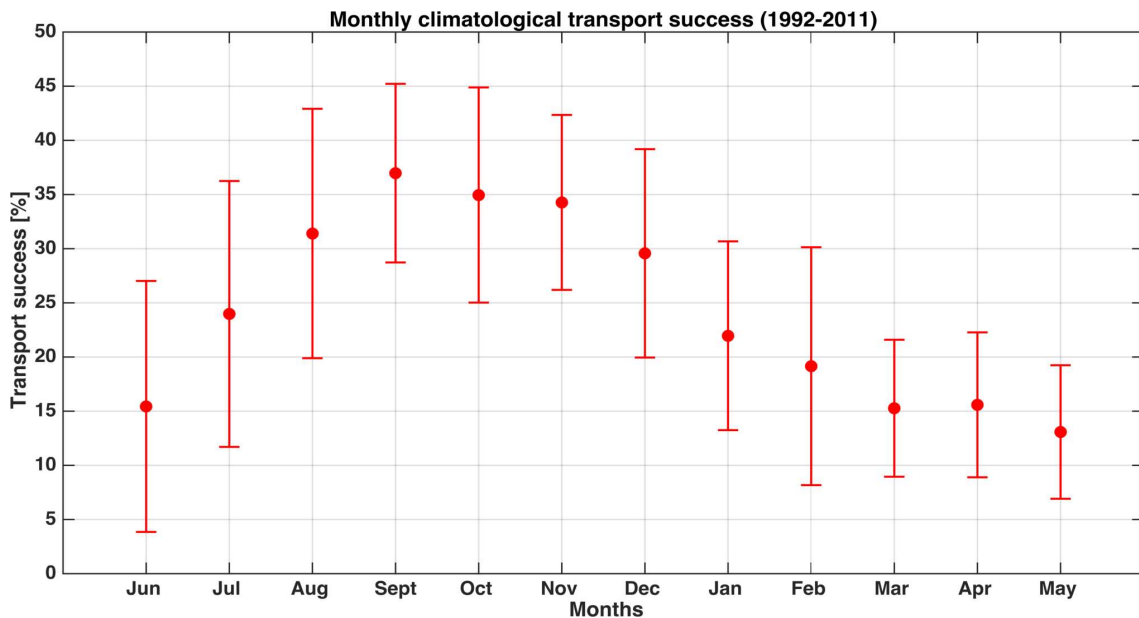


Figure 6.3: Monthly climatology of transport success \overline{TS}_m (red dots) and their standard deviation σTS_m (error-bars)

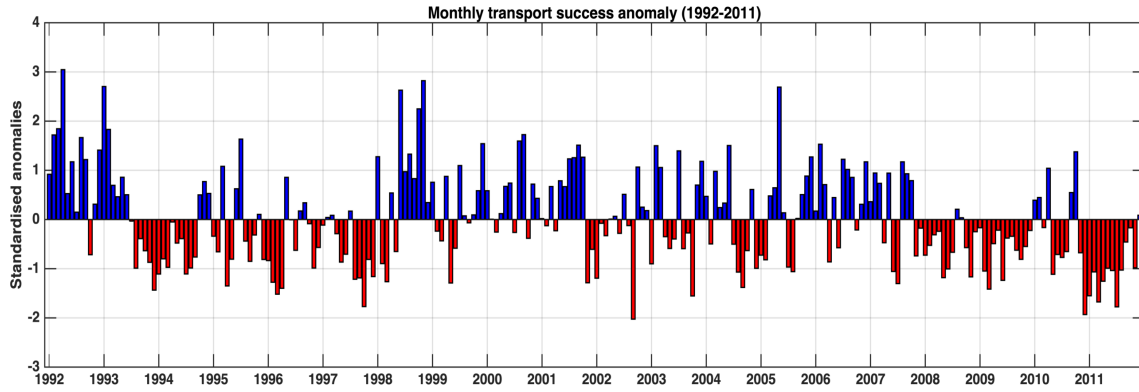


Figure 6.4: Normalised monthly transport success anomalies (TS') over the 1992-2011 period.

2015 6.3 The role of local dynamics

2016 To understand the relationship between the anomalies of transport success and the local
 2017 dynamics, anomalies of the Benguela jet, alongshore wind stress, and eddy kinetic energy
 2018 (EKE) are computed. All anomalies were computed by the removing the mean monthly
 2019 climatology from the monthly mean values and then normalised by the standard deviation
 2020 of the monthly mean values.

2021
 2022 The annual mean surface alongshore current velocities (Fig. 6.5a) show that the Benguela
 2023 jet is found between the 200 m and 500 m isobaths extending from Cape Point (34°S) to
 2024 offshore St Helena Bay (32°S) consistent with *Veitch et al. (2017)*. To investigate the
 2025 relationship between the Benguela jet variability and transport success variability, the
 2026 surface alongshore current velocity anomalies are averaged over the 200-500 m isobaths
 2027 from Cape Point to north of Cape Columbine at 32°S to form an index of the intensity
 2028 of the Benguela jet. The scatterplot in Fig. 6.5b shows normalised monthly anomalies
 2029 in transport success and the Benguela jet index. A significant positive correlation ($r =$
 2030 0.49 , $p < 0.05$) is found indicating that positive (negative) transport success anomalies
 2031 are attributed to a stronger (weaker) Benguela jet intensity. Despite the relatively strong
 2032 correlation, there are still some substantial positive (negative) events associated with neg-
 2033 ative (positive) Benguela jet anomalies.

2034 Equatorward alongshore winds (Fig. 6.6a) which are upwelling-favourable drive an offshore

2035 Ekman transport and could contribute to the cross-shelf transport of surface particles.
2036 Therefore, the interannual variability of upwelling-favourable winds strength could have
2037 an impact on the retention or offshore transport of particles. A weak, but still a signifi-
2038 cant negative relationship ($r = -0.22$, $p < 0.05$) is found between monthly alongshore wind
2039 stress anomalies and transport success anomalies (Fig. 6.6b). This correlation implies
2040 that strong equatorward wind stress (positive anomalies) and associated offshore Ekman
2041 transport tend to not be favourable for transport success. The low correlation could be
2042 because the correlations are done on monthly mean winds which filters out fluctuations
2043 of the upwelling favourable winds that induces alternating periods of intense and reduced
2044 upwelling (reduces offshore loss).

2045

2046 Another phenomenon in the SBUS is the presence of intense mesoscale eddy activity
2047 just off the shelf edge (*Veitch et al.*, 2009; *Veitch and Penven*, 2017). This eddy activity
2048 may affect particles that originate offshore as well as inshore particles that get advected
2049 into this domain of high mesoscale activity (Fig. 6.7a). To examine the relationships with
2050 transport success anomalies, anomalies of EKE averaged between Cape Point (34°S) and
2051 Cape Columbine (33°S) from the shelf edge (500 m isobath) to the 3000 m isobath are
2052 computed (Fig. 6.7a). The anomalies of transport success and EKE are shown in Fig. 6.7b
2053 and are significant negatively correlated ($r = -0.35$, $p < 0.05$). This correlation suggests
2054 that anomalously positive EKE (high mesoscale turbulence) drives a dispersion of particles
2055 and thus contributes to negative transport success. However, it is unclear as to whether
2056 the high EKE is associated with cyclonic or anti-cyclonic eddies which are both prevalent
2057 in this region.

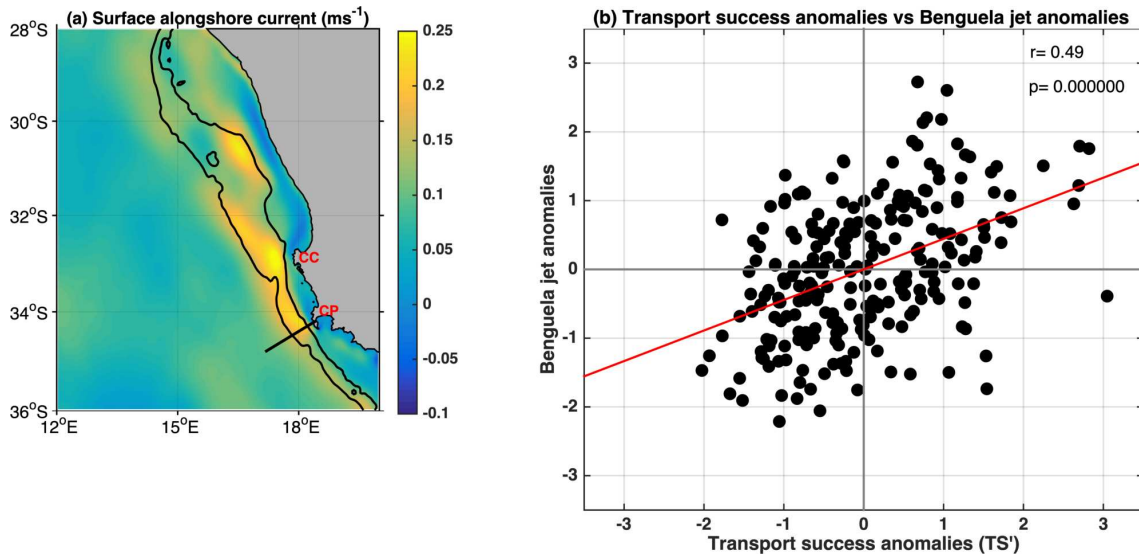


Figure 6.5: (a) Annual mean sea surface alongshore current velocities (m s^{-1}) of the R3km simulation. 200 m and 500 m isobaths are shown in black. (b) Scatterplots of normalised transport success anomalies against the Benguela jet index anomalies averaged between Cape Point (34°S) and St Helena Bay (32°S) over the 200-500 m isobaths. A positive (negative) Benguela jet anomaly suggests a stronger (weaker) than average northward jet

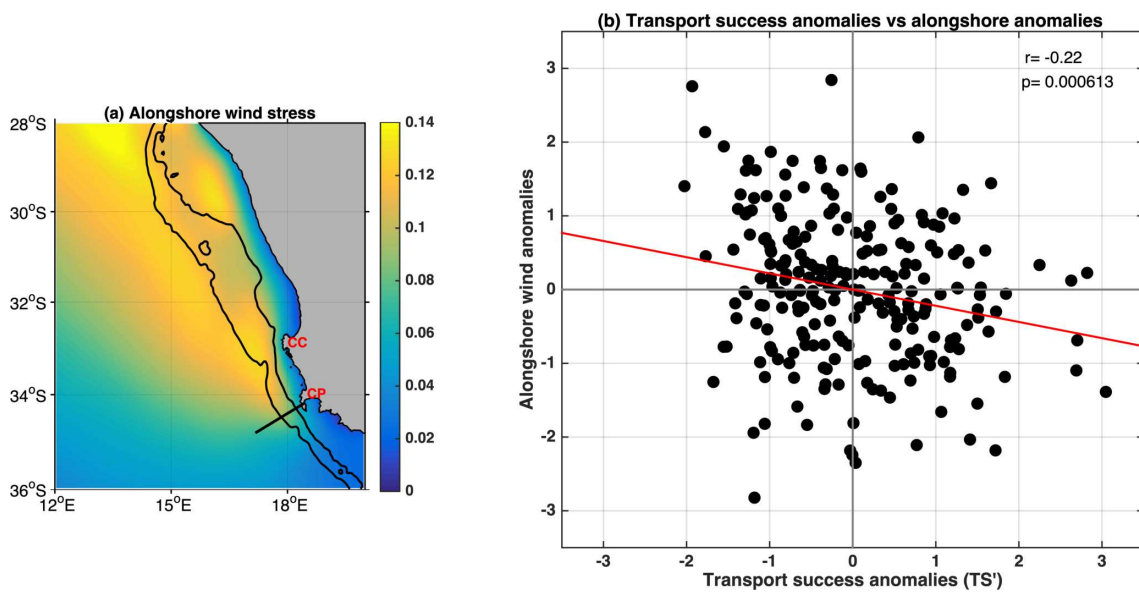


Figure 6.6: (a) January alongshore wind stress (m s^{-1}) climatological mean of the R3km simulation. 200 m and 500 m isobaths are shown in black. Same as in Fig. 6.5b but for transport success anomalies against alongshore wind stress anomalies averaged over at the transect off Cape Point (black line on the map (a)). A positive (negative) alongshore wind anomaly suggests a stronger (weaker) equatorward winds.

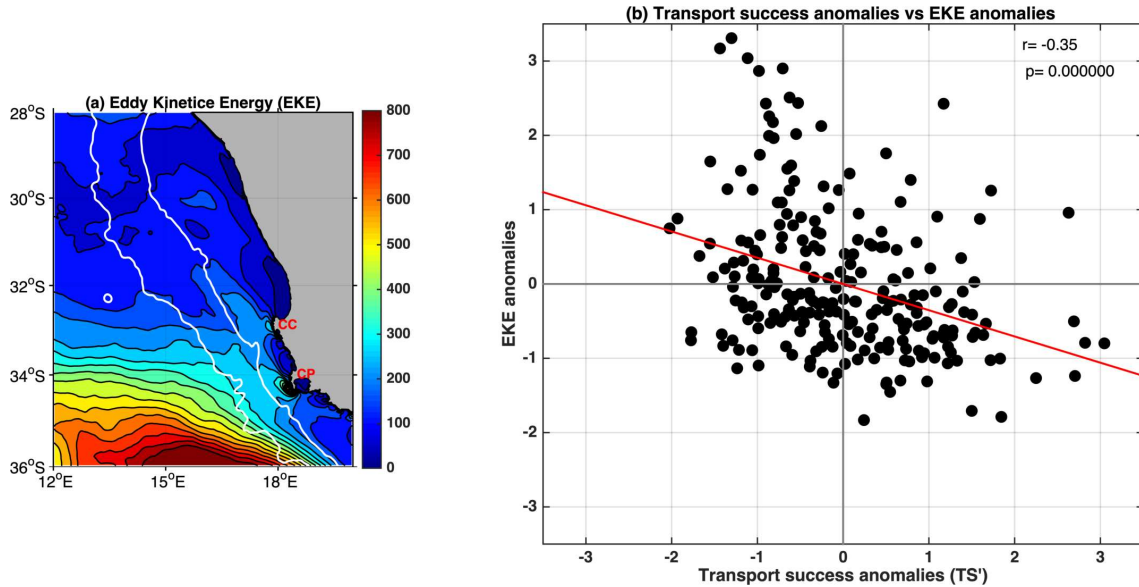


Figure 6.7: Annul mean surface EKE ($\text{cm}^{-2}\text{s}^{-2}$) of the R3km simulation. 500 m and 3000 m isobaths are superimposed in white. (b) Same as in Fig. 6.5b and Fig. 6.6b, but for normalised transport success anomalies against EKE anomalies. The EKE anomalies are averaged between Cape Point (34°S) and Cape Columbine (33°S), from the shelf edge (500 m isobath) to the 3000 m isobath. A positive (negative) EKE anomaly suggests a stronger (weaker) mesoscale turbulence.

2058 6.4 Inshore and offshore transport success

2059 The normalised transport success anomalies (TS') shown in Fig. 6.4 were further separated into inshore (particles released inshore of the 500 m isobath) and offshore (particles released offshore of the 500 m isobath) particles to quantify their relative contribution of their successful transport anomalies to the total transport success anomaly. The monthly climatological means of inshore and offshore transport success are denoted as \overline{TSI} and \overline{TSO} with their standard deviations as σTSI and σTSO , respectively. Note that the seasonal cycle of inshore and offshore transport success is described in the previous chapter (Fig. 5.16) under section 5.5.

2067 Monthly standardised anomalies are calculated as:

$$TSI'_m = \frac{TSI_m - \overline{TSI}_m}{\sigma TSI_m} \quad (6.3a)$$

$$TSO'_m = \frac{TSO_m - \overline{TSO}_m}{\sigma TSO_m} \quad (6.3b)$$

2068 Both inshore TSI' and offshore TSO' are strongly correlated with the total transport
 2069 mean TS' (respective $r = 0.83, 0.84$ $p < 0.05$). Figure 6.8 indicates that although total
 2070 negative transport is generally related to both negative inshore and offshore transport
 2071 anomalies, notably the negative anomalies associated with inshore particles (blue bars)
 2072 have a more significant role to play when there is negative total transport success. In
 2073 contrast, it is the offshore particles (orange bars) that contribute more to transport dur-
 2074 ing positive total transport success particularly for the largest values of positive transport
 2075 success that are stronger than those inshore (Fig. 6.8). However, there is a weaker rela-
 2076 tionship between inshore TSI' and offshore TSO' transport success anomalies ($r = 0.4, p$
 2077 < 0.05) which suggests that their variability might have different or independent physical
 drivers.

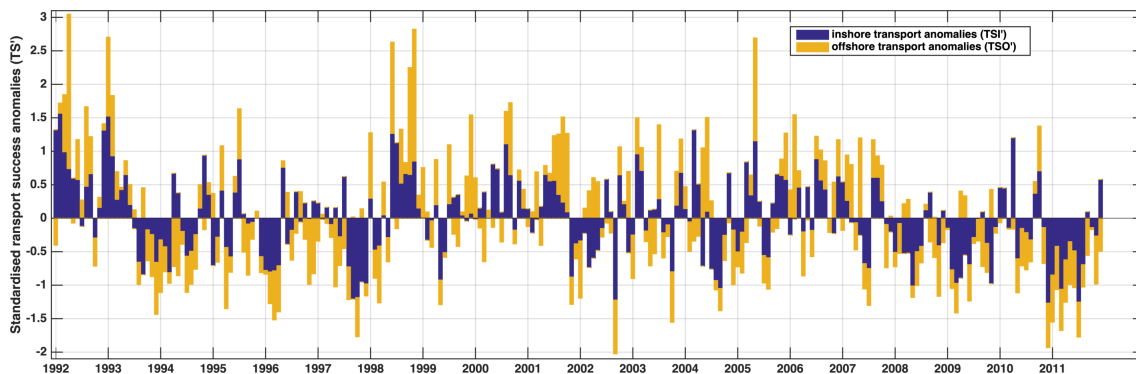


Figure 6.8: Same as in Fig. 6.4 but it is plotted in cumulated bars of inshore (TSI' : blue) and offshore (TSO' : orange) transport anomalies such that $TSI' + TSO' = TS'$

2078

2079 Similarly, the inshore and offshore transport success anomalies are related to the vari-
 2080 ability in the local dynamics. Fig. 6.9a suggests that inshore transport is positive (neg-
 2081 ative) when the Benguela Jet is stronger (weaker) than average, and indeed they are
 2082 strongly correlated ($r = 0.6, p < 0.05$). Weaker upwelling favourable wind stress (offshore
 2083 Ekman transport) tends to favour positive TSI' ($r = -0.21, p < 0.05$) (Fig. 6.9b).
 2084 In addition, a coastal poleward flow driven by negative wind stress curl is likely to drive a
 2085 southward transport of particles located inshore as previously discussed in Chapter 5. The
 2086 poleward current is derived by averaging only negative velocities ($v < 0$) over the 50 km
 2087 coastal band and integrated over the top 100 m depth at a transect off Cape Point. A pos-
 2088 itive (negative) anomaly of this flow indicate a weaker (stronger) than average poleward
 2089 current. There is a significant positive correlation ($r = 0.5, p < 0.05$) between inshore

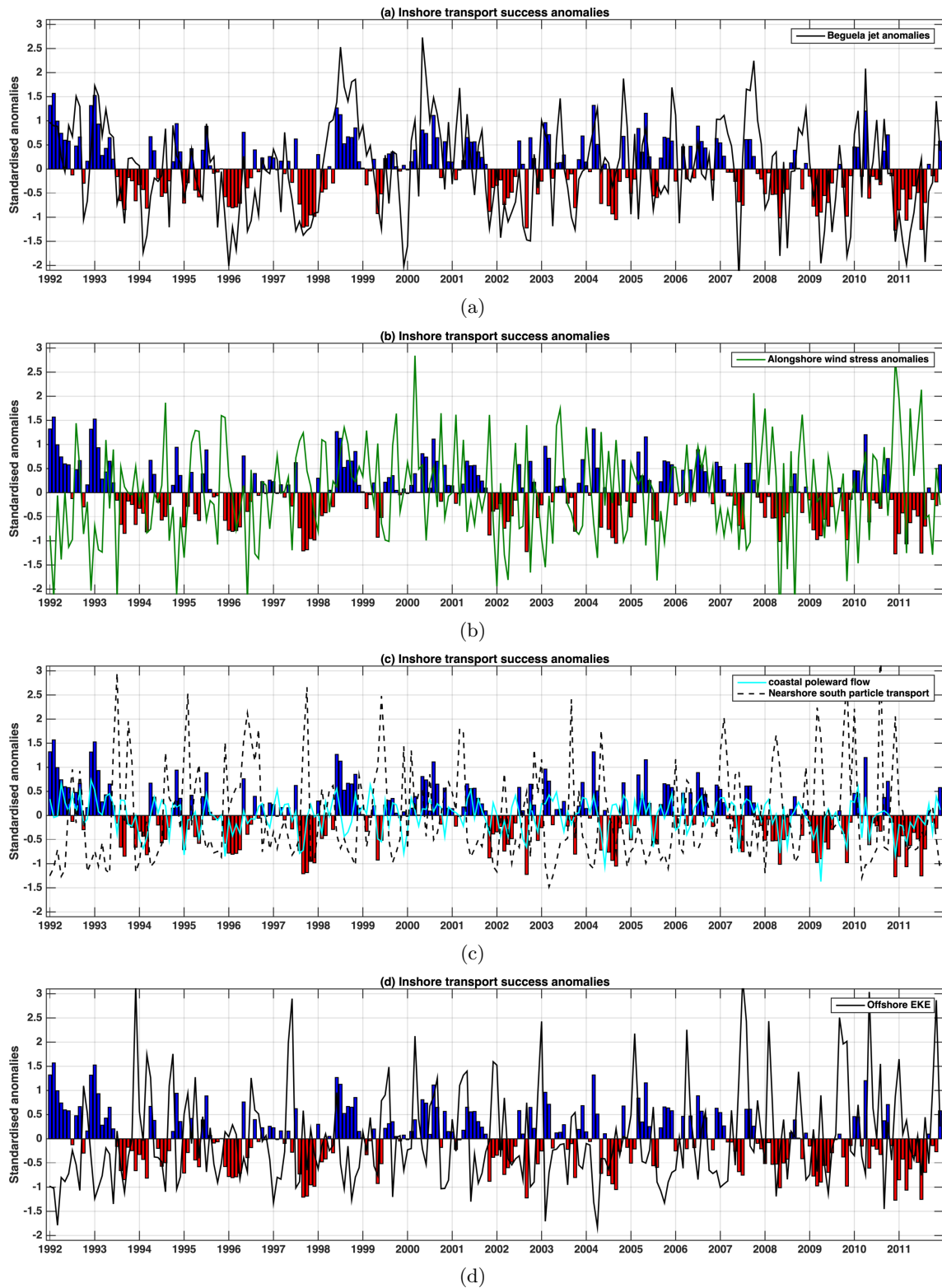


Figure 6.9: Standardised time-series of monthly inshore transport anomalies (bar graph) and superimposed is: (a) Benguela jet anomalies (black line), (b) Alongshore wind stress anomalies (green line), (c) Coastal poleward flow (cyan) plus the nearshore southward transport of particles (dashed black line) and, (d) the offshore EKE anomalies.

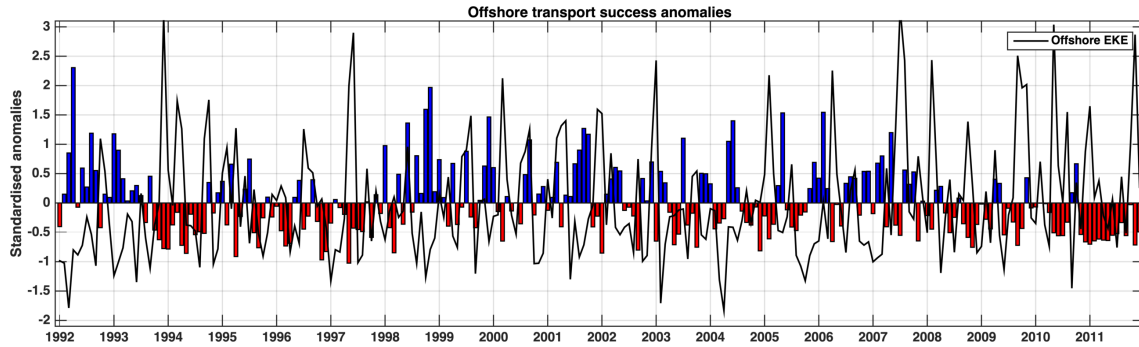


Figure 6.10: Standardised time-series of monthly offshore transport anomalies (bar graph) and EKE anomalies are superimposed as a black line.

2090 transport success and the coastal poleward flow anomalies (Fig. 6.9c). The southward
 2091 transport was determined by calculating the number of particles advected south from the
 2092 inshore domain (<200 m isobath) which crosses 35°S. Indeed, the southward transport of
 2093 these particles shows that positive southward transport anomalies are related to stronger
 2094 poleward current ($r = -0.34$, $p < 0.05$) (Fig. 6.9c). However, there are some occurrences
 2095 when this relationship does not hold, such as during mid-1998 where there was a stronger
 2096 than usual poleward flow (negative anomalies) that drove a southward transport but pos-
 2097 itive anomalies in inshore transport. Furthermore, 2011 showed a stronger than usual
 2098 poleward flow but negative anomalies of south transport did not occur.

2099

2100 Offshore mesoscale turbulence (EKE) might have an impact on inshore transport suc-
 2101 cess if the particles become advected offshore into this turbulent region. Indeed, a sig-
 2102 nificant negative correlation exists ($r = -0.28$, $p < 0.05$) (Fig. 6.9d). Correlation maps of
 2103 EKE and inshore transport success anomalies further reveal that this negative correlation
 2104 extends shoreward indicating that TSI' becomes negative when the high turbulence is lo-
 2105 cated closer to the shelf-edge.

2106

2107 The anomalies of offshore transport success and EKE are shown in Fig. 6.10 with a
 2108 significant negative correlation ($r = -0.31$, $p < 0.05$) between them. As expected increased
 2109 mesoscale turbulence leads to more dispersion which is unfavourable for offshore transport
 2110 success. However, it is essentially the increased EKE case that is important as when re-
 2111 duced EKE events are considered, there is no obvious impact on transport success. There

2112 are also cases such as during late 1994 and early 1997 when negative offshore transport
2113 was associated with a negative offshore EKE.

2114

2115 Our analysis reveals that the transport success variability cannot be described by only
2116 one individual physical driver. Future work should preferably use the combinations of
2117 multiple drivers. Although multiple drivers are typically responsible for inshore transport
2118 success anomalies, it appears that particles must remain inshore where they can trans-
2119 ported northward by the Benguela jet for positive transport success. Also, periods with
2120 anomalously strong upwelling winds result in the removal of particles away from the shelf
2121 via offshore Ekman transport. This offshore loss can be counteracted to some extent when
2122 the Benguela Jet is anomalously strong and thus able to act as a barrier. On the other
2123 hand, a weak Benguela jet does not necessarily mean negative inshore transport, espe-
2124 cially if there is no offshore or southward loss of particles. Coastal poleward flow driven
2125 by negative wind stress curl, can lead to southward advection of particles. It is the off-
2126 shore region which experiences strong mesoscale turbulence. Consequently, anomalously
2127 high mesoscale turbulence which if sufficiently large may leads to the dispersion of offshore
2128 particles or those inshore particles that have become advected offshore.

2129 **6.5 Cross-shore transport**

2130 Cross-shore transport determines the extent to which particles are impacted by the off-
2131 shore mesoscale dynamics . In this section, we investigate whether particles have drifted
2132 further offshore away from the influence of the coastal Benguela jet and how this offshore
2133 advection may have affected transport success. Furthermore, we consider how this offshore
2134 transport could be linked to the influence of mesoscale eddies.

2135

2136 To determine the cross-shore displacement (Δx) of successful particles, the distance
2137 (x) of an individual particle trajectory is calculated from the 500 m isobath at their point
2138 of release ($t_l = 1$) to a time before their entry into the target area. Δx is calculated as
2139 follows:

$$\Delta x_i(t_l) = x_i(t_l) - x_i(t_{l=1}) \quad (6.4)$$

2140 where i is the individual Lagrangian particle, x is the distance of the particle trajectory
 2141 from the 500 m isobath at time t_l (12 hourly intervals). For the purpose of this analy-
 2142 sis, four cases of cross-shore displacements are considered: two are for successful inshore
 2143 particles, and the other two are for successful offshore particles.

2144 6.5.1 Inshore particles

2145 A first group of particles was selected by identifying those having, at any time, a distance
 2146 (x) lower than 50 km. Each particle belonging to this inshore group was never displaced
 2147 during its journey more than 50 km west (offshore) of its initial distance from the 500 m
 2148 isobath. These particles are considered to have travelled inshore (TSII). Anomalies (TSII')
 2149 were standardised as follows:

$$TSII'_m = \frac{TSII_m - \overline{TSII_m}}{\sigma TS_m} \quad (6.5)$$

2150

2151 The second inshore group are defined as those that were displaced at any time, more
 2152 than 50 km west of their initial position from the 500 m isobath and are considered to
 2153 have travelled offshore (TSIO). Anomalies (TSIO') are again standardised as follows:

$$TSIO'_m = \frac{TSIO_m - \overline{TSIO_m}}{\sigma TS_m} \quad (6.6)$$

2154

2155 An example of typical trajectories considered as belonging to the TSII (cyan) and TSIO
 2156 (magenta) groups are presented in Fig. 6.11. The trajectory associated with TSIO group
 2157 is characterised by a long-distance offshore loop and it is more likely that this particle has
 2158 been impacted by offshore mesoscale turbulence.

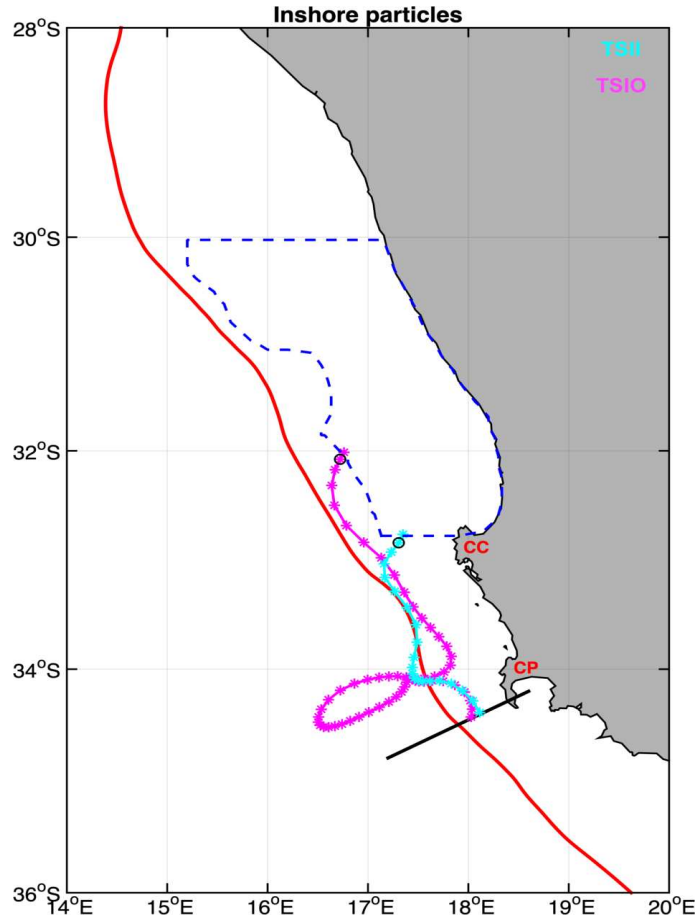


Figure 6.11: An example of the two individual trajectories plotted from their initial position on the SARP line (black line) to their entry into the target area (blue dashed polygon) corresponding to the groups of cross-shore displacement of particles originating from the inshore section: TSII in cyan and TSIO in magenta.

2159 Figure 6.12a shows time-series of normalised anomalies of inshore transport (black line)
 2160 and of successful particles that did not go further than 50 km west of their initial distance
 2161 from the 500 m isobath (TSII'). Inshore transport success anomaly is significantly corre-
 2162 lated with TSII' ($r = 0.85$, $p < 0.05$). The two time-series appear to be highly coherent.
 2163 It means that a majority of the successful particles that were released inshore did not
 2164 extend at any time more than 50 km west of their initial distance from the 500 m isobath.
 2165 The significant positive ($r = 0.59$, $p < 0.05$) correlation of TSII' with the Benguela Jet
 2166 anomalies suggest that the jet entrainment of the inshore particles is important for their
 2167 success. On the other hand, negative anomalies of TSII' are impacted by their removal
 2168 (loss) from the shelf via offshore Ekman transport and southward transport. There is a
 2169 weak but still significant negative correlation ($r = -0.17$, $p < 0.05$) between TSII anoma-

2170 lies and alongshore wind anomalies suggesting that positive wind anomalies are relatively
 2171 important. A stronger relationship (positive) exists between $TSII'$ and the poleward flow
 2172 ($r = 0.41, p < 0.05$) but that with EKE is relatively weak and negative ($r = -0.25, p < 0.05$).

2173

2174 On the other hand, there is no correlation ($r = 0.07$) between inshore transport success
 2175 anomalies (particles that were released inshore) and successful inshore particles that were
 2176 once displaced during their journey more than 50km west of their initial position from the
 2177 500m isobath ($TSIO'$) (Fig. 6.12b). Their anomalies have little impact on TSI' , most of
 2178 whose variability comes from $TSII'$ group. This further suggests that for successful trans-
 2179 port of inshore particles, the particles have to remain on the shelf under the influence of
 2180 the Benguela Jet and that they have low a chance of returning if they become transported
 2181 out into the turbulent offshore domain.

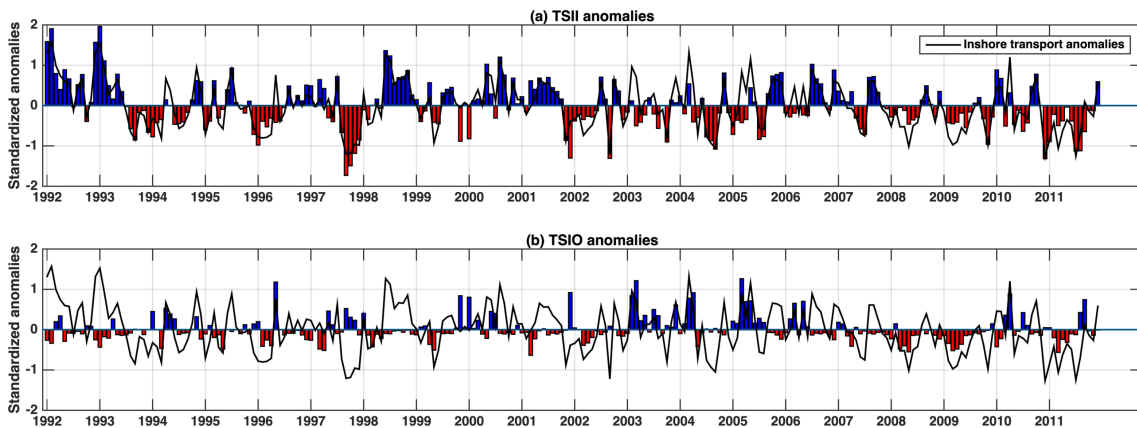


Figure 6.12: Standardised time-series (bar graph) of monthly successfully transported: (a) inshore particles that were displaced less than 50 km offshore from their initial distance from the 500 m isobath ($TSII'$) and, (b) inshore particles that were displaced more than 50 km offshore from their initial distance from the 500 m isobath ($TSIO'$) anomalies. Anomalies of inshore transport success (TSI') are superimposed in a black line.

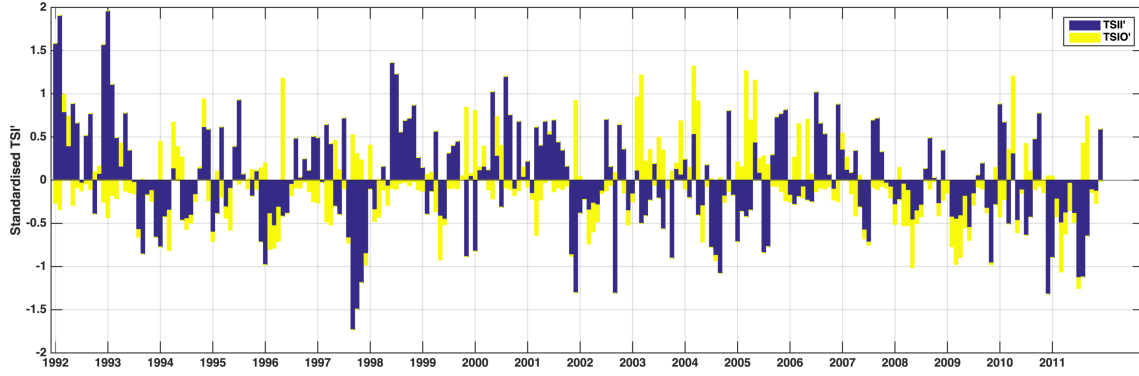


Figure 6.13: Inshore transport success anomalies (TSI') shown as cumulative bars of successful particles that were not displaced beyond 50 km of their initial distance from the 500 m isobath ($TSII'$: blue) and inshore particles that were displaced more than 50 km of their initial distance from the 500 m isobath ($TSIO'$: yellow) such that $TSI' = TSII' + TSIO'$.

2182 6.5.2 Offshore particles

2183 Similarly, the offshore displacement of particles originating in the offshore section is anal-
 2184 ysed. The TSOI group is associated with successful offshore particles that did not get
 2185 displaced more than 50 km west (offshore) of their initial distance from the 500 m isobath
 2186 and are considered to have followed the shelf-edge. An example of such a trajectory is
 2187 shown as a purple line in Fig. 6.14. Their normalised anomalies were calculated as follows:

$$TSOI'_m = \frac{TSOI_m - \overline{TSII}_m}{\sigma TS_m} \quad (6.7)$$

2188

2189 The second offshore group (TSOO) is associated with successful offshore particles that
 2190 were displaced, at least once, to a distance greater than 50 km west (offshore) of their
 2191 initial distance from the 500 m isobath. These particles were advected away from the shelf-
 2192 edge and further into the open ocean. An example of their trajectory example is shown
 2193 as the green line in Fig. 6.14. Their normalised anomalies were calculated as follows:

$$TSOO'_m = \frac{TSOO_m - \overline{TSII}_m}{\sigma TS_m} \quad (6.8)$$

2194

2195 For offshore particles, there is a positive correlation ($r = 0.9$, $p < 0.05$) between off-

2196 shore transport success anomalies and TSOI anomalies (Fig. 6.15a). A significant negative
2197 correlation ($r = -0.31$, $p < 0.05$) exists with offshore EKE anomalies.

2198

2199 There is a significant but relatively weak positive relationship ($r = 0.24$, $p < 0.05$), be-
2200 tween TSOO anomalies and offshore transport success anomalies (Fig. 6.15b). This result
2201 suggests that particles from the offshore section that become displaced further offshore
2202 can return somehow and contribute significantly to offshore transport success. However,
2203 no relationship was found between this group and EKE (mesoscale turbulence).

2204 In summary, positive transport success of offshore particles is significantly related to par-
2205 ticles that don't get advected too far from the shelf-edge, and their success is linked to low
2206 mesoscale turbulence. Interestingly, some offshore particles that get transported further
2207 into the open ocean do manage to come back inshore and contribute to positive offshore
2208 transport success. However, these offshore particles were found not to be linked to the
2209 mesoscale turbulence. Thus, cross-shore transport mechanism could be better explained
2210 by the circulation patterns associated with mesoscale eddies rather than EKE associated
2211 with their turbulence.

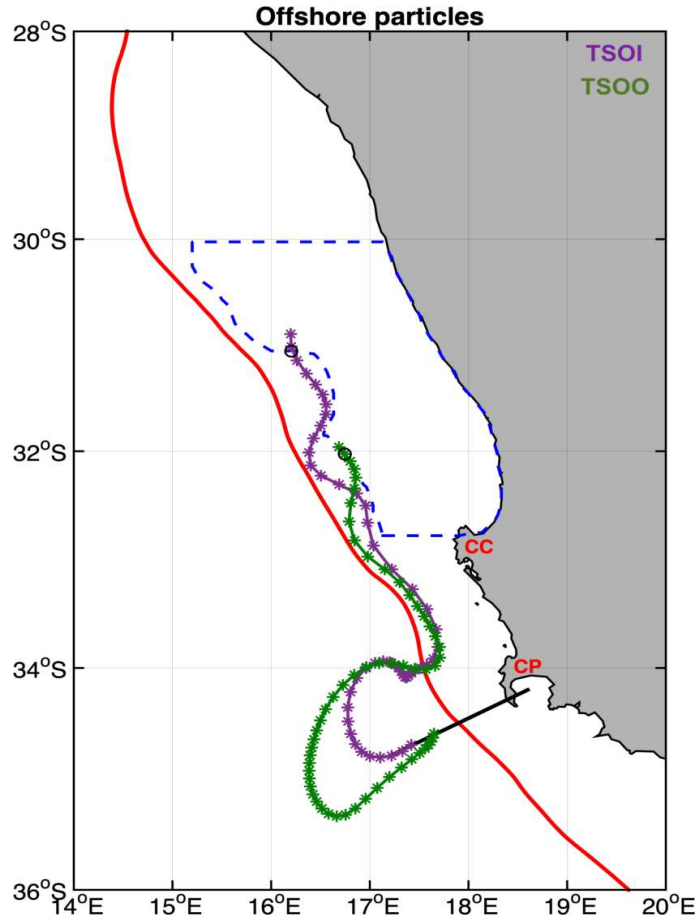


Figure 6.14: Same as Fig. 6.11 but corresponding to the groups of cross-shore displacement of particles originating from the offshore section: TSOI in purple and TSOO in green

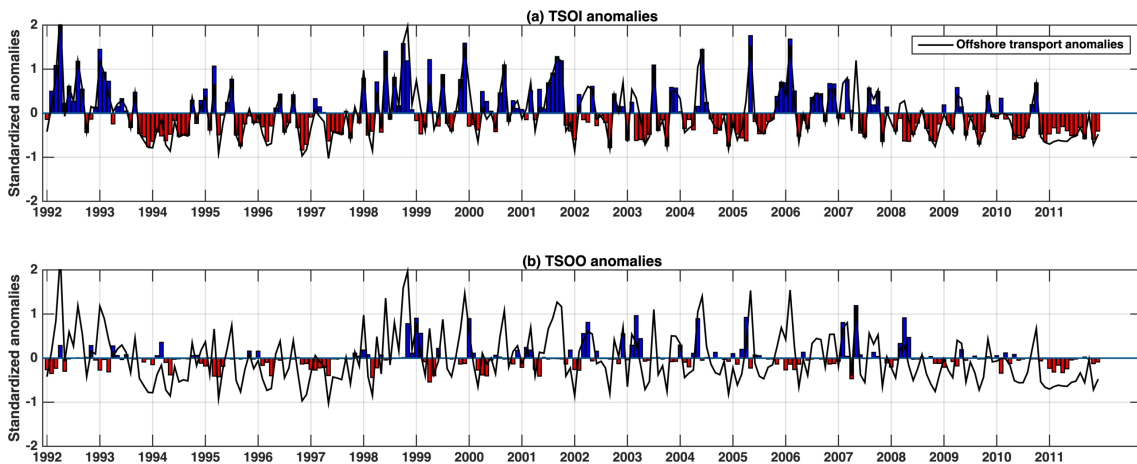


Figure 6.15: Normalised time-series (bar graph) of monthly: (a) offshore particles that were never displaced during their journey more than 50 km offshore from their initial distance from the 500 m isobath (TSOI') and, (b) offshore particles that were displaced at least once, more than 50 km offshore from their initial distance from the 500 m isobath (TSOO') anomalies. Anomalies of offshore transport success (TSO') are superimposed in a black line.

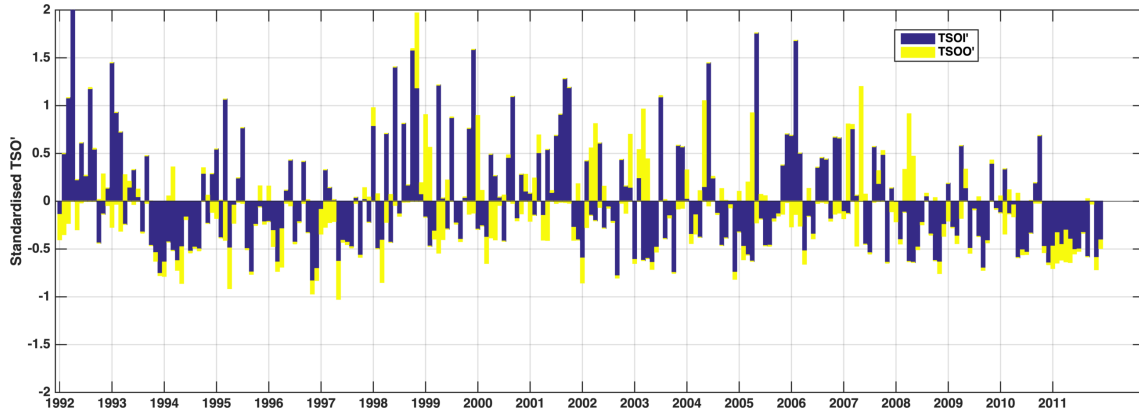


Figure 6.16: Offshore transport success anomalies (TSO') shown as cumulative bars of successful particles that were not displaced beyond 50 km of their initial distance from the 500 m isobath ($TSOI'$:blue) and inshore particles that were displaced more than 50 km of their initial distance from the 500 m isobath ($TSOO'$:yellow) such that $TSO' = TSOI' + TSOO'$

2212 6.6 The role of mesoscale eddies

2213 As previously noted, particles need to be retained over the shelf region in order to achieve
 2214 transport success, whereas the offshore particles need to be advected onshore into the
 2215 Benguela Jet away from the turbulent eddy field to become successful. It is suggested
 2216 that Ekman dynamics can play a role in the cross-shore displacement of the inshore par-
 2217 ticles.

2218 Offshore loss of inshore particles can be attributed to strong offshore Ekman transport
 2219 whereas retention is favoured by a weak offshore Ekman transport. Although that might
 2220 be true in general, there are also some events where the loss of inshore particles was not
 2221 related to offshore Ekman transport or to southward transport. Thus wind alone does not
 2222 generate the cross-shelf transport, so other mechanisms must be responsible. However, it
 2223 is also unknown as to which mechanism is responsible for the onshore transport of particles
 2224 that do not get dispersed into the open ocean.

2225 One possible mechanism could be a temporary localised cross-shore current induced by
 2226 mesoscale eddies. Indeed, visual analysis of the particle trajectories reveals that the trajec-
 2227 tories of particles found in the offshore domain are very chaotic and characterised by large
 2228 loops indicative of mesoscale eddy structures. It is also evident that offshore (westward)
 2229 transport is associated with anticyclonic eddies that pass north-westward in the vicinity

2230 of offshore Cape Point. On the other hand, shoreward (eastward) transport occurs from
2231 the circulation associated with shelf-edge cyclonic eddies in a region between just north
2232 of the SARP line and offshore of St Helena Bay.

2233

2234 However, before investigating the role of mesoscale eddies, it is useful to first charac-
2235 terise the eddy mean properties such as their sign, radius, amplitude, duration and their
2236 point of origin in the SBUS. An eddy detection and tracking algorithm was applied to the
2237 daily SSH fields from the R3km simulation for the 1992-2011 period for this purpose. The
2238 details of this algorithm may be found in Appendix B.

2239 Table 6.1 shows the mean eddy properties that were detected and tracked in the SBUS
2240 between 36°S - 32°S and 14°E - 20°E . A total of 2923 mesoscale eddies were tracked within
2241 the region with the vast majority of them being cyclonic (73.8%). The cyclonic eddies
2242 mostly originate over the shelf and along the shelf-edge (Fig. 6.17a), tend to have a smaller
2243 mean diameter (39.9 km) and amplitude of 3.1 cm and have a shorter duration (11 days).
2244 The most massive cyclonic eddy detected was 160 km in diameter and generated north-
2245 east of the Agulhas retroflection. The predominance of cyclones is consistent with the
2246 findings of *Rubio et al.* (2009); however, these authors were not able to resolve the smaller
2247 cyclonic structures between the coast and the 500 m isobath due to their coarse resolution
2248 model ($1/10^{\circ}$).

2249 The slightly larger (diameter ~ 30 km and amplitude ~ 2 cm) anticyclonic eddies mostly
2250 originate over the deep ocean, with the largest eddies (>150 km) originating further south
2251 in the region of the Agulhas Current retroflection (Fig. 6.17b). Their larger size and less
2252 frequent occurrence are consistent with what is expected of Agulhas rings. The largest
2253 anticyclonic eddy detected was 230 km, again consistent with a Agulhas ring (*Lutjeharms*
2254 *and Van Ballegooyen*, 1988; *Boebel et al.*, 2003).

2255 **Anticyclonic eddies**

2256 Anticyclones with diameters that range from 20 km to 230 km were tracked, and it was
2257 found out that there was some indication that the passage of large anticyclones near the
2258 SARP line can result in the offshore transport of particles. *Duncombe-Rae et al.* (1992)

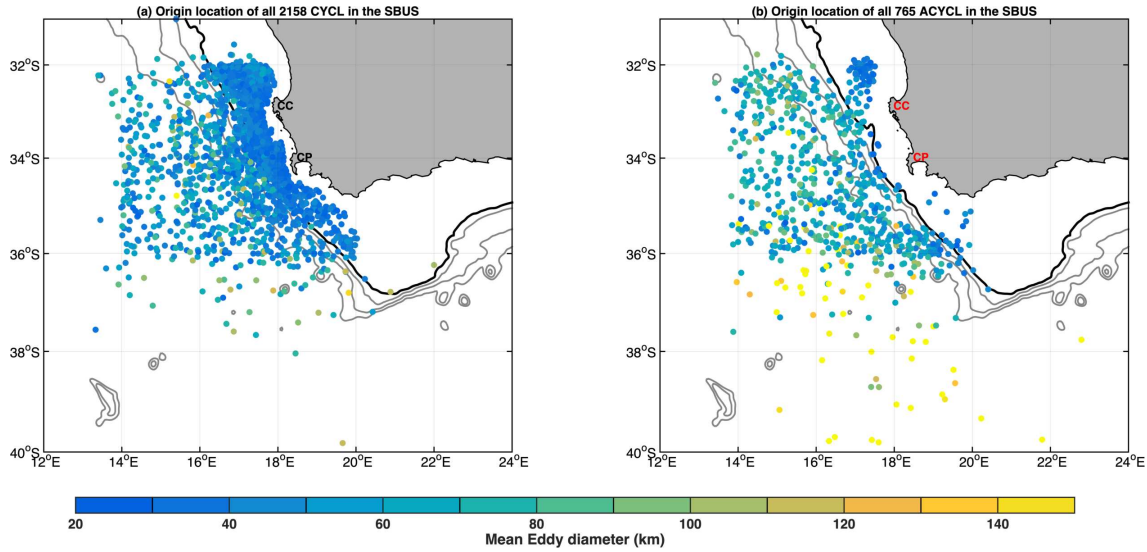


Figure 6.17: A map showing the distribution of eddies by their generation site indicated by the circles and their mean diameter (in color) during their duration in the region $36^{\circ}\text{S} - 32^{\circ}$ and $14^{\circ}\text{E} - 20^{\circ}\text{E}$ for: (a) cyclonic eddies and, (b) anti-cyclonic eddies. The background contours are the 500 m isobath in black and in grey is the 1000 m, 2000 m and 3000 m isobaths

2259 has documented the effects of an Agulhas Ring interacting with the Benguela upwelling
 2260 front which generated a long filament of cold water extending further offshore (*Payne*
 2261 *et al.*, 1992). It has been suggested that these filaments can entrain coastal biota a long
 2262 distance offshore because of their associated offshore volume transport which *Lutjeharms*
 2263 *et al.* (1991) estimated to be about 1.5 Sv.

2264

2265 To evaluate the possible influences of a Agulhas ring on the region, large anticyclones
 2266 with a radius (R_{eddy}) greater than 100 km (200 km diameter) were identified and assumed
 2267 to be Agulhas rings from their origin point and trajectories. A total of 38 Agulhas rings
 2268 were identified in the region $36^{\circ}\text{S} - 32^{\circ}$ and $14^{\circ}\text{E} - 20^{\circ}\text{E}$ between January 1992 and Decem-

	Cyclones	Anticyclones
Total number	2158 (73.8%)	765 (26.2%)
Duration (days)	11	18
Diameter (km)	39.9 (max = 160)	65.6 (max = 230)
Amplitude (cm)	3.1	5.2

Table 6.1: Summary of mean (1992-2011) cyclonic and anticyclonic eddy properties in the R3km simulation in the region $36^{\circ}\text{S} - 32^{\circ}$ and $14^{\circ}\text{E} - 20^{\circ}\text{E}$.

2269 ber 2011, their trajectories are shown in Fig 6.18a. These trajectories represent rings that
 2270 follow the northern route (Fig 6.18b) defined by *Dencausse et al.* (2010a). However, only
 2271 Agulhas rings whose paths were close to the SARP line (34°S - 35°S) are considered below.
 2272 Their closeness implies the interaction and influence of these rings on the distribution of
 2273 particles.

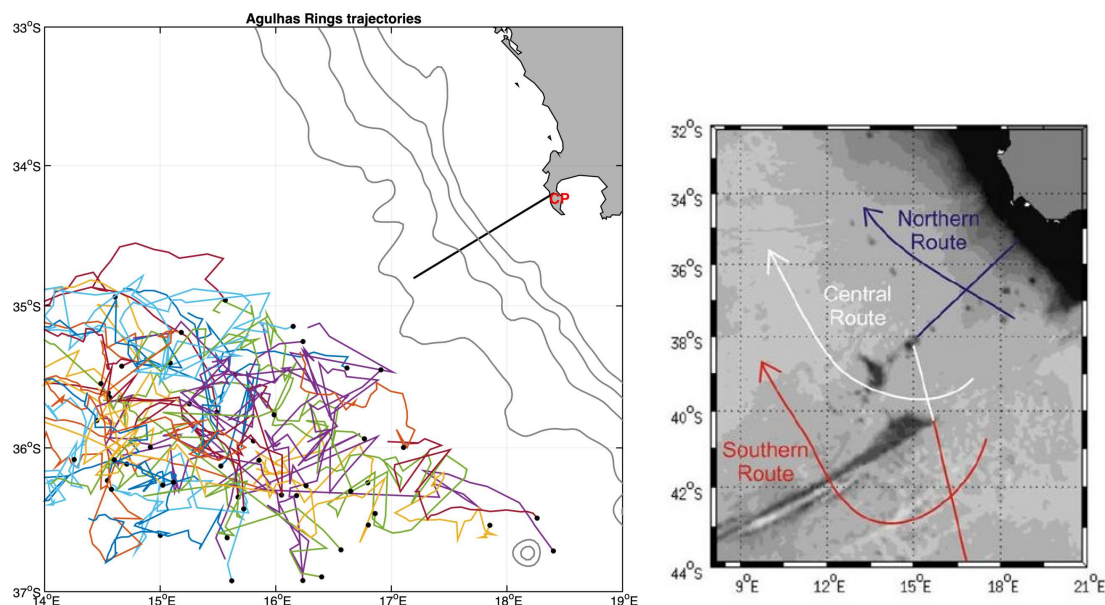


Figure 6.18: Left: trajectories of Agulhas rings (different colors) tracked into the south-eastern Cape Basin following the Northern route (blue) defined by *Dencausse et al.* (2010a) between January 1992 and December 2011. The black solid circles indicate where the rings were first detected. The black cross-shore line off Cape Point represents the SARP line and grey lines are the 500 m, 1000 m, 2000 m and 3000 m isobaths. Right: the three routes - Northern, Central and Southern defined by *Dencausse et al.* (2010a)

2274 An Agulhas Ring was considered to be in the range of influence in the region if the dis-
 2275 tance of its centre is less than $1.5 \times R_{eddy}$ from the 3000 m isobath and less than $1.5 \times R_{eddy}$
 2276 south of the SARP line. Out of the 38 Agulhas rings, 18 of them interacted with the
 2277 3000 m isobath for a total of 177 days with 8 of these interacting with the shelf-edge
 2278 (500 m isobath). The interaction duration with the shelf is relatively brief, a total of
 2279 36 days. Due to their brevity and infrequent interaction with the shelf-edge, the rings
 2280 were found not to affect the inshore transport success anomalies. Therefore only the re-
 2281 sults of their offshore influence are discussed below.

2282

2283 For each offshore interaction with the rings, their influence is evaluated through their
 2284 duration, minimum and mean distance during their interaction (Table 6.2) with their asso-
 2285 ciated offshore transport success anomalies. The results indicate that not all interactions
 2286 lead to the effect of offshore transport success. The level of impact is independent but is
 2287 also influenced by the minimum distance, mean distance and most importantly, its position
 2288 relative to the SARP line (offshore or south). Out of the 18 Agulhas ring interactions, only
 2289 5 (28%) resulted in positive offshore transport success anomalies, with the vast majority
 (72%) led to negative transport success.

Years	Months	Days of influ- ence	Minimum distance (km)	Mean dis- tance (km)	Offshore trans- port anomaly
1992	January*	12	58	91	negative (-0.40)
1993	March	10	102	115	positive (0.41)
	September*	3	86	90	positive (0.45)
1994	October	20	95	116	positive (0.35)
	November	7	120	139	negative (-0.17)
1996	November	10	121	136	negative (-0.94)
1997	April	1	105	105	negative (-0.19)
2000	January*	5	74	82	positive (0.60)
	February*	7	61	83	negative (-0.15)
2001	January	13	124	159	negative (-0.12)
	December*	7	77	98	negative (-0.22)
2002	January	5	116	127	negative (-0.85)
2004	November*	19	42	75	negative (-0.07)
	December*	7	86	102	negative (-0.82)
2008	September*	8	63	80	negative (-0.35)
	October*	31	51	77	negative (-0.56)
	November	9	83	119	negative (-0.75)
2011	October	3	91	99	negative (-0.03)

Table 6.2: A summary of the occurrence of an Agulhas Ring interaction with the 3000 m isobath in the vicinity of the SARP in terms of years, months, days of influence, its minimum distance (km) and mean distance (km). Their associated offshore transport success anomaly is also presented. The instances when the ring had a interaction with the shelf-edge (500 m isobath) is denoted by *.

2290

2291 The difference in the influence of Agulhas rings on the offshore displacement of parti-
 2292 cles can be inferred by looking at different scenarios of trajectories. Fig. 6.19 (top row)
 2293 represents a scenario associated with the passage of an Agulhas Ring that occurred in Jan-
 2294 uary 2000, which had less impact on the offshore displacement of particles or on offshore
 2295 transport success. Although the ring was located close to the 3000 m isobath, its centre
 2296 was situated well south of the SARP line. Hence, this ring had minimal impact on the

2297 cross-shore transport of particles and thus, on offshore transport success. Moreover, the
 2298 presence of a cyclonic eddy on the shelf-edge between Cape Point and Cape Columbine
 2299 (Fig. 6.19a) favoured the onshore transport of offshore particles (Fig. 6.19d).
 2300 On the other hand, the January 2002 shows how the combination of the closeness of the
 2301 Agulhas ring to the SARP line and its short interaction resulted in a large number particles
 2302 being entrained and thus transported offshore (Fig. 6.19: bottom row).

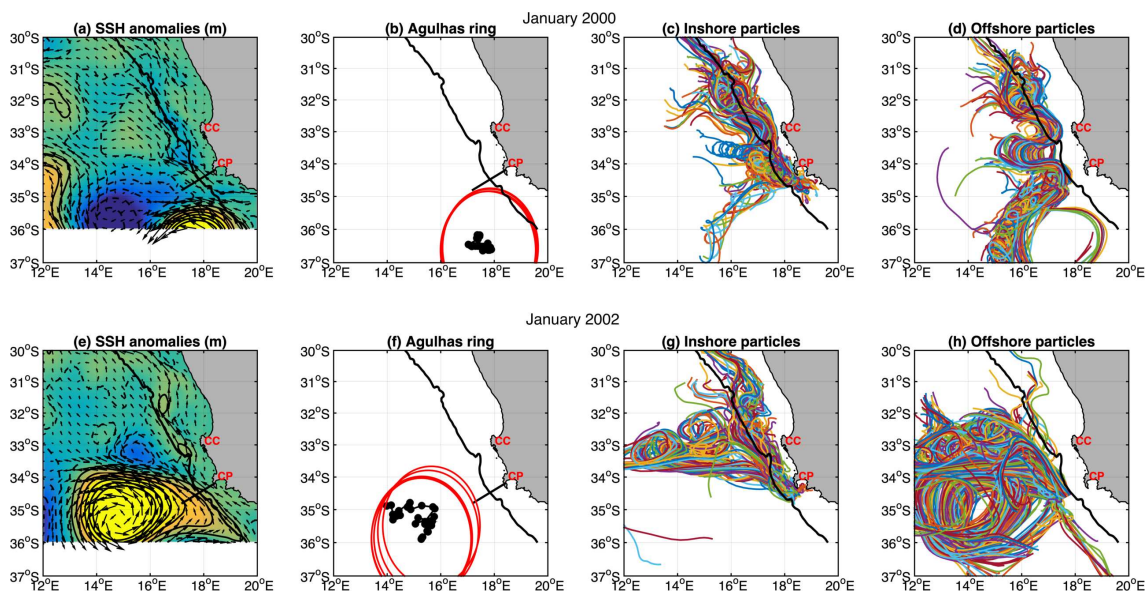


Figure 6.19: (a) Monthly mean SSH anomalies (m) in color and vectors superimposed corresponds to monthly mean surface current velocities. (b) The location of the detected Agulhas ring (black circles) during their interaction duration and the red circle represents size of the eddy at $1.5 \times R_{eddy}$. (c-d) shows the trajectories of inshore particles and offshore particles characteristic of the anticyclonic circulations. The black contour represents the 500 m isobath. The top row represents the scenario which is associated with an Agulhas ring was passed further south (January 2000) and whereas, the bottom row represents the scenario associated an Agulhas ring passing north near the SARP line (January 2002).

2303 Cyclonic eddies

2304 There is evidence that positive offshore transport anomalies (*i.e.* particles that were release
 2305 on the offshore side of the SARP line) are influenced mainly by shelf-edge cyclonic eddies
 2306 since these eddies transport offshore particles shoreward. The crucial feature favouring the
 2307 on-shelf transport of particles is the small size of the cyclonic eddies, their short duration
 2308 and their clockwise circulation. However, despite their frequent occurrence at the shelf-
 2309 edge, their influence is difficult to quantify, as particles are also affected by anticyclonic

2310 eddies at the SARP line. Fig. 6.20 shows an example (April 1992) illustrating how a
 2311 shelf-edge cyclonic eddy can affect the cross-shelf transport of particles. Note the onshore
 2312 transport of particles in the blue box (Fig. 6.20c,d) and their coincidence with the location
 2313 of shelf-edge eddies (Fig. 6.20b). This month had the highest positive offshore transport
 2314 SUCCESS.

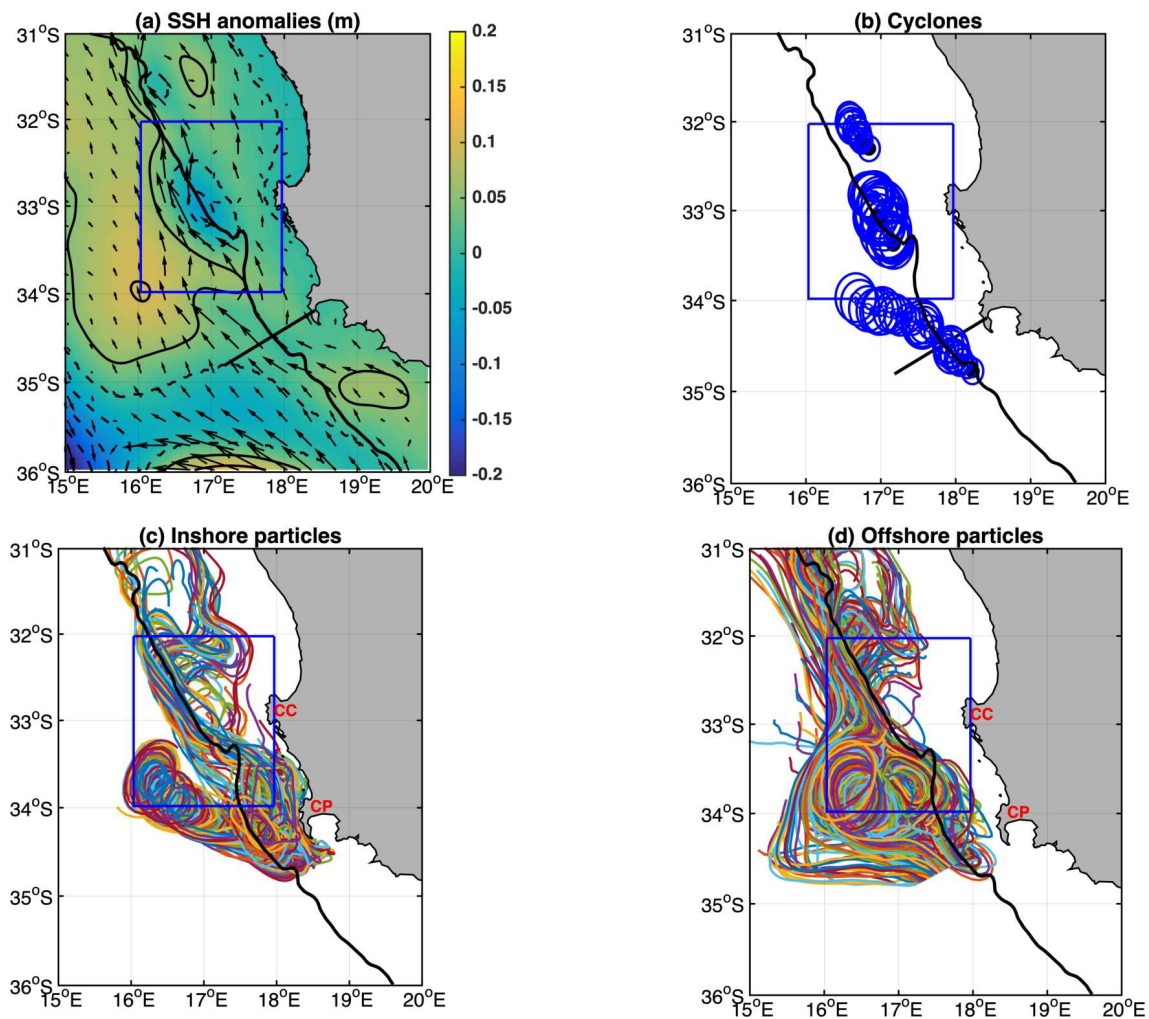


Figure 6.20: (a) April 1992 monthly mean SSH anomalies (m) in color and vectors superimposed corresponds to monthly mean surface current velocities. (b) The number of cyclonic eddies (blue circles) detected between Cape Point and Cape Columbine with their position at each time-step during their life-time is indicated by individual circles. (c-d) shows the trajectories of inshore particles and offshore particles characteristic of the cyclonic circulations.

2315 In summary, the results suggest that mesoscale eddies play a role in the cross-shore
 2316 transport of particles. However, the relationship is ambiguous because the characteris-
 2317 tics of coherent eddy structures are very variable in terms of their different polarities

2318 (circulation patterns), life-time, sizes, intensity and their proximity to the shelf or slope.
 2319 Nevertheless, transport success variability of the offshore particle is strongly linked to the
 2320 propagation of Agulhas rings, especially the north-westward propagating rings. The large
 2321 size, long occupation, a strong associated anti-clockwise circulation of the rings leads to a
 2322 more westward distribution of particles and negative offshore transport. However, these
 2323 rings may have less impact if their passage occurs further south of the SARP line.
 2324 These results suggest that shelf-edge cyclonic eddies found between Cape Point and Cape
 2325 Columbine can lead to favourable offshore transport success by transporting particles in-
 2326 shore into the Benguela Jet. However, since these eddies are present almost all the time,
 2327 it is not easy to individually link them to specific events of transport success. Thus their
 2328 influence is dependent on whether the initial conditions at the SARP are favourable for
 2329 the northward transport instead of offshore.
 2330 The quantification of the impact of mesoscale eddies is further complicated by the co-
 2331 existence of both cyclonic and anticyclonic eddies. Fig. 6.21 shows that there is a positive
 2332 correlation between anomalies of transport success and nearshore EKE between north
 2333 of Cape Point and offshore off St Helena where Fig. 6.17a indicate cyclonic eddies are
 prevalent.

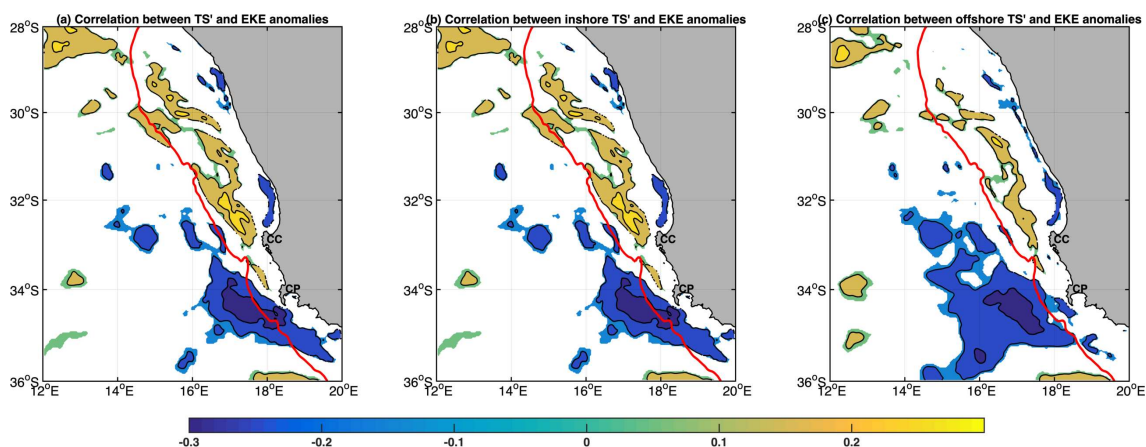


Figure 6.21: Correlation map showing only the significant correlation at 95% confidence level between monthly transport success time-series and EKE anomalies: (a) total transport success anomalies (TS'), (b) inshore transport success anomalies (TSI') and, (c) offshore transport success anomalies (TSO'). The 500 m isobath separating the inshore and offshore region is superimposed as the red line.

2334

2335 This result supports the positive impact of these eddies on transport success. Note
 2336 that the nearshore negative correlation off Cape Point to the south of it seen in Fig. 6.21b

2337 corresponds to cyclonic eddies whose presence here was observed to transport particles
2338 southward, and thereby contributing to negative inshore transport success. Whereas the
2339 region of negative correlation offshore coincides with the north-westward propagation of
2340 anticyclonic eddies with the offshore influence limited to 15°E. This impact of the anticy-
2341 clonic eddies (Agulhas rings) depends on their closeness to the SARP line.

2342 **6.7 Discussion and conclusions**

2343 In the southern Benguela upwelling system, variability in annual recruitment of small
2344 pelagic fish has long been a significant concern to fisheries oceanographers. Successful
2345 advection is one of the factors required for the survival of fish larvae, and this chapter was
2346 aimed at relating anomalous transport success events to local physical processes such as i)
2347 wind surface forcing (Ekman transport, Benguela Jet, poleward current) and ii) mesoscale
2348 dynamics. The interannual variability in physical transport processes is investigated using
2349 a 3 km high-resolution simulation forced with 6-hourly atmospheric CFSR data provided
2350 for 20 years (1992-2011). The approach consisted of computing monthly transport success
2351 anomalies and characterising their wind-driven ocean circulation and mesoscale turbu-
2352 lence.

2353

2354 In the absence of mesoscale eddy influence (especially anticyclones), the Benguela Jet
2355 was found to play a pivotal role in the successful transport variability of particles originat-
2356 ing inshore. If inshore particles remain entrained in the coastal Benguela Jet, they have a
2357 higher probability of being successfully transported. More importantly, it was shown that
2358 mesoscale eddies also contribute significantly to the variability of successful transport for
2359 particles originating offshore in particular. There is a difference in the role of cyclonic
2360 shelf-edge eddies off Cape Point that cause negative transport success and the cyclonic
2361 shelf-edge eddies off Cape Columbine and St Helena Bay that lead to more positive trans-
2362 port success.

2363 In contrast, the passage of large anticyclonic eddies (Agulhas rings) in the vicinity of Cape
2364 Point can have a negative effect if they pass close enough to interact with the particles
2365 originating offshore.

2366

2367 Over the shelf, most of the variability in the transport success of particles originating
2368 from inshore is associated with a coherent Benguela Jet between the 200 m and 500 m
2369 isobath. Negative events of transport success could be related to their removal by wind-
2370 driven shelf circulation such as strong offshore Ekman transport and by a poleward current.
2371 However, the influence of Ekman transport is relatively small because a strong Benguela
2372 Jet can inhibit the offshore transport of inshore particles by the wind. Hence, the impact
2373 of large-scale interannual climate modes perturbations such as ENSO and SAM which in-
2374 fluence wind forcing over the region was found to be small (not shown) and was therefore
2375 not discussed in this chapter.

2376

2377 For particles originating offshore to have positive transport success, the particle should
2378 not get advected offshore into the open ocean away from the influence of the shelf-edge
2379 cyclonic eddies. The passage of large anticyclonic eddies close to the SARP line was found
2380 to have a negative impact on transport success as they are associated with westward sur-
2381 face velocities, which advect particles away from the shelf-edge. This impact is limited
2382 to being only likely near Cape Point (35°S-34°S). This is not surprising, since the region
2383 is located in the vicinity of the 'Cape Cauldron' (*Lutjeharms et al.*, 2003) where there is
2384 high turbulence associated with the north-westward passage of Agulhas rings and other
2385 eddies (*Veitch et al.*, 2009; *Veitch and Penven*, 2017). Interaction of an Agulhas ring
2386 with the southern Benguela shelf-edge was found to be relatively infrequent since only 18
2387 interactions occurred over the 20 years model simulation.

2388

2389 This study investigates in detail how physical processes can impact transport success
2390 between Cape Point and St Helena Bay. The next step is to investigate whether our results
2391 can help to explain and understand anchovy recruitment variability. This will require the
2392 incorporation of biological properties such as egg density, diel vertical motion, mortality
2393 due to starvation and predation (*Huggett et al.*, 2003; *Mullon et al.*, 2003; *Parada et al.*,
2394 2003, 2008; *Koné et al.*, 2013). This is however out of scope of this thesis but we are
2395 convinced that the physical processes we identified and the interannual variability of their

2396 contribution to transport success could be used to bring some light on the processes gov-
2397 erning small pelagic fish recruitment variability in the SBUS.

2398

2399 This study confirms that the variability Benguela Jet is a primary driving mechanism
2400 the alongshore transport success and thus can help to explain some of the variability ob-
2401 served in recruitment strength of small pelagic fish. Therefore, it is essential to study
2402 response of the jet to large-scale influences such as Agulhas leakage, modes climate of
2403 variability as well as its long-term trends in order to anticipate changes that might occur
2404 in the light of climate change. It is known for example that the Agulhas leakage itself
2405 is sensitive to both the mid-latitudes westerlies and the trade winds over the South In-
2406 dian Ocean (*Durgadoo et al., 2013; Loveday et al., 2014*). A better understanding of the
2407 Benguela Jet would help in the management of the marine ecosystem in the area.

2408

2409 This chapter also showed that the cross-shore transport contributes to the variability
2410 in transport success through the activity of mesoscale eddies. It is therefore critical to
2411 understand the dynamics of the cross-shore transport due to its importance for variabil-
2412 ity of regional ecosystems. Mesoscale eddies were found to exert a stronger influence of
2413 transport success than the local winds.

2414

2415 It is important to note that the results presented may be sensitive to model resolution
2416 and to the ability of the model to adequately capture the Agulhas leakage. Overestimation
2417 of the Agulhas leakage and their associated mesoscale turbulence which often occurs in
2418 coarse resolution models is likely to lead to unrealistic estimates of transport success.
2419 Moreover, underestimation of the abundance of small cyclonic shelf-edge eddies in such
2420 models could also lead to the underestimation of the inshore advection of offshore particles
2421 and associated transport success.

Chapter 7

Summary and future research

There have been numerous previous works interested in the alongshore Lagrangian connectivity in the southern Benguela upwelling system (SBUS) for fisheries related applications (*Boyd et al.*, 1998; *Hutchings et al.*, 1998; *Huggett et al.*, 2003; *Mullon et al.*, 2003; *Parada et al.*, 2003; *Lett et al.*, 2006; *Miller et al.*, 2006; *Parada et al.*, 2008; *Koné et al.*, 2013). In those studies, transport success was investigated as one of the factors modulating recruitment success and used as proxy for the Western Agulhas Bank/southern Benguela connectivity. These studies were based on the advection of Lagrangian particles by ocean currents produced by regional ocean circulation models. They discussed the sensitivity to spawning depth, the spawning area and some biological parameters like mortality (below a given temperature threshold and time-threshold to reach the recruitment area) (*Huggett et al.*, 2003; *Mullon et al.*, 2003; *Parada et al.*, 2003; *Lett et al.*, 2006; *Miller et al.*, 2006; *Parada et al.*, 2008; *Koné et al.*, 2013). However, none of this previous work analysed in detail the characteristics of the ocean dynamics that led to the simulated transport success and variability. In addition, most of these studies used outputs from ocean models with a coarse horizontal resolution (9 km for the finest one), which is not sufficient to accurately represent the mesoscale structures associated with the Benguela Jet and even less the inner shelf (less than 100 m depth) flows. Moreover, the ocean models were forced with atmospheric seasonal monthly climatology precluding studies of externally forced interannual variability of transport success.

In this thesis, numerical model simulations of the SBUS coupled with particle tracking

2444 experiments were used to investigate Lagrangian pathways in the SBUS and their link to
2445 the oceanic circulation. This work is motivated by questions arising from the effect of
2446 ocean currents on fish eggs and larvae during their journey from the spawning to nursery
2447 grounds in the southern Benguela ecosystem. We used newly developed 20-year realistic
2448 simulations of the Benguela upwelling system and a series of Lagrangian experiments to
2449 identify the pathways connecting the spawning and nursery grounds as well as important
2450 processes impacting the connectivity between the two regions.

2451

2452 In **Chapter 3**, a mesoscale eddy resolving ($1/15^\circ$, 'BENGR15') numerical simulation
2453 of the Benguela system the Regional Ocean Modelling System (ROMS) used as a refer-
2454 ence simulation is described. The simulation spans from 1989 to 2010 (22 years) with bulk
2455 surface forcing (wind, heat and freshwater fluxes) that used CFSR atmospheric reanalysis
2456 and a seasonal monthly climatology extracted from SODA at the lateral open boundaries.
2457 In contrast to previous studies, our ocean model is forced by 6-hourly variable surface
2458 forcing. The aim is to build a robust model with realistic mesoscale eddy field, Agulhas
2459 leakage and the coastal circulation.

2460 The BENGR15 configuration is able to capture the general mean circulation patterns
2461 and the seasonal variability typical of an upwelling system such as the equatorward jet
2462 (Benguela jet), a slope-edge poleward undercurrent (>400 m depth) and shelf-edge insta-
2463 bilities related to the jet. The model was also able to capture the strong eddy kinetic energy
2464 (EKE) associated with the Agulhas Current and the Agulhas retroflexion. However, the
2465 simulated retroflexion overestimates the leakage of warm and saline Agulhas waters into
2466 the South Atlantic Ocean. This overestimation resulted in the model exhibiting a warm
2467 bias and strong cross-shore temperature gradients that drive an associated strong Benguela
2468 jet. Although the simulated jet is much stronger than observed, the model adequately rep-
2469 represents the characteristic seasonal variability of the jet. This provided the confidence in
2470 using this simulation to study the seasonal variability of the alongshore connectivity.

2471

2472 In **Chapter 4**, the BENGR15 configuration is used to characterise the seasonal along-
2473 shore connectivity between Cape Point (34°S) and St Helena Bay (32°S) and to also

2474 identify important physical processes impacting this connectivity. We show that trans-
2475 port success has a strong seasonal cycle related to the structure of the alongshore currents
2476 and their characteristic variability, including the wind-driven Ekman transport.
2477 It was found that transport success peaks in spring (September-November) when there is
2478 an efficient fast inshore pathway taken by particles reaching St Helena Bay within 15 days.
2479 This period is the optimal window, as the jet is in its intensified state, closer to the shore,
2480 and relatively stable (coherent). At this time of the year, offshore Ekman transport is still
2481 low enough to allow particles to remain on the inshore side of the jet.
2482 However in summer, during the peak upwelling season, the jet veers offshore which is
2483 attributable to the intensification and the broadening of the slope-edge poleward under-
2484 current induced by negative wind stress curl. When this happens, particles follow the
2485 offshore route where they experience an increased dispersion with less successful rate due
2486 to offshore increasing EKE values. Also, there is a decrease in transport success of inshore
2487 particles as they are advected offshore by Ekman transport. Particles taking the offshore
2488 route were found to take longer (20-30 days) to reach St Helena Bay.
2489 These results highlight some of the physical processes essential for the dispersal of fish
2490 eggs and larvae in the SBUS. For favourable transport between the spawning and nursery
2491 grounds, fish eggs and larvae have to be entrained within the fast inshore route taken
2492 by the Benguela jet and avoid being carried offshore into the turbulent region when the
2493 jet splits into several branches. The latter was also found to be linked to wind induced
2494 offshore Ekman transport.

2495

2496 In **Chapter 5**, the sensitivity of physical transport processes to increasing model
2497 resolution is discussed by nesting a 3-km resolution child grid (R3km) within the 7-km
2498 parent grid (R7km). The new simulation was run for 22 years from 1989 to 2011 with
2499 3 years considered a spin-up time. For consistency, the simulation was also forced with
2500 6-hourly CFSR reanalysis surface forcing, and the STRM bathymetry was interpolated on
2501 the R3km grid. The high resolution was used to study the impact of increased resolution
2502 on the fine scale shelf dynamical features and their effect on transport success. There is
2503 some impact on the dispersion of particles associated with the differences in mesoscale

2504 variability and the poleward coastal current. However, there is not much change in the
2505 seasonal cycle of transport success because the Benguela Jet which is the a primary driver
2506 of the alongshore connectivity is always present in both simulations.

2507 It was found that increasing the resolution results in topographically controlled features
2508 specifically over the shelf region. With the R3km, the Benguela jet is confined between
2509 200 m and 500 m isobaths, whereas the lower resolution R7km grid resulted in a broader
2510 offshore jet particularly in summer. An inner shelf (<200 m isobath) poleward flow emerges
2511 that is most intense in autumn in the R3km, and this poleward flow was found to be re-
2512 sponsible for the southward transport of particles located on the shelf. Consequently, the
2513 strongest southward transport occurred in autumn. A shelf-edge constrained jet in R3km
2514 was found to be beneficial for transport of inshore particles (originating <500 m isobath)
2515 during spring and summer, whereas the intensification of the poleward flow reduced in-
2516 shore transport success in autumn and winter.

2517 Surprisingly, there was also an unexpected impact on the quantity of warm Agulhas Cur-
2518 rent leakage into the SBUS by the increase of model resolution. There is a reduction
2519 in the Agulhas leakage in the higher resolution model. As a consequence, the stratifica-
2520 tion and the cross-shore density gradient were impacted. There were differences in the
2521 intensity of the jet related to the quantity of warm Agulhas water influx in two simula-
2522 tions. Topographical smoothing in the lower resolution R7km simulation led to an Agulhas
2523 retroreflection that occurs more westward with more leakage of warm and salty Agulhas wa-
2524 ters into the south Atlantic Ocean. Moreover, without the topographical blockage of the
2525 shallow sea-mounts located south-east of the Agulhas Bank, most of the leakage estimated
2526 at 18.3 Sv entered close to the shelf-edge of SBUS extending north to Cape Columbine in
2527 R7km. This enhanced the cross-shelf temperature gradient driving a more intense jet in
2528 the R7km simulation. However, increasing the resolution reduced this leakage by 2 Sv to
2529 16.3 Sv due to a better representation of topography resulting in a weaker jet and particles
2530 taking longer to reach St Helena Bay in the R3km. Notably, offshore particles (originating
2531 >500 m isobath) had a better probability of transport success in the higher resolution due
2532 to a reduction of EKE along the shelf-edge. On the other hand, offshore particles in R7km
2533 are exposed to higher EKE and tend to be largely dispersed into the open ocean with low

2534 probability of successful transport. However, although the Agulhas leakage was different
2535 in the two simulation, this did not have much impact on the seasonal cycle of alongshore
2536 connectivity as it is mostly driven by the Benguela jet present in both simulations.

2537

2538 Having identified the physical drivers (*i.e.* the Benguela jet, Ekman transport, pole-
2539 ward current and shelf-edge mesoscale turbulence) of the seasonal cycle of advective path-
2540 ways and transport success, in **Chapter 6** we explore interannual variability of these
2541 drivers on the interannual variability of transport success. The interannual variability of
2542 transport success is investigated using Lagrangian experiments with particles being ad-
2543 vected by velocity fields from the R3km simulation described in previous **Chapter 5**.
2544 This study found that in addition to the Benguela Jet, mesoscale eddies also play a role
2545 in the interannual variability of transport success.

2546 It was found that there is an interaction between competing processes that impact trans-
2547 port success variability. However, we find that transport success is predominately influ-
2548 enced by the Benguela jet variability such that particles must either remain over the shelf
2549 in the jet or are advected towards the shelf into it. In an event that particles are removed
2550 away from the shelf and from the influence of the jet either by offshore Ekman transport
2551 or by the poleward current, this can lead to a reduced contribution of inshore particles to
2552 transport success.

2553 The results also showed that transport success of offshore particles can contribute signif-
2554 icantly to total transport success anomalies especially strong positive anomalies. Never-
2555 theless, their contribution is dependent on cross-shore displacement and mesoscale eddies
2556 are able to impact on the cross-shore distance that particles are able to reach through
2557 their ability to drive a cross-isobath flow. The presence of an anti-cyclonic eddy either
2558 nearshore, south or offshore of the SARP line leads to an enhanced offshore transport of
2559 particles and negative anomalies of offshore transport success. However, the extent of their
2560 impact is related to their close proximity to the SARP line whereas, cyclonic eddies located
2561 between Cape Point and Cape Columbine can favor onshore transport leading to positive
2562 transport success. Moreover, the results suggest that the size of the eddy structures is also
2563 important. Circulations associated with large eddies such as an Agulhus ring can drive

2564 a more pronounced offshore long-distance advection of particles than small eddies. This
2565 is consistent with the effect of shelf-cyclonic eddies, where particles can undergo a small
2566 loop offshore and back inshore.

2567 **7.1 Key Findings**

2568 The key findings of Chapter 4, 5 and 6 are summarised by addressing the scientific ques-
2569 tions and their conclusions.

2570 **Chapter 4: What are the spatio-temporal variability and the drivers of the** 2571 **alongshore connectivity between Cape Point and Cape Columbine?**

2572 Two clear main Lagrangian pathways are identified. The results show that there is a very
2573 efficient inshore route that is characterised by a short connection time (<15 days). This
2574 route is mostly efficient during spring and facilitated by a nearshore and coherent (one
2575 branch) equatorward along-shelf jet. Weaker upwelling enables particles to remain over the
2576 shelf and within the jet. An offshore route is predominant during peak and late upwelling
2577 season when the jet splits into multiple branches linked to the poleward undercurrent
2578 intensification. There is increased dispersion in the offshore region with less successful
2579 rate due to offshore increasing EKE values, and particles take longer (20-30 days) to reach
2580 St Helena Bay. The successful transport from the Agulhas Bank to St Helena is attributed
2581 to wind induced modulation of the jet including the full three dimensional structure of
2582 the oceanic circulation and to its inherent turbulent nature.

2583 **Chapter 5: What is the influence of the ocean model spatial resolution on the** 2584 **shelf circulation features of the SBUS shelf?**

2585 Increasing the model resolution led to a better representation of the bathymetry, in partic-
2586 ular in regions with steeper slopes and also in the shallow inner shelf region, which in turn
2587 modified the cross-shore positions and intensities of the jet including the emergence of an
2588 inner shelf poleward flow. Furthermore, the higher resolution configuration allows repro-
2589 duction of a realistic pathway of the Agulhas Rings and thus of the Agulhas leakage into

2590 the South Atlantic. This is particularly true for the central route which is defined as eddies
2591 that flow between the Agulhas Ridge and the Schmitt-Ott Seamount and Erica Seamount
2592 (*Dencausse et al.*, 2010b). There is a reduced leakage of warm Agulhas water influx into
2593 the southern Atlantic as the leakage is blocked by these bathymetric features, which are
2594 smoothed out resolved in the coarser resolution. This has an impact on the intensity of
2595 the Benguela jet, and as such, on the absolute values of transport success. However, the
2596 seasonal cycle of transport success and the physical processes responsible remain the same.
2597 The equatorward jet that is narrower and located between the 200 m and 500 m isobaths
2598 in the higher resolution R3km simulation is favourable for transport success of inshore par-
2599 ticles in spring and summer. Moreover, the intensity of the Benguela jet is weaker due to
2600 the reduced Agulhas leakage and thus the EKE associated with the baroclinic/barotropic
2601 instabilities is much lower. This led to the offshore particles experiencing less dispersion
2602 and having a higher probability of transport success which contributed significantly to the
2603 total transport success.

2604 **Chapter 6: How are anomalous transport events related to the local wind**
2605 **forcing and the turbulent eddy field?**

2606 Interestingly, the variability of wind had little influence on the interannual variation of
2607 transport success, in contrast to previous studies. This is because offshore transport due
2608 to strengthened upwelling could be prevented by a strong Benguela Jet. On the other hand
2609 mesoscale eddies have an influence on the cross-shore displacement of offshore particles.
2610 The interaction of large anti-cyclonic such as Agulhas ring with the shelf-edge at the SARP
2611 line could lead to an offshore advection of particles originating from the outer shelf into a
2612 region of high turbulence. This offshore advection can contribute to negative anomalies of
2613 transport success. However, in the case where these particles don't get advected offshore
2614 away from the shelf-edge, shelf-cyclonic eddies can advect them inshore into the Benguela
2615 Jet.

2616 7.2 Future work

2617 The findings obtained from this work provide us with a better understanding of Lagrangian
2618 transport dynamics and their association with the seasonal and interannual variability
2619 of ocean circulation. The high resolution model configuration of the SBUS was found
2620 to reproduce the main characteristics of the currents in the region and their long-term
2621 fluctuations. There are several ways in which these results can be extended or applied:

- 2622 • Retention of larvae in the area of high productive area is one the key factors pro-
2623 posed by Bakun’s fundamental triad processes (*Bakun, 1996*) that determines the
2624 variability in the stock of small pelagic fishes. In this study, the retention of par-
2625 ticles in the target area after their arrival was not considered. Since the particles
2626 have to remain in the target area for specified time to be considered as successful
2627 recruits, it is important to also study cross-shore transport and retention processes
2628 in St Helena Bay. The consideration of retention time is expected to have an impact
2629 on the present transport success probabilities.
- 2630 • The occurrence of inner shelf poleward flows is believed to be very important in
2631 the SBUS for the generation, propagation and dissipation of harmful algal blooms
2632 in St Helena Bay and on the Namaqua shelf (*Pitcher and Nelson, 2006*). Thus
2633 the high resolution model (R3km) of the SBUS shelf region could provide a tempo-
2634 rally and spatially cohesive dataset for other investigations of the small-scale and
2635 high frequency dynamics that contribute to often catastrophic harmful algal bloom
2636 events. The R3km resolution can also be used indirectly to provide lateral boundary
2637 conditions to have a more higher resolution nested model (i.e. 1 km).
- 2638 • Several other processes, that were not investigated in this thesis, may lead to in-
2639 traseasonal variability of the inner shelf circulation. These processes include sub-
2640 mesoscale eddy dynamics generated by frontogenesis events in the mesoscale eddy
2641 field of the coastal transition zone (*Capet et al., 2008b*), the role of inertial waves
2642 generated by the diurnal cycle of the wind forcing (*Fawcett et al., 2008*), Stokes
2643 drift transport, and bolus transport from internal tides (*Spingys, 2017*). The former
2644 enhances the presence of high vertical velocities, which may impact the structure

2645 of primary production (*Gruber et al.*, 2011). These events occur on a time scale of
2646 a few days and on spatial scales of a few kilometres. Unfortunately, the available
2647 computing resources prevented the study of such fine scales. These scales could be
2648 responsible for large amplitude pycnocline fluctuations (*Orlić and Pasarić*, 2011),
2649 enhanced diapycnal mixing, and trigger the occurrence of harmful algal blooms (*Lu-*
2650 *cas et al.*, 2014). However, these processes are challenging to model in a realistic
2651 configuration of the southern Benguela shelf, and would require a spatial and time
2652 resolution that is beyond our available computing resources.

2653 • Mesoscale eddies, particularly anti-cyclonic eddies located near the shelf were found
2654 to have an impact on the distribution of particles by advecting them offshore. How-
2655 ever, a field survey can be carried out to quantify the role of these eddies in the
2656 cross-shelf transport of biota. Comparing larval fish and plankton biomass entrained
2657 in a cyclonic and anti-cyclonic eddy that interacted with the shelf can help us to
2658 contrast and understand the ecological consequences associated with these features.
2659 Finding juveniles in an eddy that has travelled farther in the open ocean could also
2660 determine whether eddies can maintain the survival of late-stage larvae.

2661 • Available satellite altimetry data can be used to detect eddies in the region going
2662 back in time to determine if anomalous fish recruitment events can be linked to the
2663 effect of mesoscale eddies transport.

2664 • Our findings suggest that particles on the SARP line off Cape Point are likely to be
2665 more affected by the turbulent mesoscale eddy field than particles located further
2666 north. Thus, the SARP line may not be the best transect to survey larvae in order
2667 to estimate their subsequent recruitment. This possibility needs to be investigated
2668 by carrying out a sensitivity study releasing particles at different offshore transects
2669 all along the northward route.

2670 • Our model set-up considers a monthly climatology at lateral boundaries which filters
2671 out the interannual signal such as equatorial coastal Kelvin waves generated by
2672 Atlantic Ninã and Benguela Ninõ. Some assessment of the relative roles of local
2673 forcing vs forcing at the lateral boundaries should be considered in future efforts

2674 that studies the interannual variability in physical transport processes.

2675 • Finally, the Benguela Jet has been suggested as to play a key role in carrying fish eggs
2676 and larvae from their food-poor Agulhas Bank spawning grounds to the nutrient-
2677 rich inshore nurseries on the west coast (*Hutchings et al.*, 1998). Our results provide
2678 robust evidence that this jet does indeed modulate alongshore transport success at
2679 both seasonal and interannual time-scales. Due to its importance in recruitment of
2680 small pelagic fishes, it is therefore important to establish a long-term monitoring (a
2681 mooring array) of the jet. This will provide the ocean modelling community with
2682 temporally coherent data that can be used to evaluate the simulated characteristics
2683 of the jet and to track long-term trends associated with it.

2684 Appendix A

2685 Abbreviations used for the 2686 description of the ROMS model

- 2687 • x, y, z - are the zonal, meridional and vertical directions (m) in the Cartesian coordinate
- 2688 system.
- 2689 • u, v, w - are the components of the vector velocity \vec{v} (m s^{-1}) in x, y, z respectively.
- 2690 • t - is the time in seconds (s).
- 2691 • f - is the Coriolis parameter (s^{-1}).
- 2692 • ϕ - is the dynamic pressure ($\text{m}^2 \text{s}^{-2}$).
- 2693 • D_u, D_v, D_T, D_S - are the dissipative terms.
- 2694 • T - is the potential temperature ($^{\circ}\text{C}$).
- 2695 • S - is the salinity (psu)
- 2696 • \vec{g} - is the acceleration of gravity (m s^{-2}).
- 2697 • ρ - is the seawater density (kg m^{-3}).
- 2698 • ρ_o - is the mean seawater density (kg m^{-3}).
- 2699 • P - is the total pressure (N m^{-2} or Kg m^{-2}).
- 2700 • ζ - is the sea surface variation (m).
- 2701 • τ_s^x, τ_s^y - are the surface wind stress (N m^{-2}), in x and y directions respectively.
- 2702 • τ_b^x, τ_b^y - are the bottom wind stress (N m^{-2}), in x and y directions respectively.
- 2703 • Q_{net} - is the surface heat flux (W m^{-2}).
- 2704 • K_{Mv} - is the coefficient of vertical eddy viscosity ($\text{m}^2 \text{s}^{-1}$).

- 2705 • K_{Tv} , K_{Sv} - are coefficients of vertical eddy diffusivity ($\text{m}^2 \text{s}^{-1}$).
- 2706 • K_{Mh} - is the coefficient of horizontal eddy viscosity ($\text{m}^2 \text{s}^{-1}$).
- 2707 • K_{Th} , K_{Sh} - are coefficients of horizontal eddy diffusivity ($\text{m}^2 \text{s}^{-1}$).
- 2708 • E_r - is the evaporation rate (cm day^{-1}).
- 2709 • P_r - is the precipitation rate (cm day^{-1}).
- 2710 • C_p - is the heat capacity of seawater ($\text{J kg}^{-1} \text{ }^\circ\text{C}$).
- 2711 • H - is the total ocean depth (m).
- 2712 • u_{10} , v_{10} - are the horizontal components of the wind vector in x and y respectively,
- 2713 measured at 10 m above sea level.
- 2714 • $C_{D \text{ surface}}$ is a non-dimensional wind-drag coefficient.
- 2715 • $C_{D \text{ bottom}}$ is a non-dimensional bottom-drag coefficient.
- 2716 • ρ_{air} - is the density of atmospheric air (kg m^{-3}).
- 2717 • r - is the coefficient of linear bottom friction.
- 2718 • γ_2 - is the coefficient of quadratic bottom friction.
- 2719

2720 Appendix B

2721 Eddy detecting and tracking 2722 algorithm

2723 An automatic eddy tracking algorithm developed by *Penven et al.* (2005) in Matlab is
2724 used to detect and study mesoscale eddy structures in the SBUS. Generally, there are
2725 two methods used to detect eddies in the ocean based on geometric criteria (*Chelton*
2726 *et al.*, 2011) and the dynamical method (*Isern-Fontanet et al.*, 2006; *Chelton et al.*, 2007)
2727 using local minima SSH, maxima SSH and Okubo-Weiss. The geometric method typically
2728 applied to altimetry data (*Halo et al.*, 2013) is based on detecting closed loops by first
2729 identifying local minima and maxima SSH fields as eddy centre points. Eddy boundaries
2730 corresponds to the largest closed loop SSH contour, however, a maximum eddy length-
2731 scale threshold is defined to exclude large close loops structures such as gyres (*Chelton*
2732 *et al.*, 2011).

2733 On the other hand, the dynamical method is based on the local dynamical properties of the
2734 flow which compares relative vorticity to strain (deformation); where a negative Okubo-
2735 Weiss ($W < 0$) parameter indicates that relative vorticity dominates over strain and this
2736 is true for the inner region of an eddy (*Isern-Fontanet et al.*, 2006; *Chelton et al.*, 2007).
2737 The Okubo-Weiss parameter which measures the relative importance of strain tensor and
2738 vorticity (*Okubo*, 1970; *Weiss*, 1991) (Equation B) has to be assessed against a threshold
2739 value.

$$W = S_n^2 + S_s^2 - \xi^2 \quad (\text{B.1})$$

2740

2741 where the components are:

$$S_n^2 = \frac{\partial u}{\partial x} - \frac{\partial v}{\partial y}, \quad S_s^2 = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}, \quad \xi^2 = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \quad (\text{B.2})$$

2742

2743 where S_n (S_s) is the normal (shear) strain component of the strain tensor; ξ is the vertical
2744 component of relative vorticity; u and v are respective zonal and meridional components
2745 of geostrophic velocity computed from SSH fields.

2746 In our analysis, the combination of two methods is used, such that an eddy feature is
2747 considered as a feature that is characterised by a closed loops of SSH and a negative Okubo-
2748 Weiss parameter in our ocean model. The detailed description of this combined method is
2749 found in *Halo et al.* (2013, 2014) and has been proven to be more robust in consistent eddy
2750 detection. Firstly, using the Okubo-Weiss parameter, only regions dominated by vorticity
2751 (i.e. negative W) are considered. To avoid the detection of basin-scale gyres, detection
2752 of eddies is limited to a radius of less than 400 km. Next, the inner regions are detected
2753 by using SSH contour lines, and a 2 cm contour interval was used. Although, we use the
2754 same algorithm as *Halo et al.* (2013, 2014), some of the tunable parameters such as the
2755 maximum size of a closed loop (Halo = 600 km, chosen = 400 km), the number of the
2756 Hanning filter on W was set as zero (Halo = 2) in our analysis.

2757 After an eddy structure is detected, it is tracked using the methodology of *Penven et al.*
2758 (2005), where changes are tracked over time. An eddy is detected in the first frame and
2759 is considered the same eddy if there is a minimum generalised change of some of their
2760 properties in the subsequent frames:

$$W_{e_1, e_2} = \sqrt{\left(\frac{\Delta X}{X_0}\right)^2 + \left(\frac{\Delta R}{R_0}\right)^2 + \left(\frac{\Delta \xi}{\xi_0}\right)^2} \quad (\text{B.3})$$

2761 Where

2762

2763 • ΔX is the spatial distance between two eddy centres e_1 and e_2

2764

2765 • ΔR is the variation in diameter between e_1 and e_2

2766

2767 • $\Delta\xi$ is the variation in relative vorticity between e_1 and e_2

2768

2769 • X_0 is the typical eddy length scale: 30 km

2770

2771 • R_0 is a typical eddy radius scale: 30 km

2772

2773 • ξ_0 is a typical eddy vorticity scale: 10^{-5} s^{-1}

2774

2775 In addition, for an eddy to be tracked as the same eddy between frames, it must retain its
2776 polarity. In the case where this condition is violated, $W_{e_1, e_2} = \infty$. A unique identifier is
2777 assigned to all eddies tracked which the eddy retains through its lifetime. Where an eddy
2778 splits, a new identifiers are assigned for each child eddy. The track gets terminated if no
2779 subsequent eddy is found.

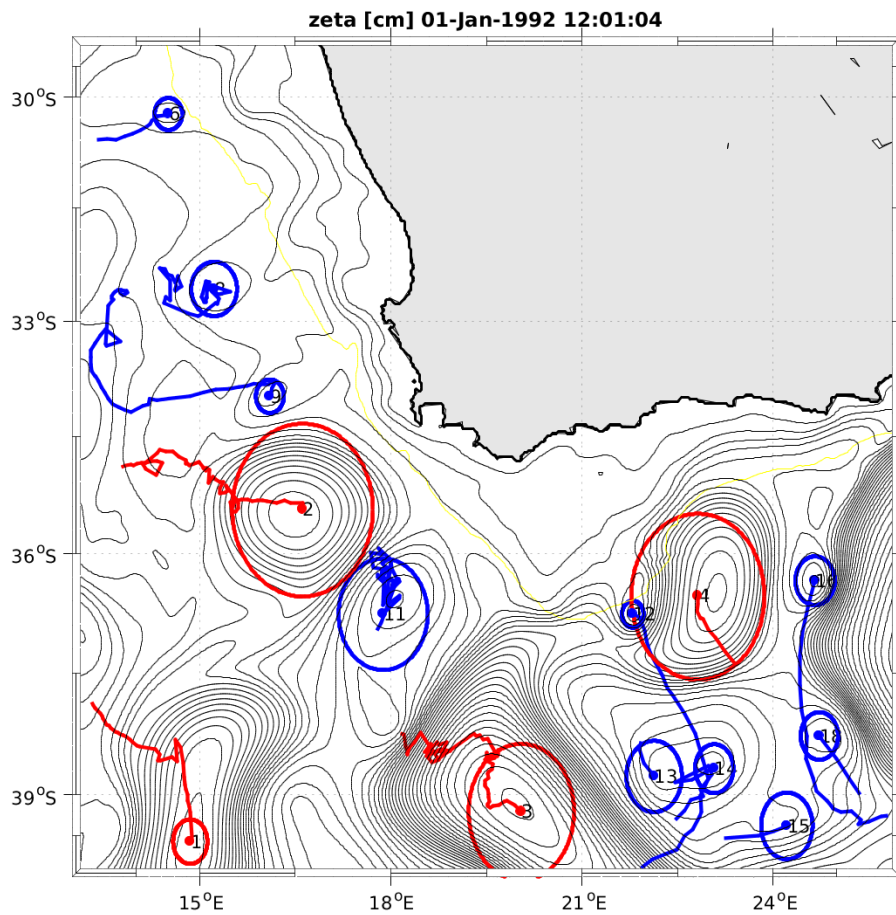


Figure B.1: Snapshot (01 January 1992) of eddies detected by the eddy detection algorithm in the R3km SSH field, contoured at 5 cm. Red (blue) circles indicate anti-cyclones (cyclones) and their trajectories throughout their lifetime are shown by red and blues lines, respectively. All eddies are tracked irrespective of their lifetime and the number shown on each eddy represents their identifiers (ID numbers). The yellow line traces the 500 m isobath.

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Titre : La variabilité du transport lagrangien dans la partie sud du système d'upwelling du courant du Benguela.

Mots clés : succès du transport, le Benguela jet, les tourbillons à méso-échelle.

Résumé : Cette étude a pour cadre le système d'upwelling côtier du Benguela sud (SUBS) dans lequel l'advection par les courants marins influe sur de nombreux processus biologiques. Le rôle de ces courants est exacerbé par la séparation géographique existant entre les zones de ponte et de nurserie de plusieurs populations de poissons marins. Des simulations numériques basées sur un modèle hydrodynamique du SUBS sont exploitées à l'aide d'outils lagrangiens pour identifier les trajectoires suivies par des particules entre la péninsule du Cap (34°S) et la baie de Sainte-Hélène (32°S) et pour explorer les liens entre ces trajectoires et la circulation océanique.

Le rapport entre le nombre de particules atteignant la baie de Sainte-Hélène et le nombre total de particules relâchées est utilisé pour quantifier la connectivité entre les deux régions.

Plusieurs processus physiques importants sont identifiés : (i) le Benguela jet, (ii) le transport d'Ekman vers le large, (iii) le courant s'écoulant vers le sud dans la bordure littorale et (iv) les tourbillons à méso-échelle.

Le Benguela jet est le mécanisme dominant la connectivité aux échelles de temps saisonnières et interannuelles. Le rôle contrasté des tourbillons océaniques suivant leur localisation et leurs caractéristiques a également été documenté.

L'identification de ces processus physiques et la quantification de leurs impacts sur la connectivité entre les zones de ponte et de nurserie dans le SUBS sont des résultats qui vont indéniablement contribuer dans de futurs travaux à cerner le rôle des processus d'advection sur la survie larvaire et sur la dynamique des populations marines dans cet écosystème.

Title : The variability of Lagrangian transport in the southern Benguela Current upwelling system.

Keywords : Transport success, Lagrangian pathways, variability, Benguela Jet, mesoscale eddies.

Abstract : This study analyses the physical mechanisms that impact Lagrangian pathways and transport in the southern Benguela upwelling system (SBUS), an environment in which currents are key components of many important ecological processes, including the dispersal of marine larvae. Physical advection by currents is an important mechanism for egg and larvae transport success in the SBUS since the spawning areas and recruitment areas are separated by a long distance. Numerical model simulations of the SBUS coupled with particle tracking experiments are used to investigate Lagrangian pathways between the Cape Peninsula (34°S) and St Helena Bay (32°S) and how they are linked to the oceanic circulation. Transport success, given by the ratio of the number of particles that reach St Helena Bay over the total number of particles released, is used to quantify the alongshore connectivity between the two regions.

We have identified and quantified the following physical drivers: (i) Benguela Jet, (ii) offshore Ekman transport, (iii) inner shelf poleward current, (iv) mesoscale eddies to be responsible for the spatial and temporal variability of the alongshore connectivity. The Benguela Jet was found to be the dominant driver of the connectivity at both seasonal and interannual timescales. Moreover, the presence of anti-cyclonic eddies near the shelf-edge negatively impact transport success by advecting particles into the open ocean. The opposite occurs with shelf-edge eddies as they transport particles onshore onto the shelf and the Benguela Jet. These findings will provide a valuable information for the future studies on the role of the physical drivers that impact transport of larvae and eggs.