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Cyclonic Eddies in the Cape Basin

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Declaration

I, Candice Hall, declare that contents of this thesis represent my own work. I declare that the opinions contained within are my own, and not necessarily those of the University of Cape Town. Equally, this thesis has not been previously submitted for academic examination towards any qualification.
Acknowledgements

Firstly I would like to thank Prof. Johann R.E. Lutjeharms (University of Cape Town) for his unlimited assistance with the concept and design of my proposal, as well as guidance in completion of this thesis. Without him this study would not have been accomplished.

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Abstract

A great deal of attention has been paid to the inter-ocean exchange of thermohaline properties in the Agulhas Retroflection region. Recent observations have shown that the highly energetic field of the southern half of the Cape Basin consists of both cyclonic and anticyclonic eddies. These eddies interact with each other, resulting in the vigorous stirring of these water mass. Investigations have shown that the cyclonic eddies tend to be smaller and outnumber the anticyclonic rings. Nonetheless, very little is known of their characteristics and the nature in which they are formed. Using remote sensing data, confirmed with hydrographic data, this study determines the location, frequency and seasonality of cyclonic eddy formation; their size, trajectories and lifespan; physical components and associations with Agulhas Rings.

Cyclonic eddies were seen to split, merge and link with other cyclonic eddies, with splitting events creating child cyclonic eddies. The 105 parent and 157 child cyclonic eddies identified during this study show that an average of 11 parent and 17 child cyclonic eddies were formed annually. 31.58 % follow an overall west-southwest direction, with 27.37 % translocating west north-westward. Poleward translocation speeds average at 0.3 km/day, whereas translocation speeds obtained from all directional components averages at 2.153 km/day for parent and 2.975 km/day for child cyclonic eddies. Parent cyclonic eddies lived for approximately 254 days, whereas child cyclonic eddies survived for a mean of 188 days. Of note was a significant variation of lifespan between parent and child cyclonic eddies formed in both the north and south of the study area. 77 % of northern and 93 % of southern cyclonic eddies were formed directly adjacent to positive sea level anomalies or Agulhas Rings, resulting in an total overall association of 82.93 % parent and 89.63 % child cyclonic eddies. Cyclonic eddy groups were seen to merge at a rate of 16.38 parent and 14 child cyclonic eddies per year, whereas topography appeared to affect the demise of 17.00 % of the investigated cyclonic eddies.

Therefore this study may form a basis for further investigations into the influence Cape Basin cyclonic eddies have on the meridional transfer of heat, salt, nutrients, oxygen and carbon concentrations in the South-East Atlantic Ocean. A more in-depth study using model outputs and targeted in situ hydrographical data would again enhance cyclonic eddy knowledge.
CHAPTER 1

Introduction

The oceans of the world play an important role in the thermoregulation of our planet. Evaporation from the surface waters at low latitudes increases the salinity of the upper layers of the tropical oceans. Currents carry these warm, salty waters towards the poles where they are cooled by exposure to the colder high latitudes. The processes of evaporation and pack ice formation found near the poles increase the salinity of the surface waters. As density is a function of salinity, these cooler, more saline waters sink to form waters that return, at depth, to the low latitudes. Therefore heat is transferred from the warmer tropics to the colder poles. This north-south movement of heat is known as the meridional heat flux.

![Figure 1.1: The great ocean conveyor belt (Broecker, 1987).](image)

As there is, in essence, only one major ocean punctuated by continents, waters of differing temperature and salinity may be found circulating throughout the interlinking ocean basins, regulating the earth’s heat budget. The process of this thermohaline transfer around the globe, also known as the oceanic conveyor belt shown in the simplistic figure 1.1 (Broecker, 1987), plays a vital part in the maintenance of our current world climate (Broecker, 1991).
A major indicator of ocean circulation is the volume and signature properties of water masses transported between the ocean basins. The reasons for this are twofold. Firstly, the signature properties act as identification agents or tracers for the source of the water masses. Secondly, it is climatically important to know the volume of heat and salt transfer between the ocean basins, as these variables drive the mixing and overturning circulation of the oceans (de Ruijter et al., 1999) and as such, influence the global climate.

In order to predict future global climate change, much attention has been focused on thermoregulation and circulation within the ocean. In a collaboration between approximately 30 nations, the large-scale 1990 – 2002 World Ocean Circulation Experiment (WOCE) used altimetry, physical and chemical hydrography, as well as numerical ocean models to determine heat flux and transport in four of the world’s oceans. The success of the WOCE has lead to the implementation of the following programs: CLIVAR, a global study of ocean climate variability and predictability; GODAE, the Global Ocean Data Assimilation Experiment; and ARGO, a global array of temperature/salinity profiling floats (WOCE, 2005). In this way a plausible attempt is being made to monitor and study the global overturning circulation of the ocean.

Two main forces drive ocean circulation. Wind affects the surface layers but its influence does not penetrate deeper than 1000 meters. Water renewal deeper than that is achieved by thermohaline circulation; simply, currents that are driven by density differences produced by temperature (thermal) and salinity (haline) effects. Thermohaline circulation, in turn, is forced by the formation of water masses, which are created by surface processes in specific locations, such as polar cooling and tropical evaporation. Temperature and salinity are properties that are relatively conservative when not affected by processes in the upper mixed layer (~50 – 150 m) of the ocean, such as evaporation, precipitation or cooling. Therefore the temperature-salinity (T-S) combinations of water masses at depth act as tracers when following water mass movement, allowing for the slow at-depth thermohaline changes due to advection and mixing. The ability of a T-S relationships acting as a tracer for a particular water masses is eloquently depicted in figure 1.2. From this picture it is clearly possible to see the vast difference between, for example, Mediterranean Water and Antarctic Intermediate Water. Hence thermohaline properties play an important role in our monitoring of the ocean.
Deep convection in the Weddell and Ross Seas forms Antarctic Bottom Water (AABW), while most Antarctic Intermediate Water (AAIW) is formed by deep convection south of South America and spreads into all the oceans with the Antarctic Circumpolar Current. Evaporation produces Mediterranean (MedW) and Red Sea (RedSW) waters, which are intrusions of high temperature, high salinity waters from those two seas.

Figure 1.2: T-S plot of water in the Atlantic Ocean basin (modified from Polzin, 1983).

Figure 1.3: Contour plot of salinity as a function of depth in the western basins of the Atlantic from the Arctic Ocean to Antarctica. The plot clearly shows extensive cores, one at depths near 1000 m extending from 50°N to 20°S; the other at near 2000m extending from 20°N to 50°S. The upper is the Antarctic Intermediate Water, the lower is the North Atlantic Deep Water. The arrows mark the assumed direction of the flow in the cores. The Antarctic Bottom Water fills the deepest levels from 50°S to 30°S (from Lynn and Reid, 1968).
Importantly for Atlantic circulation, deep convection in the Arctic Ocean, the Greenland Sea and the Labrador Sea, forms North Atlantic Deep Water (NADW), the major water mass found in the Atlantic Ocean. The contour plot in figure 1.3 shows the relationship between depth and salinity in the Atlantic Ocean, highlighting the prominent water masses found there. The upper core represents AAIW (~ 1000 m); below that NADW (~ 2000 m) and AABW may be found along the ocean floor. Of interest is the NADW, as the formation of this water mass is thought by many to be an essential key that drives the Atlantic Ocean circulation, therefore forcing global ocean circulation (Broecker, 1991).

Dr Wallace Broecker, the creator of the previously mentioned ‘conveyor belt’ theory (Broecker, 1987), has emphasized the importance of inter-ocean exchange of heat and salt. He states that ‘the ocean’s conveyor appears to be driven by the salt left behind as the result of water-vapour transport through the atmosphere from the Atlantic to the Pacific basin’ (Broecker, 1991). The Atlantic is the only ocean where net heat is transported from the Southern to the Northern hemisphere. In other words, heat is transported across the Equator, as opposed to strictly polewards from the equator as in the other oceans (de Ruijter et al., 1999). Within the Indian and Pacific Oceans heat transport across the equator is approximately zero.

This heat transfer allows surface waters of the Atlantic Ocean to be exposed to evaporation processes for longer than water in the adjoining basins, as it travels towards the North Pole. Hence the Atlantic Ocean surface waters are found to be approximately 1-3 g/litre higher in salt content than the Pacific (Broecker, 1991), which is conducive to cooling and subsequent sinking as mentioned above. As seen in figure 1.1 (page 7), the more dense surface waters of the Greenland and Labrador Seas, upon cooling, are able to sink to greater depths than on the same latitudes within the Pacific. The resulting formation of North Atlantic Deep Water (NADW) is a major contributor to the ocean conveyor circulation as the area is one of the few places in the world where such waters may be formed (Broecker, 1997). This newly formed NADW spreads to fill most of the Atlantic Ocean and is ultimately a key component of global ocean circulation.
Hence the physical and chemical properties of the waters that flow into the Atlantic Ocean are crucial with regard to NADW formation (Gordon et al., 1992) and consequently global ocean circulation, as seen in figure 1.1 (page 7). For example, should the salinity of the water decrease significantly prior to the cooling process, the formation of NADW may be reduced sufficiently to stop Arctic deep convection and therefore the ocean conveyor belt. This cessation would have catastrophic effects on the global climate, especially with regards to maintaining the relatively mild (non ice-age) winters that the northern reaches of the globe, such as Europe, currently experience (Broecker, 1991).

Figure 1.4, taken from Ganachaud and Wunsch (2003), provides an overview of many studies conducted on horizontal Atlantic Ocean heat transport. In the figure, the numbers above the arrows show quite clearly that at the 30°S ocean heat loss / heat gain junction, the heat divergences are not as gradual as they are in the Northern Hemisphere. In fact, many of the studies (Maedonald and Wunsch, 1996; Ganachaud and Wunsch, 2003; Holfort and Stieglfer, 2001) seem to indicate that there is an abnormal increase in heat loss within this area. Due to the positive northward heat transfer, this dip must have an effect on the overall heat transfer within the Atlantic.

![Figure 1.4: Heat transports across hydrographic sections. Black cones with thick error bars indicate advective horizontal heat transports across hydrographic sections—except at 7.5°N, see text—numbers above the arrows indicate the divergences, or air sea fluxes, positive red for ocean heat gain, negative blue for ocean heat loss. Heat, or energy transports are referred to 0°C. Northward heat transport is positive (Ganachaud and Wunsch, 2003).](image-url)
The meridional flux of waters in the Atlantic Ocean logically highlights the importance of the water transport into and out of the Atlantic Ocean. The area south of the African continent is thought to be one of the main inflows of water into the Atlantic (Gordon and Piola, 1983; Gordon, 1985 and 1986; Broecker, 1991). The degree of heat and salt exchange between the Indian and Atlantic oceans is deemed by many (Gordon et al., 1992; Lutjeharms, 1996; de Ruijter et al., 1999; Weijer et al., 1999; Sloyan and Rintoul, 2001) to be pivotal to the overturning circulation of the ocean.

Thompson et al. (1997) believe that water leakage from the Agulhas Current into the Atlantic is partly accountable for the large northward heat flux in the South Atlantic. Weijer et al. (1999) took this a step further to deduce that this large heat flux would result in a concurrent large atmospheric heat loss, affecting evaporation rates and therefore the net surface increase of salinity in the Southern Atlantic.

Returning to Weijer et al.'s 1999 study, they showed that 'overturning strength is very sensitive to details of the interbasin fluxes of heat and salt. This means that an impact of the Agulhas leakage on the overturning circulation is probable'. Therefore, Weijer et al. (1999) concluded that thermosaline inter-ocean fluxes, instigated by both the ocean and the atmosphere, have an effect on global ocean circulation. This has also been shown to be particularly crucial south of Africa, where the influence of the Agulhas Current dominates.

The Agulhas Current, shown in figure 1.5, is a fast, poleward flowing current that is prominent along the east coast of southern Africa and forms the western boundary current for the South Indian Ocean (Lutjeharms, 1996). The Agulhas Current has surface speeds of up to 2 m/s (Wyrtki, 1973) and a volume flux of about $70 \times 10^6 \text{ m}^3/\text{s}$ (Bryden and Beal, 2001). As depicted in figure 1.5, once the Agulhas Current rounds the tip of the African continent and loses the stabilizing effect of the continental shelf, it retroreflects sharply towards the east, becoming the Agulhas Return Current (Lutjeharms and Ansorge, 2001). This water flows eastwards, eventually becoming the South Indian Ocean Current (Lutjeharms, 1996), which forms the southern boundary of the South Indian Ocean gyre.
Of interest to inter-ocean exchange of water, however, is the region where the Agulhas Current retroreflects. Recent altimetric (van Leeuwen et al., 2000) and in situ studies (Lutjeharms et al., 2003) show that the Natal Pulse (Lutjeharms and Roberts, 1988) formed at the Natal Bight (about 31°S) appears to cause a perturbation in the Agulhas Current that may have an important impact in the region of the Agulhas Retroreflection (Lutjeharms, 1996; de Ruijter et al., 1999; van Leeuwen et al., 2000), as shown in figure 1.5. They believe that this meander causes a loop of warm, salty water to be occluded from the Agulhas Current into the Atlantic Ocean in the region of the Agulhas Retroreflection (Lutjeharms, 1996). Many studies have confirmed the presence of these rings (Lutjeharms and van Ballegooijen, 1988; Duncombe Rae et al., 1992; Byrne et al., 1995; de Ruijter et al., 1999; Garzoli et al., 1999; etc.), now known as Agulhas Rings. Importantly, these rings provide a "vessel" for the transfer of warm, salty water from the Indian to the Atlantic Ocean and form a principal link between these two oceans.
Hence, in regions such as the Cape Basin (see figure 1.5), where converging water masses cause intense mixing and stirring, the thermohaline relationship is generally complex (Lutjeharms, 1996). Knowing the physical signature of the surrounding water is paramount in discerning the hydrographic difference between the background water mass and targeted eddy features. Table 1.1 (Valentine et al., 1993) shows a hydrographic summary of water masses found within the southern Agulhas. Knowledge of the hydrographic temperature and salinity data of eddies allows scientists to resolve the origin and possible evolution of the water features (Duncombe Rae et al., 1992; Lutjeharms, 1996; Garzoli et al., 1999), as the closed nature of eddies ensures that they retain many of their original signature characteristics. Once these eddies are identified, their lifespan, trajectories and effect on the surrounding region may be studied.

Table 1.1: Summary of the water masses of the southern Agulhas region (after Valentine et al., 1993)

<table>
<thead>
<tr>
<th>Water Mass</th>
<th>Temperature Range (°C)</th>
<th>Salinity Range (psu)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface Water</td>
<td>16.0 - 26.0</td>
<td>&gt; 35.50</td>
</tr>
<tr>
<td>Central Water</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(i) Southeast Atlantic Ocean</td>
<td>6.0 - 16.0</td>
<td>34.50 - 35.50</td>
</tr>
<tr>
<td>(ii) Southwest Indian Ocean</td>
<td>8.0 - 15.0</td>
<td>34.50 - 35.50</td>
</tr>
<tr>
<td>Antarctic Intermediate Water</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Characteristic T/S</td>
<td>2.2</td>
<td>33.87</td>
</tr>
<tr>
<td>(i) Southeast Atlantic</td>
<td>2.0 - 6.0</td>
<td>33.80 - 34.80</td>
</tr>
<tr>
<td>(ii) Southwest Indian</td>
<td>2.0 - 10.0</td>
<td>33.80 - 34.80</td>
</tr>
<tr>
<td>Deep Water</td>
<td></td>
<td></td>
</tr>
<tr>
<td>North Atlantic Deep Water (Southeast Atlantic)</td>
<td>1.5 - 4.0</td>
<td>34.50 - 35.00</td>
</tr>
<tr>
<td>Circumpolar Deep Water (Southwest Indian)</td>
<td>0.1 - 2.0</td>
<td>34.63 - 34.73</td>
</tr>
<tr>
<td>Antarctic Bottom Water</td>
<td>-0.9 - 1.7</td>
<td>34.64 - 34.72</td>
</tr>
</tbody>
</table>

Agulhas Rings have been observed to drift westwards as anti-cyclonic eddies (e.g. Byrne et al., 1995), away from the Agulhas Retroflection and into the South Atlantic (Lutjeharms, 1996; Richardson et al., 2003). Numerous studies and programs (Duncan, 1968; Bang, 1970; Olson and Evans, 1986; Lutjeharms and Gordon, 1987; Gordon and Haxby, 1990; Van Ballegooeyen et al., 1994; Lutjeharms, 1996; Boebel et al., 1998; Lutjeharms et al., 2000) have looked at the importance of Agulhas Rings with respect to heat, increased salinity and energy leakage from the Indian Ocean into the Atlantic.
Individual Agulhas Rings have been observed via altimetry (TOPEX/Poseidon and ERS-2 satellites) and surveyed hydrographically using expendable bathythermographs (XBT), conductivity-temperature-depth-oxygen (CTDO) profiles, lowered acoustic Doppler current profiler (LADCP) and a hull-mounted acoustic Doppler current profiler (ADCP) (e.g. Garzoli et al., 1999; van Aken et al., 2003). This gathering of data ensures a continual increase in knowledge of the process by which Agulhas Rings form and their leakage into the Atlantic.

De Ruijter et al.'s 1999 review of Indian-Atlantic inter-ocean exchange documents the energy transfer results from a number of previous studies, as shown in table 1.2. To examine one such study more closely, Byrne et al. (1995) calculated that each of their fifteen (15) studied Agulhas Rings typically transported $0.8 \text{ Sv}$ ($1 \text{ Sverdrup} = 10^6 \text{m}^3 \text{s}^{-1}$) of warm, salty Indian Ocean water into the Atlantic. They determined that the standard kinetic energy value per Agulhas Ring was $4.5 \times 10^{15} \text{ J}$ (Joules), which results in an average available potential energy value of $18 \times 10^{15} \text{ J}$. However, their study showed that the total energy entering the Atlantic via Agulhas Rings could be as high as $70 \times 10^{15} \text{ J}$, deeming them to be 'significant components of the Indian / South Atlantic interbasin exchange and of the South Atlantic meridional thermohaline flux' (Byrne et al., 1995). This proven influx of heat and salt has a direct effect on the surrounding water (Duncombe Rae et al., 1992) and may form a crucial link in the formation of NADW and hence influence ocean circulation (Gordon, 1986).

**Table 1.2: Agulhas Ring properties as found in literature (modified from de Ruijter et al., 1999)**

<table>
<thead>
<tr>
<th>Source</th>
<th>Average Potential Energy ($10^{15} \text{ J}$)</th>
<th>Kinetic Energy ($10^{15} \text{ J}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Olson and Evans (1986)</td>
<td>30.5</td>
<td>6.2</td>
</tr>
<tr>
<td>Duncombe Rae et al. (1992)</td>
<td>38.8</td>
<td>2.3</td>
</tr>
<tr>
<td>Byrne et al. (1995)</td>
<td>18</td>
<td>4.5</td>
</tr>
<tr>
<td>Clement and Gordon (1995)</td>
<td>7.0</td>
<td>7.0</td>
</tr>
<tr>
<td>Duncombe Rae et al. (1996)</td>
<td>11.3</td>
<td>2.01</td>
</tr>
<tr>
<td>Garzoli et al. (1996)</td>
<td>2.8 – 3.8</td>
<td>-</td>
</tr>
<tr>
<td>Golli et al. (1997)</td>
<td>24</td>
<td>-</td>
</tr>
</tbody>
</table>
However, these are not the only eddies to be found within this region. While analysing data from the Benguela Sources and Transports (BEST) program, Duncombe Rae et al. (1996, p. 11,958) described "a seemingly persistent phenomenon is what appears to be a 'recoil' effect in the [inverted echosounder] record; that is, the thermocline appears to shallow appreciably after passage of an [anticyclonic] eddy before relaxing to the local mean". It is possible that this 'shallow thermocline recoil' actually represents a southern hemisphere cyclonic eddy that is passing behind the identified anticyclonic ring. Data collected during the Cape of Good Hope Experiment (KAPEX) allowed Boebel et al. (2003) to confirm that floats released into the area known as the Cape Basin (see figure 1.5, page 13) alternate between cyclonic and anticyclonic motion at intermediate depths, supporting the theory that postulates the existence of cyclonic eddies within the Cape Basin.

The presence of cyclonic eddies between the anticyclonic Agulhas Rings (as shown in figure 1.5, page 13) may have a large impact on estimated potential energy, water mixing, the spin-down of Agulhas Ring etc. (Boebel et al., 2003; Richardson and Garzoli, 2003). The discovery of these cyclonic eddies denotes that the region's dynamic energy previously attributed solely to Agulhas Rings may have to be reconsidered. Interaction between the cyclonic and anticyclonic eddies could possibly result in interactive stirring and mixing. Such an action may defuse the warm, salty water of an Agulhas Ring into the background water far more quickly than was previously thought to happen by natural degradation. This theory could possibly explain how the water properties between the southern and northern arms of the South Atlantic subtropical gyre can differ to such a degree, as defined by Boebel et al. (2003). Mesoscale eddies may also conceivably have biological effects on the background waters of the Cape Basin, as new nutrients are mixed into the euphotic zone within the turbulent region.

These cyclonic eddies may furthermore affect the recruitment of commercial fish stocks that are still in their egg and larval stage. Much work has focused on the recruitment of important pelagic species such as anchovy (Engraulis capensis) and sardines (Sardinops sagax) within the Benguela Current (e.g. Boyd et al., 1992; Painting et al., 1998; Skogen et al., 2003). The southern Benguela upwelling regime is especially important as it has been envisaged as the recipient of a funnel of water from the Agulhas Bank (Boyd et al., 1992), as seen in figure 1.6.
However, Boyd et al. noticed that this 'funnelling' system seemed to have holes in it resulting in a considerable loss of eggs offshore. It could be possible that these 'holes' were actually cyclonic eddies that, due to their ability to advect massive volumes of water away from the coastline and into the nutrient-barron center of the South Atlantic gyre, contributed to the loss of these potential recruits.

![Figure 1.6: Hutchings (1992) larval transport scenario (not to scale), showing the prominent funnelling of plankton around the Cape of Good Hope.](image)

As little is known about these cyclonic eddies, it is possibly that they may be having a large impact on the physical mixing and stirring processes within the Cape Basin. These cyclones may be crucial components of the circulation in the region and it is therefore essential to establish what is already known about them.
CHAPTER 2

Knowledge of eddies in the Cape Basin

The definition of an eddy is a 'circulation system in which the water follows closed circular or elliptic paths, ... can be cyclonic or anti-cyclonic' (Tomczak and Godfrey, 2002). To date, the most definitive study of the cyclones found within the Agulhas region was conducted by Boebel et al. (2003). They compared sea-surface height data with in situ Lagrangian velocity measurements at intermediate depth gathered during the KAPEX float program around southern Africa. The aim of this project was to determine the inter-ocean exchange between the Atlantic and Indian Oceans, in particular to establish a 'zone of turbulent stirring and mixing [between the two Oceans] in the south-eastern Cape Basin' (Boebel et al., 2003).

Using RAFOS float data (figure 2.1) and MODAS-2D SSH data (figure 2.2 below), they found that most cyclonic eddies are limited to the south-eastern Cape Basin, in a region now known as the Cape Cauldron (Boebel et al., 2003). Within their three years of data (1997 - 1999) they recorded sixty-two cyclonic eddies and only twenty-nine anticyclonic Agulhas Rings. They determined that although the anticyclones were observed to drift in a north-westerly pattern at an average of $3.8 \pm 1.2$ cm/s, the cyclones drifted in a west south-westward direction at approximately $3.6 \pm 0.6$ cm/s (Boebel et al., 2003).
Boebel et al. (2003) have furthermore determined that the average forty-day lifespan of a cyclone was shorter than that of an anticyclonic ring. However, they admit that their study only identified those eddies with life spans of over two months, possibly filtering out equally effective but fleeting eddies. Once leaving the Cape Basin these cyclones may last far longer than the anticipated 2–3 months. As for size, their RAFOS data suggests that these cyclones may reach a diameter of 120 km (Boebel et al., 2003).

![Figure 2.2: Trajectories of cyclones in the Cape Couldeon as estimated from MODAS-2D SSH for the years 1997–1999. Isobaths are at 0, 3000, and 5000 m. Bold trajectories end at the triangle (Boebel et al., 2003).](image)

In addition, using Ollitrault's (1994) routine, they determined the average kinetic energy of both cyclonic and anticyclonic features, as shown in Table 2.1 (page 20). The rows represent zonal (u) and meridional (v) velocity components using both the 650 to 1150 m RAFOS float information and geostrophic MODAS surface velocity (Boebel et al., 2003). Notable within this table are the higher frequency of cyclones and in some cases, such as within the Cape Basin, their greater circular velocity than that of the anticyclones.

Due to their location formation, Boebel et al. (2003) believed that the Cape Basin cyclones contain water with properties identical to the background flow. They suggested that these cyclones may form from ‘interaction[s] of Agulhas Rings with the shelf, to the local upwelling along the East African shelf (Shannon and Nelson, 1996) or to an intermediate eastern boundary current, which is suggested to emerge out of the Angola Basin to flow polewards along the African shelf break’ (Shannon and Hunter, 1988, as quoted by Boebel et al., 2003).
Table 2.1: Average kinematic properties of cyclonic and anticyclonic features within the Cape Cauldron as detected by MODAS-2D and RAFOS floats (Boebel et al., 2003)

<table>
<thead>
<tr>
<th>Feature</th>
<th>Cyclones</th>
<th>Anticyclones</th>
</tr>
</thead>
<tbody>
<tr>
<td>MODAS-2D derived features</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Number of features</td>
<td>43</td>
<td>29</td>
</tr>
<tr>
<td>$u$ (cm s$^{-1}$)</td>
<td>$3.6 \pm 0.6$</td>
<td>$3.6 \pm 1.0$</td>
</tr>
<tr>
<td>$r$ (cm s$^{-1}$)</td>
<td>$0.4 \pm 0.5$</td>
<td>$1.3 \pm 0.7$</td>
</tr>
<tr>
<td>Ratio of no. of features</td>
<td>$\approx 3:2$</td>
<td></td>
</tr>
<tr>
<td>RAFOS trajectory segments (all floats)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Number of features</td>
<td>23</td>
<td>12</td>
</tr>
<tr>
<td>$u$ (cm s$^{-1}$)</td>
<td>$1.8 \pm 2.8$</td>
<td>$1.0 \pm 1.4$</td>
</tr>
<tr>
<td>$r$ (cm s$^{-1}$)</td>
<td>$3.0 \pm 5.8$</td>
<td>$3.0 \pm 5.0$</td>
</tr>
<tr>
<td>Ratio of no. of features</td>
<td>$\approx 2:1$</td>
<td></td>
</tr>
<tr>
<td>RAFOS trajectory segments (Cape Basin floats only)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Number of features</td>
<td>17</td>
<td>11</td>
</tr>
<tr>
<td>$u$ (cm s$^{-1}$)</td>
<td>$0.7 \pm 3.2$</td>
<td>$0.9 \pm 1.6$</td>
</tr>
<tr>
<td>$r$ (cm s$^{-1}$)</td>
<td>$3.1 \pm 2.5$</td>
<td>$2.9 \pm 5.8$</td>
</tr>
<tr>
<td>Ratio of no. of features</td>
<td>$\approx 3:2$</td>
<td></td>
</tr>
</tbody>
</table>

Boebel et al. (2003) believed that the frequency of cyclones found within their study region highlighted the importance of these previously disregarded features, with respect to eddy mixing and stirring capabilities. Secondly, they suggested that cyclonic eddies may contain sub-antarctic or even subtropical water [obviously dependent on their formation location], which, during turbulent processes, could have an impact on the background flow and subsequent heat budget. They hoped to obtain information on the water properties of such cyclones during the planned Dutch – South African Mixing of Agulhas Ring Experiment (MARE). They felt that this information, combined with a long-term inter-ocean exchange program, would allow one to interpret more accurately the heat budget between these two important oceans.

In order to investigate the meridional transfers of heat, freshwater, carbon and nutrients, Morrow et al. (2004) have completed a study encompassing cyclonic eddies found within the world's oceans. They used satellite altimetry to detect and follow cyclonic and anticyclonic eddies in the South-East Indian Ocean (Leeuwin Current eddies), South-East Atlantic Ocean (Agulhas Current and Cape Cauldron eddies) and North-East Pacific Ocean (Californian Current eddies).

Their main concern was how to analytically identify eddies from five years (1996 – 2000) of sea level anomaly maps. They chose to apply Isern-Fontanet et al.'s (2003) method of 'defining the eddy cores as regions in which the second invariant of the velocity gradient tensor (Q) is positive' (Morrow et al., 2004). They tested various
threshold values for each area and applied $Q_0 = 2 \times 10^{-4} \text{s}^{-1}$ to the Agulhas region, where the overall eddy energy is considered to be higher than their other two study areas. This allowed them to develop an automatic eddy propagation detection system, which was repeated at ten-day intervals.

Using this detection program they considered eddies within the Cape Cauldron that had life spans longer than four months. They determined that both cyclonic and anticyclonic eddies are large enough for the $\beta$-effect (figure 2.3) to impact on their trajectory, with planetary and relative vorticity pressure on the cyclone's perimeter affecting its meridional movement.

![Figure 2.3: Meridional drift of vortices on a $\beta$-plane (after Cushman-Roisin, 1994). This diagram depicts typical Northern Hemisphere eddies. In the Southern Hemisphere, anticyclones would move equatorward, while cyclones should move in a poleward direction (Morrow et al., 2004).](image)

Their case study showed that westward movement was determined by the negotiation of 'large topographical features and [a] relatively strong background flow, both of which can perturb the propagation pathways' (Morrow et al., 2004) as seen in figure 2.4. Garzoli et al. (1999), in contradiction, shows that the background flow in the Cape Basin is very weak.

In all, 66% of the cyclones observed by Morrow et al. (2004) travelled in a west-south-westerly direction at approximately 0.3 km/day, as shown in figure 2.4. In their conclusion they felt that further studies are needed to quantify the importance of these cyclonic and anticyclonic eddies in the counter-balance heat budget relationship between eastern and western boundary currents.
In a recent paper Chelton et al. (2007) used automated eddy tracking procedures to identify and track mesoscale vortices on a global scale. Eddies with amplitudes of 5–25 cm and diameters of 100–200 km were noted to provide more than 50% of the oceans variability. However, Chelton et al. (2007) expressed caution in that a smaller eddy size may be mistaken as noise in the merged sea surface height fields, as well as due to the particular threshold values of the Okubo-Weiss parameters chosen for the automotive detection.

Of particular interest is their observation that only a quarter of the global eddies studied translocated more than 10° from due west during the ten-year study period (1992 to August 2002). They determined that "globally, the percentages of eddies that propagated with equatorward deflection, purely zonally (0° ± 1°), and with poleward deflection, respectively, were 34%, 8% and 58% for the cyclonic eddies... the fact that nearly 1/3 of the observed eddies of each polarity had meridional deflections opposite of that expected may be consequences of eddy-eddy interactions and advection by background currents." (Chelton et al., 2007).

As Cape Basin cyclonic eddies are relatively new to the oceanographic world, not much research has been conducted on these mesoscale features. However, many scientists have studied the warm-water, anticyclonic eddies, known as Agulhas Rings.
The tools utilised within this study are the same as those that have been used to investigate Agulhas Rings, helping to establish the viability of these techniques. For example, XBT (Expendable Bathythermographs) probes, CTD (Current, temperature, depth profiler) data, NOAA-11 satellite imagery and GEOSAT altimeter data were used to characterize an Agulhas Ring detected between April to May 1989 by Duncombe Rae et al. (1992). The hydrographic data allowed the authors to obtain vertical temperature sections of the anticyclone to determine its size parameters and physical components.

Satellite altimetry data were used to track the Agulhas Ring from its origin and to follow its trajectory for as long as possible (curtailed due to data limitations). These data were also used to determine the Agulhas Ring’s interaction with the background Benguela Current. Size parameters, formation frequency and location, physical components and trajectories of cyclonic eddies are important information needed to understand the cyclonic eddies in this ocean region. A similar method of data gathering can therefore be used in this case.

Fu (2006) used the same method of satellite altimeter to follow eddies across the entire South Atlantic Ocean, aiming to provide a greater understanding of eddy propagation velocity, essentially providing an additional tool for ocean model testing. Although he mainly focused on the Western Atlantic, he established that northwestward eddies (anticyclones) progress at a speed of 3 – 4 km/day, once they have left the Agulhas Current region (Fu, 2006).

Byrne et al. (1995) used hydrographic data to determine Agulhas Ring size parameters and distribution, while satellite altimeter data were used to observe the studied Agulhas Ring paths and decay rates. Importantly, they discovered that ocean floor topography has an impact on the trajectory of the Agulhas Rings, slowing the anticyclonic eddies down over areas of steep topography, as shown in figure 2.5. These topographical features may also have an impact on the trajectory of the cyclonic eddies. Additionally, it is important to determine for how long these cyclonic eddies are able to exert an influence on the region. Therefore their decay rates and trajectory are valuable considerations.
In 1997 three Agulhas Rings were observed using TOPEX/POSEIDON satellite altimetry, XBT's, CTD's (conductivity-temperature-depth-oxygen) profiles, an ADCP (acoustic Doppler current profiler), an LADCP (lowered ADCP) and satellite-tracked surface drifters (Garzoli et al., 1999). Garzoli et al. (1999) published a paper identifying the origins of these anticyclonic Agulhas Rings and presented the rings' characteristics. They noted that each ring had individual structural properties. One of the rings exhibited properties of both Agulhas water and Subtropical Front water, allowing speculation that it had previously incorporated another eddy into its mass (Garzoli et al., 1999). Thus it is possible that many cyclonic and anticyclonic eddies interact with each other within the South Atlantic and especially within the highly energetic region of the Cape Basin. The question as to whether cyclonic eddies are associated with Agulhas Rings, with regards to formation, or whether they simply interact with each other is not known at present.

De Ruijter et al. (1999) undertook an in-depth study of Indian-Atlantic inter-ocean exchange, which would not have been complete without the inclusion of the effect of anticyclonic Agulhas Rings on the inter-ocean exchange. They noticed that the season had an effect on the Agulhas Rings that remained within the area of their formation over that season. Some Agulhas Rings have been observed to remain within their formation area for more than a year (Goñi et al., 1997).
Rings occluded before austral winter were associated with large thermosaline fluxes, whereas the rings shed in spring and early summer were subjected to far fewer changes and were noticeably different to winter rings, as determined from hydrographic data (Arhan et al., 1999; de Ruijter et al., 1999). If Agulhas Rings are affected by seasonality, it is logical to assume that cyclonic eddies in the same region are as susceptible to the changing seasons.

De Ruijter et al. (1999) tracked Agulhas Rings to their disintegration using satellite altimetry and many were found to reach as far across the Atlantic Basin as South America. This discovery allowed de Ruijter et al. (1999) to determine that Agulhas Rings provide a significant amount of anticyclonic vorticity to the South Atlantic, emphasizing their impact on the South Atlantic’s water transport, thermosalinity and energy levels. In this way, de Ruijter et al. (1999) believed that Agulhas Rings affect the stratification of the South Atlantic and therefore the global stratification, hence also playing a role in thermohaline overturning circulation.

Ferrari and Boccaletti (2004) agreed that horizontal eddy motion would have an effect on the background water and significantly modify water exchange rates in the localized area, therefore influencing the global exchange of ocean water properties. If anticyclonic eddies have such a significant effect on overturning circulation, it is therefore exceedingly worthwhile to consider a future study regarding the extent of cyclonic eddy vorticity within the same region.

Guilivi and Gordon (2006) approached the inter-ocean exchange topic via a hydrographical study of the isopycnal displacement of thermoclines in the Cape Basin. Using data from 24,486 hydrographic stations sampled between 1901 – 2002, they identified positive and negative depth anomalies, representing anticyclonic and cyclonic eddies respectively. Relying on potential temperature, salinity and oxygen properties identified at the thermocline and other depths, they identified the origin of the congregated eddy water. They noted that different formation locations and formation depths encapsulated differing water masses, namely Indian Ocean; tropical and subtropical South Atlantic waters.
Geographically, Guilivi and Gordon (2006) determined that eddies throughout the Cape Basin were found to contain water from the Agulhas Retroflection. Cyclones south of 30 °S and west of the Walvis Ridge generally contained South Atlantic water, while those west of 10 °E and north of the Walvis Ridge usually contained tropical Atlantic waters (Guilivi and Gordon, 2006).

Therefore little is known about the timing and frequency of cyclonic eddy formation, how often they interact with Agulhas Rings, their size parameters, trajectory and life spans. To understand their impact within the Atlantic Ocean we have to have a better understanding of these cyclonic eddies, features that may well ultimately change the currently accepted concept of Indian-Atlantic water exchange and its effect on global ocean circulation.
CHAPTER 3

Key Research Questions

The key questions of this thesis can therefore be divided neatly into three chapters.

3.1 Cape Basin cyclonic eddy formation

As shown in the knowledge review above, very little is known about the formation location and occurrence frequency of these cyclonic eddies. For the purpose of this study the focus will be primarily on those eddies formed between 23 – 35 °S; 5 – 17 °E and not those formed in the region of the Agulhas Retroflection, since the latter have been shown to have a totally different origin. Formation locality, occurrence rate and seasonality will form the focal point of this section. Questions such as whether formation location may be found within upwelling regions will also be addressed. In addition, the size characteristic of these eddies during formation will be considered.

3.2 Cape Basin cyclonic eddy lifestyles

Of equal importance in understanding these cyclonic eddies and their effects on the region are their lifespan and trajectory; size parameters, physical components, as well as the splitting and merging frequency of parent and child cyclonic eddies. Questions such as whether bottom topography has an effect on either the trajectory or potential splitting and merging of cyclones will be addressed. Hydrographic data will be used – where available – to correlate anomalies of sea surface height with eddy characteristics.

3.3 Cape Basin cyclone associations

An obvious question would be whether the generation of cyclonic eddies is associated with the passing of Agulhas Rings. Does the formation of a cyclonic eddy require the presence of an Agulhas Ring and are the splitting and merging of parent and child cyclonic eddies influenced by Agulhas Rings?
CHAPTER 4

Methods and data

4.1 Data

Over the past decade many scientists have studied the cyclonic and anticyclonic features of the ocean in the Cape Basin, as mentioned within the knowledge review of this study. On occasion, sea surface temperature images have been used to identify mesoscale features, such as Agulhas Rings that have propagated into the Atlantic Ocean. However, many oceanic eddies do not have surface temperature expressions that are consistently different from the background water throughout the year, for example, those found within the Gulf of Mexico (Zavala-Hidalgo et al., 2003). Therefore satellite altimetry collected by the Colorado Center for Astrodynamic Research was used to determine sea surface height (SSH) anomalies, allowing the authors to negate seasonal temperature variability.

Lutjeharms et al. (2003) have used two sets of satellite data: sea surface thermal infrared data (collected by a NOAA satellite Advanced Very High Resolution Radiometer) and sea surface height data (MODAS-2D, TOPEX/Poseidon and ERS2 altimetric data). RAFOS floats were deployed to form the Lagrangian drift part of their study. The thermal data were apparently subject to inopportune cloudiness, while the sea surface height data appeared to be associated conclusively with the data recorded by RAFOS floats. One of the main results to emerge from their study was that sea surface height data appeared to be a very useful tool to track vortices in intermediate waters within the Agulhas Current system.

Boebel and Barron (2003) have looked at the relationship between float trajectories and altimeter-derived geostrophic velocities. They concluded that the MODAS-2D sea surface height data accurately depict cyclonic and anticyclonic features within the wider Agulhas Retroflection region. They believed that the sea surface height data contained sufficient resolution to track these mesoscale features with confidence. Boebel et al. (2003) did just that and used the same RAFOS float and MODAS-2D SSH data sources to investigate inter-ocean exchange within the Cape Basin itself.
Due to the efforts of those above in proving the effectiveness and resolution of current satellite altimetry, ten years of altimetric sea height anomaly data (1992-2002) were used to answer the majority of the key questions in this study. As mentioned, the sea height anomalies were calculated from gridded TOPEX/Poseidon and ERS1-2 sea level anomaly data that were produced by Ssalto/Duacs and distributed by Archivage, Validation et Interprétation des données des Satellites Océanographiques (AVISO), with support from CNRS (http://www.aviso.oceanojobs.com/html/donnees/products/hauteturs/globalmsla_uk.html, 2005).

These sea level anomaly data are gridded sea surface heights that are computed with respect to a seven-year mean. They are available in near-real and delayed time. For the purpose of this study a high resolution, 1/3° x 1/3°, Mercator grid product was deemed the most effective to investigate cyclonic eddies. More information on the processing of these data may be found at http://www.jason.oceanojobs.com/html/donnees/traitemen/welcome_uk.html (2007). This sea height anomaly product was downloaded in NetCDF format. Matlab version 7.0.4 was used to graphically display these data and then join geographical regions to represent the entire study area.

However, these satellite altimetry data are only useful if there is conclusive proof that the actual sea height anomalies used during this study truly represent cyclonic eddies in this region. A number of studies that used altimetry have been conducted on Agulhas Rings that occlude from the Agulhas Retroflection (Lutjeharms, 2007; Schmid et al., 2003; Arhan et al., 1999; Gründlingh, 1995; Lutjeharms and van Ballegoojen, 1988; to name but a few). As identified by Schmid et al., 2003 (p: 145), a 'regime of enhanced sea-surface temperatures and salinities...[where] the 10 °C isotherm [deepens] from 200 dbar...to 800 dbar...before rising again...is indicative of an anticyclonic circulation pattern and typical for Agulhas Rings.' This description shows clearly that a convex pattern in the thermocline indicates an anticyclonic eddy.

It is therefore logical to assume that, within the southern hemisphere, a dramatic doming of the thermocline, identifiable by a rise in the 10 °C isotherm, would indicate the presence of a cyclonic eddy, as shown in figure 4.1 (Boebel et al., 2003) below. Therefore, concurrent hydrographic and altimetric data depicting a rise in the
thermocline and with a depression in the mean sea surface height, respectively, would confirm the existence of a cyclonic eddy.

![Diagram of ADCP sections and XBT sections](image)

**Figure 4.1: ADCP sections of KAPEX leg c data, combined with XBT sections. ADCP velocities are assigned to the center depth for each bin, e.g. to 150 m for the 325-375 m bin. The x-axes are scaled to be approximately equivalent for the three subplots. Positions of the Agulhas Current (AC), the Agulhas Return Current (ARC) and of a cyclonic eddy are indicated (modified from Rochel et al., 2003).**

During this study available hydrographic data were investigated to determine whether a rise in the thermocline coincided with an altimetric signal that indicated a depression in the mean sea surface height at the same location. Should the data sets both reveal the presence of a cyclonic eddy in their respective manner, it is possible to determine that each method is accurate enough to resolve mesoscale features and may be used as a tool in isolation.

Hydrographical data were gathered from a myriad of data sources and programmes, such as ARGO (XBT High Density Repeat Lines; [http://www.nrml.nos.nl/phod/hd除此_ds/hd除此_getdata.htm](http://www.nrml.nos.nl/phod/hd除此_ds/hd除此_getdata.htm)), BENT (Benguela: Sources and Transport: June 1992 - November 1993; Garzoli et al., 1996), KAPEX (Cape of Good Hope Experiments: April 1997 - June 1998), MARE (Mixing of Agulhas Rings Experiment, 2005; [http://kelina.noaa.gov/projects/mare](http://kelina.noaa.gov/projects/mare)), SAVE (South Atlantic Ventilation Experiment: 1987 - 1989; [http://cdiac.esd.ornl.gov/lp/oceans/save/](http://cdiac.esd.ornl.gov/lp/oceans/save/)), CLIVAR and CHIDO (Carbon Hydrographic Data Office) WHPO cruises ([http://cehdo.ornl.edu/maps/atl_south.htm](http://cehdo.ornl.edu/maps/atl_south.htm)) and WOCE (World Ocean Circulation Experiment: 1990 - 2002; [http://www.nodc.noaa.gov/OC5/](http://www.nodc.noaa.gov/OC5/)). Certain data were provided by SAICDO (Southern African Data Centre for Oceanography; [http://saicdo.esri.co.za/data.htm](http://saicdo.esri.co.za/data.htm)), the rest was obtainable from the corresponding links above.
Figure 4.2: Hydrographical data available for use during this study. Each point represents the location of a hydrographical station. Bottom topography is represented by the colour legend.

Figure 4.3: Temporal distribution of hydrographical data collected for the region represented in figure 4.2 above, as displayed per month and year (Ocean Data View).

Figure 4.2 shows an overview of the hydrographical data collected for use within this study. Figure 4.3 shows a temporal distribution of the gathered data. Unfortunately, although a large collection was gathered, not many hydrographical data are available that correspond to identified altimetric cyclonic eddies within the study area. However, the following cyclonic eddy is one of the ten cyclones that were concurrent in both data sets and is given here as an example.
Cyclone 92E is identified as a sea surface height (SSH) negative anomaly from TOPEX/Poseidon and ERS data (MATLAB and ACDSec), as shown in figure 4.4 below. Cyclone 92E is confirmed using hydrographical temperature and salinity data obtained from WOCE surveys within the study area. The map (c) in figure 4.5 shows the line of hydrographical stations that provided the data for the Ocean Data View temperature (a) and salinity (d) sections shown in figure 4.5.

![Cyclone 92E and SSH High](image)

**Figure 4.4**: Altimetry image 20 and 27 January 1993 (AVISO 2005). Cyclone 92E and a positive sea surface high are indicated by black arrows. Hydrographic sampling track line is displayed with a white line.

Cyclone 92E is flanked in figure 4.5 a & d by a SSH high, as identified in the altimetry data. Figure 4.5 a & d show how the isotherms and isohalines draw closer to the surface under the influence of Cyclone 92E, affecting the water column to a depth of approximately 1000 m. Meanwhile, the SSH positive anomaly depresses the isohaline and isotherms, ensuring an increase of thermohaline values at greater depths. Other examples of cyclonic eddies that appear in both data sets may be found in Appendix D (page: 151; figures 9.25 – 9.37).
Hence, there is conclusive proof that the actual sea height anomalies used during this study truly represent cyclonic eddies in this region. Next, in order to investigate their behaviour, it is essential to determine whether it is possible to consistently track these features throughout their lifespan.

![Temperature section through the region](image)

![Salinity section through the region](image)

**Figure 4.3:** Cyclonic eddy 02E. (a) Temperature section through the region; (b) thermohaline IT/SA plot of the section; (c) map showing hydrographic WOCE stations; and (d) salinity section through the region.

Appendix A (page 135) contains a brief overview of an individual cyclonic eddy that was identified during the study period. In this way an attempt is made to show the reader how it is possible to track these cyclonic eddies as the single entities that they are, moving independently yet reacting to the influence of anticyclonic eddies and the background water flow. Cyclonic eddies were considered formed and traceable after they were 10 cm deeper than the average mean sea level at formation. Their lifespan was deemed over once their trace was less than −10 cm deep in sea surface height.

Each log entry in Appendix A represents a statement of activity of the specific cyclonic eddy. Most entries in the appendix are weekly, as per the gridded weekly output of the AVISO data, unless the cyclonic eddy has remained stationary. Should that be the case, the next entry reflects its new movement. The location of the cyclonic
eddy is logged first, then the date (day, month, year), followed by a description of its activity. This information has been collated and displayed graphically in Figure 4.6 below, where each position corresponds to the figures in Appendix A. In both figure 4.6 and Appendix A, it is possible to see the cyclonic eddy's translocation and variability over time.

![Figure 4.6: The translocation of cyclonic eddy 94A. The center of the eddy is indicated by a dot and a solid line indicates the outline. Positions correspond to the following dates: A: 5 January 1994; B: 23 March 1994; C: 18 May 1994; D: 26 July 1994; E: 24 August 1994; F: 21 September 1994; G: 7 December 1994; H: 11 January 1995; J: 12 July 1995 (and J: 21 January 1996. Original AVISO data and shape explanations for cyclonic eddy 94A may be found in Appendix A.)](image)

From Figure 4.6 above it is possible to determine that cyclonic eddy 94A was formed (position 'A' in Figure 4.6) on the continental slope at 34 °S 19 °E, within the Cape Peninsula Upwelling cell. It was formed directly north of a positive sea surface height anomaly, generated within the region of the Agulhas Retraction. Over its lifespan of 24 months, 94A moved from its formation location at 34 °S 16 °E to 33 °S 11 °W, therefore mainly westerly with a very slight south angle. Over its lifespan, 94A achieved a size diameter of approximately 3 square degrees as a single cyclone (positions 'D' and 'E' in Figure 4.6). However, once entrained with the other SSH low in a dipole system (position 'D'), their actual depression size is much greater. As typical of this data set, the sea surface height of 94A varies between -10 to -40 cm from the average mean of the region.
Therefore it is possible to ascertain how varied cyclonic eddy motion is in association with other eddies and their impacts. Splitting, merging and linkage between cyclonic eddies, as well as adjacent anticyclonic eddies, necessitates the need for careful consideration of each step of their lifespan. Automated methods may be used to detect and track mesoscale features but are in the development stage (Seigel and Weiss, 1997; Luo and Jameson, 2002; Chelton et al., 2007; Doglioli et al., 2007). Automation would expedite the collection these data and potentially remove subjective bias introduced by individual researchers. However, due to the subjective process of selecting the particular threshold value used during the automatic search criteria, it was thought that a visual inspection of each cyclonic eddy was the most thorough method applicable. In this way it would be possible to ascertain the complicated behaviour and movement of cyclonic eddies within the Cape Basin.
4.2 Statistical analysis

One type of data was present within the data set, known as discrete variables. Discrete variables "can take on only certain values" (Zar, 1984). Hence these data values consist of whole numbers, such as the number of cyclones formed per year. Due to the nature of the questions asked, different statistical methods were used to test the null hypotheses.

When appropriate, the data were considered to determine any deviation from a Gaussian distribution (normality testing). This was achieved by two methods; one interpretive and the other computed. Histograms, which plot the frequency distribution of the data to be analysed, indicated the skewness (whether the data are symmetrical or "off centre") and/or kurtosis (sharpness of the peak of the distribution curve) of the data. These were created using R version 2.6.0 (© The R Foundation for Statistical Computing) and Statistica version 6.1 (© StatSoft, Inc. www.statsoft.com) software.

Computed normality tests were conducted using GraphPad InStat version 3, minimizing the potential for user error. InStat examines for normality by applying a Kolmogorov-Smirnov test (described below) to calculate a KS statistic ($D$); whereby a high value $D$ suggests a large discrepancy between the Gaussian and the data's distribution. Ultimately, should the data originate from a Gaussian population, the resulting probability ($P$) value would indicate the possibility that the randomly selected sample of said particular size would have a KS distance as large, or larger, than detected (GraphPad InStat, 2007).

Not all data passed the Kolmogorov-Smirnov normality test, such as the child cyclonic eddy annual lifespan data. Various methods may be employed to transform the data to a lognormal distribution (normally used for measurement); in this study notably the conversion of the data to a Poisson distribution using a square root of a count; or conversion to a binomial distribution using a arcsine of square root of proportion function. Subsequent failure to convert the data successfully necessitated the use of nonparametric methods for determining statistical relationships in these instances.
GOODNESS OF FIT TESTS

Goodness of fit (GOF) tests determines normality of the data and subsequently proceeds to draw conclusions about certain parameters of the distribution, such as whether the difference between the observed and expected frequencies of the sample are too extreme to be attributed purely to random sampling fluctuations (Underhill, 1985). The number of samples within the data set dictates the correct GOF test to use, as a one-sample data set asks whether the sample arises from a hypothesized distribution, while a two-sample data population queries whether the two independent samples arise from the same distribution (S-PLUS 6 for Windows Guide to Statistics, 2001). The two best-known GOF tests used during this study are the Chi-square and, as mentioned above, the Kolmogorov-Smirnov statistical test.

The chi-square test applies only in the one-sample case; the Kolmogorov-Smirnov test can be used in both the one-sample and two-sample cases. Both the chi-square and Kolmogorov-Smirnov GOF tests work for many different distributions.

THE KOLMOGOROV-SMIRNOV GOODNESS-OF-FIT TEST

The Kolmogorov-Smirnov (K-S) goodness-of-fit test is a cumulative distribution function test and was applied to the data to test for normality. Quite simply, the test statistic (D) is calculated as the difference between the cumulative frequency of the observed values ($F_1$) and the cumulative frequency of the expected values ($F_2$). Therefore the K-S test is based on the greatest vertical distance between the distributions. Should the calculated $D$-statistic be greater than the critical $D$-statistic, it is possible to reject the null hypothesis and assume that the data are not from a normal distribution (Weisstein, 1999).

As clearly documented in S-PLUS 6 for Windows Guide to Statistics (2001), there are three possible hypotheses and associated test statistic formulae that may be used as appropriate.
a) For a two-sided situation:

$H_0: F_1(x) = F_2(x)$ for all $x$

$H_A: F_1(x) \neq F_2(x)$ for at least one value of $x$

Appropriate test statistic:

$T = \sup_x \mid F_1(x) - F_2(x) \mid$

b) For a one-sided "less" situation:

$H_0: F_1(x) \geq F_2(x)$ for all $x$

$H_A: F_1(x) < F_2(x)$ for at least one value of $x$

Appropriate test statistic:

$T = \sup_x \mid F_1(x) - F_2(x) \mid$

c) For a one-sided "greater" situation:

$H_0: F_1(x) \leq F_2(x)$ for all $x$

$H_A: F_1(x) > F_2(x)$ for at least one value of $x$

Appropriate test statistic:

$T^* = \sup_x \mid F_1(x) - F_2(x) \mid$

As mentioned, this test was mainly used in this study to determine whether the considered data represents a normal distribution. Should the data pass the normality test, a parametric statistical test such as a $t$-test or ANOVA was applied.

**THE CHI-SQUARE GOODNESS OF FIT TEST**

The chi-square test is the most frequently used statistical test in this study. To quote S-PLUS 6 for Windows Guide to Statistics, Volume 1 (2001): "It is a one-sample test that examines the frequency distribution of $n$ observations grouped into $k$ classes and assumes a normalised distribution. The observed counts $O_i$ in each class are compared to the expected counts $E_i$ from the hypothesized distribution. The test statistic, due to Pearson, is:

$$
\chi^2 = \sum_{i=1}^{k} \frac{(O_i - E_i)^2}{E_i}
$$
Under the null hypothesis that the sample comes from the hypothesized distribution, the test statistic $\chi^2$ has a distribution with $k - 1$ degrees of freedom. For any significance level $\alpha$, reject the null hypothesis if $\chi^2$ is greater than the critical value for which:

$$P(\chi^2 > v) = \alpha.$$ 

Chi-square tests apply to any type of variable: continuous, discrete, or a combination of these. For large sample sizes ($n \geq 50$), the chi-square is the only valid test when the hypothesized distribution is discrete. In addition, the chi-square test easily adapts to the situation when parameters of a distribution are estimated.

In this study the chi-square test was used to determine the statistical significance of annual, seasonal and monthly frequency of formation of both parent and child cyclonic eddies. It was also used to consider annual, parent and child merge data.

**ANOVA**

The analysis of variance (ANOVA) statistical test is used to compare the mean annual, seasonal and monthly life span of the various cyclonic eddies. This test is appropriate for multisample analysis and is not affected by sample size (Zar, 1984). Importantly, ANOVA has been proven to be robust with respect to data normality. Therefore, should the data distribution be abnormal, ANOVA would minimize the impact of deviations from normality.

ANOVA indicates the significance of the difference between the samples means. An $F$-value is calculated by dividing the variances between the samples with the variances within the samples. This calculated $F$-value is compared with the critical $F$-value, found in the $F$-table. Should the calculated $F$-value be less than critical $F$-value, the null hypothesis is not rejected.

The hypotheses for ANOVA are as follows:

- $H_0$: All means are the same.
- $H_1$: Not all means are the same.
Let \( n \) be the number of values in each sample with mean \( \bar{x} \) and standard deviation \( s \), \( k \) the number of samples that are compared. Let \( N \) be the grand total of all the values in all the samples and \( \bar{x} \) the mean of all the values (df represents degrees of freedom).

Then, \[
\bar{x} = \frac{\sum n \bar{x}}{N}
\]

Variance between samples = \[
\frac{\sum n (x - \bar{x})^2}{k - 1}
\]
with df = \( k - 1 \)

Variance within samples = \[
\frac{\sum (n - 1)s^2}{N - k}
\]
with df = \( N - k \)

Calculated \( F = \) \frac{\text{Variance within samples}}{\text{Variance between samples}}

To draw a conclusion, this calculated \( F \)-value is then compared with the corresponding critical \( F \)-value. Should the \( F \)-critical value be greater than the calculated \( F \)-value, the null hypothesis is not rejected.

**BARTLETT’S TEST**

Bartlett’s test (Snedecor and Cochran, 1983) is used to test if \( k \) samples have equal variances. The Bartlett test may be used to verify whether variances are equal across sample groups and is sensitive to deviations from normality. Bartlett’s test is used to test the null hypothesis, \( H_0 \), that all \( k \) population variances are equal against the alternative hypothesis, \( H_a \), which assumes that at least two populations are different.

If there are \( k \) samples with size \( n_i \) and sample variance \( S_i^2 \) then Bartlett’s test statistic is:

\[
X^2 = \frac{(N - k) \ln(S_p^2) - \sum_{i=1}^{k}(n_i - 1) \ln(S_i^2)}{1 + \frac{1}{n(\bar{n} - 1)} \left[ \frac{\sum_{i=1}^{k}(\frac{1}{n_i - 1} - \frac{1}{\bar{n}})}{\bar{n} - k} \right]}
\]
\[ N - \sum_{i=1}^{k} n_i \quad \text{and} \quad S_p^2 = \frac{1}{N - g} \sum_{i} (n_i - 1) S_i^2 \]

where \( S_p^2 \) is the pooled estimate for the variance.

The test statistic has approximately a \( \chi^2 \) distribution. Thus the null hypothesis is rejected if \( \chi^2 > \chi_{k-1, \alpha}^2 \) (where \( \chi_{k-1, \alpha}^2 \) is the upper tail critical value for the \( \chi^2 \) distribution).

Bartlett’s test is a modification of the corresponding likelihood ratio test designed to make the approximation to the \( \chi^2 \) distribution better (Bartlett, 1937).

T-TEST

The unpaired \( t \)-test was used to determine whether there was a relationship between the lifespan of parent cyclonic eddies formed within the northern and southern regions of the study area. This test compares the means of two samples to show the possibility that both samples were drawn from the same population, based on two assumptions. These assumptions are that the observations have a common normal (or Gaussian) distribution with mean \( \mu \) and variance \( \sigma^2 \), and secondly that the observations are independent.

Should these two assumptions be met, the two-tailed \( t \)-test is robust with regard to the equality of sample size and the normality of the data (Zar, 1984). Another point to consider before using this test is that a \( t \)-test confidence interval of 95% may have a much higher error rate than 5% when there is a small amount of positive correlation in the data. Additionally, most modern robust methods are oriented toward obtaining insensitivity toward outliers generated by heavy-tailed nearly normal distributions, and are not designed to cope with serial correlation” (Zar, 1984). Hence a Welch correction was added to the \( t \)-test analysis (Heidelberger and Welch, 1981).

The equations for Student \( t \)-test for samples \( x \) and sample size \( n \) are as follows:

\[ H_0: x_1 = x_2 \text{ (both means are the same)} \]

\[ H_a: x_1 \neq x_2 \text{ (both means are not equal)} \]
Therefore,

\[ t = \frac{x_1 - x_2}{\sqrt{\frac{s_1^2}{n_1} + \frac{s_2^2}{n_2}}} \]

Associated degrees of freedom:

\[ df = (n_1 - 1) + (n_2 - 1) - n_1 + n_2 - 2 \]

Subsequently, if the calculated t-value is greater than the critical t-value, found using the calculated degrees of freedom (df), the null hypothesis is not rejected, allowing for the assumption that both means are the same. However, as statistics is not an absolute science, i.e., no test is completely accurate due to Type I and II errors, the level of confidence that may be assigned to the test should be calculated.

The confidence limit may be denoted as:

\[ x_1 - x_2 \pm t_{(\alpha/2), df} \sqrt{\frac{s_1^2}{n_1} + \frac{s_2^2}{n_2}} \]

Statistical tests only really allow the user to predict the propensity that the same test will return the same result when applied to the data. Therefore, should the level of confidence be set at \( \alpha = 0.05 \) (95%), it is then possible to assume that 95 out of 100 tests will return the same result.

**MANN-WHITNEY NON-PARAMETRIC TEST**

The Mann-Whitney test statistic, also known as the Wilcoxon rank sum statistic, is a non-parametric method for comparing two independent samples. It was used in this case to investigate the relationship between child cyclonic eddies formed within the northern and southern regions of the study area.

The test itself is most easily described in terms of treatment and control groups. Given a set of \( m + n \) units, the \( n \) set is randomly selected and assigned as a control group, leaving \( m \) units for a treatment group. After measuring the effect of the treatment on all units, the \( m + n \) observations are grouped together and ranked in order of size. If
the sum of the ranks in the control group is too small or too large, then it is possible to state that the treatment had an effect. Therefore the Mann-Whitney sum statistic portrays the probability characteristics of the test values, where the rank sum statistic takes on values found between the following equations:

\[ U = \frac{mr(m + 1)}{2} \]
\[ U' = \frac{mr(m + 1)}{2} \]

**KRUSKAL-WALLIS NON-PARAMETRIC ANOVA**

McDonald (2008) describes the Kruskal-Wallis test (Kruskal and Wallis, 1952) clearly: "The Kruskal-Wallis test is most commonly used when there is one nominal variable and one measurement variable, and the measurement variable does not meet the normality assumption of an ANOVA. It is the non-parametric analogue of a one-way ANOVA. A one-way ANOVA may yield inaccurate estimates of the P-value when the data are very far from normally distributed. The Kruskal Wallis test does not make assumptions about normality. Like most non-parametric tests, it is performed on ranked data, so the measurement observations are converted to their ranks in the overall data set: the smallest value gets a rank of 1, the next smallest gets a rank of 2, and so on. The loss of information involved in substituting ranks for the original values can make this a less powerful test than an ANOVA, so the ANOVA should be used if the data meet the assumptions." (McDonald, 2008)

The test statistic is given by:

\[ K = (N - 1) \frac{\sum_{g=1}^{G} n_g \left( \bar{r}_g - \bar{r} \right)^2}{\sum_{g=1}^{G} \sum_{j=1}^{n_g} (r_{ij} - \bar{r})^2} \]

where:

- \( n_g \) is the number of observations in group \( g \)
- \( r_{ij} \) is the rank (among all observations) of observation \( j \) from group \( i \)
- \( N \) is the total number of observations across all groups

\[ \bar{r}_g = \frac{\sum_{j=1}^{n_g} r_{ij}}{n_g} \]

\[ \bar{r} = \frac{(N + 1)}{2} \] is the average of all the \( r_g \).
The denominator of the expression for $K$ is exactly $(N - 1) N(N + 1) / 12$. Thus,

$$K = \frac{12}{N(N + 1)} \sum_{i=1}^{G} n_i (\bar{r}_i - \bar{r})^2$$

A correction for ties can be made by dividing $K$ by

$$1 - \frac{\sum_{i=1}^{G} (t_i^3 - t_i)}{N^3 - N},$$

where $G$ is the number of groupings of different tied ranks, and $t_i$ is the number of tied values within group $i$ that are tied at a particular value. This correction usually makes little difference in the value of $K$ unless there are a large number of ties.

Finally, the $p$-value is approximated by $\Pr(\chi^2_{g-1} \geq K)$. If some $n_i$'s are small (i.e., less than 5) the probability distribution of $K$ can be quite different from this chi-square distribution. If a table of the chi-square probability distribution is available, the critical value of chi-square, $\chi^2_{g-1}$, can be found by entering the table at $g - 1$ degrees of freedom and looking under the desired significance or alpha level. The null hypothesis of equal population medians would then be rejected if $K \geq \chi^2_{g-1}$. Appropriate multiple comparisons would then be performed on the group medians.

To recap, after determining data normality and considering the type of result required from the statistical test, the various methods explained above would be available for use to analyse the following data. Chapter five deals with the first section of key questions detailed in chapter three.
CHAPTER 5

Cyclonic Eddy Formation in the Cape Basin

5.1 Formation locations

Most of the Cape Basin parent cyclonic eddies in this study appear to be formed near to the coast of Southern Africa (figure 5.1). From the figure it seems apparent that a large proportion of the parent cyclonic eddies are formed along the edge of the continental shelf, or the shelf slope. On closer inspection, there seem to be specific regions of higher incidence of cyclone creation.

![Cyclonic eddy parent and child formation location in the Cape Basin (1992 - 2002)](image)

Figure 5.1: Cyclonic eddy formation location within the Cape Basin (1992 - 2002). Black circles represent parent cyclones, while red crosses represent child cyclones that have split from the parent cyclone.

The formation of child cyclonic eddies is a product of parent cyclone splitting. This may happen as a parent cyclonic eddy occludes a smaller cyclonic eddy or as the parent cyclonic eddy splits into two. After splitting, the parent cyclonic eddy is considered to be the cyclonic eddy that retains the larger size and original trajectory. The formation of child cyclonic eddies has a far wider formation region. As is apparent in figure 5.1, most splitting occurs within the deeper Cape Basin. However, once the northern-most cyclones have crossed the Walvis Ridge, splitting still occurs.
During the study period, no southern cyclones split after they had left the Cape Basin, even though a few of these cyclones did proceed into the region above the Mid-Atlantic Ridge (as shown in the trajectories section further on in chapter 6; figure 6.1, page 82). This act of splitting will be discussed in more detail within the cyclone lifestyles portion, found within chapter 6 of this thesis.

Some have suggested that cyclonic eddy formation may be along the continental shelf (Boebel et al., 2003; Shannon and Nelson, 1996). To test this theory, formation location data were separated zonally into those cyclonic eddies formed to the east of the continental slope edge and those formed within the deeper Basin. The edge of the continental slope was defined as following the 4000 m isobath, using the following geographical limits: north of 30°S, the slope was east of 10°E; between 30 - 32.0°S, the slope was east of 12°E; between 32 - 34.0°S; the slope was east of 14°E and between 34 - 36.0°S, the slope was east of 16°E.

Table 5.1: Continental slope formation frequency of parent and child cyclonic eddies

<table>
<thead>
<tr>
<th></th>
<th>Cyclones formed to the east of the defined continental slope edge</th>
<th>Cyclones formed to the west of the defined continental slope edge</th>
</tr>
</thead>
<tbody>
<tr>
<td>Parent cyclonic eddies</td>
<td>97</td>
<td>8</td>
</tr>
<tr>
<td>Child cyclonic eddies</td>
<td>34</td>
<td>123</td>
</tr>
</tbody>
</table>

With a ratio of 12:1, table 5.1 clearly shows that most parent cyclonic eddies considered during this study were formed to the east of the 4000 m isobath, indicating continental slope formation. Conversely, child cyclonic eddies were found to form predominately to the west of the 4000 m isobath, as expected due to their formation process (splitting from already formed parent cyclonic eddies). Such an overwhelming ratio of parent cyclonic eddies formed at the slope certainly supports the shelf formation theory proposed by Boebel et al., 2003.

As mentioned above, the formation of most parent cyclonic eddies do appear to occur in clusters along the continental slope, although a few parent cyclonic eddies were found to form along on the continent’s coastline, near Cape Point, Lamberts Bay and Port Nolloth. A review of the literature (Shannon and Nelson, 1996; Hutchings, 1992; Lutjeharms and Meeuwis, 1987) shows that the geographical locations of these formation hotspots agree substantially with the location of known upwelling cells.
Here, upwelling inducing winds (Shannon and Nelson, 1996) and steep shelf topography promote seasonal and perennial upwelling (Taunton-Clark, 1985), as shown in Figure 5.2.

![Figure 5.2: Parent cyclonic eddy formation location with regards to known upwelling cells. The pink circles represent the Central Namibia Upwelling Cell; the orange circles represent the Luderitz Cell and the yellow circles represent the Namaqua Cell. The purple circle represents upwelling around Dassen Island, the green circles represent the Columbine Cell, the red circles represent the Cape Peninsula Cell and the blue circles represent cyclonic eddies not formed within these known cells.](image)

Defining the upwelling cells as per Shannon and Nelson, 1996, the Central Namibia cell (~ 24 °S); the Luderitz cell (27 – 28 °S); the Namaqua cell (31 °S); the Columbine cell (33 °S) and the Cape Peninsula cell (34 °S) may all be found along the West Coast of Southern Africa (Figure 5.2). Near Dassen Island, 33 °S, waters sometimes form a "mushroom head" beyond the shelf as they flow past the island (Shannon and Nelson, 1996). The region where a cyclonic eddy might form part of such a vortex dipole is visible in Figure 5.2 and is identified by a purple circle.

Additionally, zonally orientated fronts are thought to develop adjacent to major upwelling cells: to the north of the perennial Luderitz cell and to the north of the Cape Peninsula and Columbine cells in summer (Shannon and Nelson, 1996). These
upwelling fronts are 'coincident with the zone of maximum cyclonic wind-stress curl and [lie] immediately south of an area where wind-induced turbulence is significantly lower and stratification of the shelf waters is stronger' (Shannon and Nelson, 1996). Therefore it is possible that these tongues promote cyclonic eddy formation within coastal areas that are not part of a recognised upwelling cell.

With upwelling cells defined as such, it is possible to determine that within the study period and region, sixty-eight out of ninety-six parent cyclonic eddies were formed within these cells, including the upwelling region around Dassen Island.

### Table 5.2: Number of parent cyclonic eddies formed per upwelling location

<table>
<thead>
<tr>
<th>Upwelling Location</th>
<th>Number formed</th>
<th>Percentage ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>Central Namibia Cell</td>
<td>4</td>
<td>5.88%</td>
</tr>
<tr>
<td>Lüderitz Cell</td>
<td>29</td>
<td>42.65%</td>
</tr>
<tr>
<td>Namaqua Cell</td>
<td>22</td>
<td>32.35%</td>
</tr>
<tr>
<td>Dassen Island</td>
<td>1</td>
<td>1.47%</td>
</tr>
<tr>
<td>Columbine Cell</td>
<td>6</td>
<td>8.82%</td>
</tr>
<tr>
<td>Cape Peninsula Cell</td>
<td>6</td>
<td>8.82%</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>68</strong></td>
<td></td>
</tr>
</tbody>
</table>

Table 5.2 shows a more comprehensive breakdown of the number of parent cyclonic eddies formed within each upwelling cell during the study period. In addition, both figure 5.2 and table 5.2 shows that not all the parent cyclonic eddies were formed upon the continental slope but that some were generated within the Cape Basin itself.

As shown in figures 5.3 and 5.4, the formation locations of both parent and child cyclonic eddies do not seem to reflect marked annual change in distribution patterns. Along the continental shelf, most years are represented in each known upwelling cell. Therefore it is possible to deduce that each year parent cyclones are formed within all cell locations along the continental shelf.
Figure 5.3: Cyclonic eddy parent formation (1992 - 2002). The colours of the circles represent a specific year.

Figure 5.4: Cyclonic eddy child formation (1992 - 2002). The colours of the crosses represent a specific year.
With no obvious annual effect on formation location, seasonality is the next time scale to consider due to the seasonality of upwelling in the south as opposed to the northern region of the study area. Figure 5.5 details the seasonal formation location of both parent and child cyclonic eddies over the study period. As is visible, the formation location of both parent and child cyclonic eddies does not appear to exhibit any clear seasonal pattern.

![Diagram of cyclonic eddy formation](image)

**Figure 5.5: Seasonal parent and child cyclonic eddy formation (1992 - 2002).** The colour of the circles and crosses represent a specific season of parent and child cyclonic eddy formation respectively.

Throughout the 1992 - 2001 study period, (figure 5.6) parent cyclonic eddies formed in winter seemed to originate entirely on the edge of the continental shelf and on the continental slope. Only two cyclones were formed on the shelf south of Cape Point and on the Agathas Bank. The same continental slope location is true for most of the parent cyclonic eddies formed in spring (figure 5.7). In that parent cyclonic eddies formed north of 30°S seem to be entirely on the edge of the continental shelf or in the deeper offshore waters. Apart from a few parent cyclonic eddies formed at the Lamberts Bay and Cape Point locations, this is also true for the southern region of the study area during springtime.
Figure 3.6: Winter parent and child cyclonic eddy formation (1992 - 2002). The black circles and crosses represent parent and child cyclonic eddy formation respectively.

Figure 3.7: Spring parent and child cyclonic eddy formation (1992 - 2002). The green circles and crosses represent parent and child cyclonic eddy formation respectively.
Summer (figure 5.8) and autumn (figure 5.9), on the other hand, see a larger number of parent cyclonic eddies generated on the actual slope, as well as nearer to the continent. In fact, Port Nolloth and Lambert's Bay appear to be locations where cyclones often form during these seasons. As with winter, summer formation location seems to be restricted to the edge of the continental shelf and not in the actual basin. Autumn follows suit and only the odd parent cyclonic eddy may be found to generate off the shelf.

Child cyclonic eddy formation during the study period was mainly restricted to the Cape Basin. Interestingly, each season had a number of child cyclonic eddies that were formed as their parents collide with the Walvis Ridge and Vema Seamount. Topographical barriers have been known to split these eddies (Boebel et al., 2003); Schouten et al. (2000) showed Agulhas Rings splitting at the Vema Seamount. These data (figure 5.5) showed that the bathymetry had the same affect on these cyclonic eddies, with seven obvious examples in winter’s figure 5.6, four in spring’s figure 5.7, eight in summer’s figure 5.8 and three features in autumn’s figure 5.9.

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**Figure 5.8:** Summer parent and child cyclonic eddy formation (1992 - 2002). The red circles and crosses represent parent and child cyclonic eddy formation respectively.
Other child cyclonic eddies were formed outside of the Cape Basin and seem to be derived from parent cyclonic eddies that were able to transverse the Walvis Ridge. At this point it is important to remember that each parent cyclonic eddy is capable of shedding more than one child, so that the child cyclonic eddies formed to the west of the Cape Basin may be one in a string of eddies created by a parent. Additionally, child cyclonic eddies also have the potential to split themselves, creating a veritable family of siblings from one original parent cyclonic eddy.

During the winter season (figure 5.6), five child cyclonic eddies were formed just south of the 30 °S line of latitude. However, as mentioned above, this may have been due to a single, active cyclonic eddy family. This will be possible to confirm when looking at formation locations of individual years. In spring (figure 5.7) four child cyclonic eddies were formed to the east of the Walvis Ridge, although their location suggest that their formation is not related to one particular parent cyclone. During both the summer (figure 5.8) and autumn season (figure 5.9) four child cyclonic eddies were formed. Again their relation is difficult to determine from a seasonal viewpoint and an annual perspective may shed some light on this.
Figure 5.10: Cape Basin cyclonic eddy parent and child formation location during January - December 1993.

Figure 5.11: Cape Basin cyclonic eddy parent and child formation location during January - December 1997.
Figure 5.10 shows a typical year, with most of the parent cycloic eddy formation on the shelf as given above. One parent cyclone was formed in the Cape Basin itself. Due to its proximity to the Vema Seamount, it may be possible to speculate that the Vema Seamount caused the formation of this cycloic eddy. Other years to emulate this formation pattern are 1994, 1995, 1996 and 2000. Maps of their formation location may be found in Appendix B (page 142); figures 9.9 to 9.12. Further discussion regarding child eddy formation may be found in the cyclone splitting section of Cyclone Lifestyles (page 112).

Other years display a different picture in that a few parent cycloic eddies are formed on the continental shelf itself. Figure 5.11 shows a representative example of this coastal development offshore of Lamberts Bay and Port Nolloth. Again there is the atypical parent cycloic eddy formed within the Cape Basin itself. Other years displaying formation on the continental shelf are 1998, 1999 and 2001. Maps depicting formation locations found during these years may be found in Appendix B; figures 9.13 to 9.15.

A note-worthy pattern to emerge from the formation location of the cycloic eddies is a discernable split in formation at 30 °S (figure 5.12). Within the northern group, only five child cycloic eddies were formed and all of the splits occurred before the Mid-Atlantic Ridge. Interestingly, the child cycloic eddies formed along the Walvis Ridge in the northern group seem to be associated with the deepest point in the Walvis Ridge. To speculate a little at this point would be to suggest that this formation process could indicate that the Walvis Ridge potentially slows the parent cycloic eddies down, allowing for the creation of a child cyclone, but doesn't have enough of a drag on the parent cyclones to reduce their energy sufficiently to prohibit future child formation.

Within the southern group, these splits occurred between 30 and 32 °S, following a path that crossed the Mid-Atlantic Ridge into the Atlantic Ocean Basin. Although the topography seemed to create a barrier, encouraging eddy splitting, it did not seem to impede all cyclonic trajectories. In addition, due to the increased energetic activity in the lower portion of the study area (Boebel et al., 2003), cycloic eddies formed south of 30 °S may have more energy and therefore a greater ability to propagate and transverse obstacles.
Figure 3.12: Northern versus southern cyclonic eddy formation (1995 - 2002). The black circles and crosses represent the northern portion of the study area, while the red circles and crosses represent the southern portion.

As mentioned, the westerly winds are thought to play an effect on wind rotation as far north as 30 °S (Shannon and Nelson, 1996). North of this latitude, upwelling cells such as the Central Namibia and Lüderitz cells are known to upwell perennially, peaking in late winter (Shannon and Nelson, 1996). South of 30 °S, the Columbine and Cape Peninsula upwelling cells are documented as occurring more seasonally than those in the north in Shannon and Nelson (1996). Additionally, the zonally orientated fronts are believed to develop to the north of the perennial Lüderitz upwelling cell and the peak summer Cape Peninsula and Columbine cells.

Hence a northern versus southern pattern should emerge from the seasonal parent cyclonic eddy data if cyclonic eddy formation is affected by coastal upwelling. These questions will be addressed within the next section concerning cyclonic eddy formation frequency.
Therefore, to summarise, parent cyclonic eddy formation appears to be mainly along the West Coast continental shelf, while child cyclonic eddy formation occurs to the west of the shelf-defining 4000 m isobath. Additionally, there may be a possible link between formation location and renowned upwelling cells within the region. Having established the formation location of these cyclonic eddies; the next section will investigate their frequency of formation.
5.2 Frequency of formation of cyclonic eddies

When considering the frequency formation of cyclonic eddies within this study it is important to remember that 1992 does not constitute a full year of data. It will therefore not be considered when contemplating formation frequency of either parent or child cyclonic eddies, as shown in figure 5.13. Additionally, as parent cyclonic eddies are able to split to form child cyclonic eddies at any point in their existence and may survive for longer than one year, it is possible, for instance, that some parent cyclonic eddies formed in 2001 would be able to produce child cyclonic eddies in 2002 and beyond. With a satellite data gap of February 2002, 2001 child cyclonic eddy formation frequency has been excluded from this portion of the analysis to prevent potential data bias.

![Annual formation frequency of cyclonic eddies (1993 - 2001)](image)

*Figure 5.13: Annual formation frequency of cyclonic eddies (1993 - 2001). The blue bars represent parent cyclonic formation frequency and the red diamonds represent child formation frequency.*

The formation frequency of 98 parent and 150 child cyclonic eddies formed between January 1993 and December 2001 is considered. A descriptive analysis of the annual formation frequency determines a mean of 10.89 and 16.67 for parent and child cyclonic eddies respectively. For parent cyclonic eddy frequency data the standard error is 0.56 and the standard deviation is 1.69, with a confidence level of 1.30. For the child cyclonic eddy frequency data the standard error is 1.5 and the standard deviation is 4.5, with a confidence level of 3.459.
As depicted in figure 5.13, 2000 produced the fewest number of parent cyclonic eddies (7), while thirteen were created in 1997, resulting in an inter-annual variability of only six over the study period of nine years. Annual child cyclonic eddy formation seemed to be similar, ranging between fifteen created in 2000 and twenty-two produced in 1993.

To determine whether the number of cyclonic eddies formed per year deviated from the expected annual rate over the study period, a chi-square goodness of fit test was applied to the discrete data (represented in table 5.3). Due to the nature of a chi-square test, data normality is not important. When considering parent and child cyclonic eddy formation rates, the null hypothesis (H₀) states that there is no difference in formation frequency between years. Naturally the alternative hypothesis (Hₐ) states that there is a difference.

Table 5.3: Annual formation frequency of parent and child cyclonic eddies between 1993 and 2000

<table>
<thead>
<tr>
<th>Year</th>
<th>Observed number of parent cyclones</th>
<th>Observed number of child cyclones</th>
</tr>
</thead>
<tbody>
<tr>
<td>1993</td>
<td>10</td>
<td>22</td>
</tr>
<tr>
<td>1994</td>
<td>12</td>
<td>19</td>
</tr>
<tr>
<td>1995</td>
<td>12</td>
<td>16</td>
</tr>
<tr>
<td>1996</td>
<td>11</td>
<td>19</td>
</tr>
<tr>
<td>1997</td>
<td>13</td>
<td>19</td>
</tr>
<tr>
<td>1998</td>
<td>11</td>
<td>17</td>
</tr>
<tr>
<td>1999</td>
<td>11</td>
<td>17</td>
</tr>
<tr>
<td>2000</td>
<td>7</td>
<td>15</td>
</tr>
<tr>
<td>2001</td>
<td>11</td>
<td>n/a</td>
</tr>
<tr>
<td>Totals</td>
<td>98</td>
<td>144</td>
</tr>
</tbody>
</table>

In the case of annual parent cyclonic eddy formation, the chi-square test statistic ($\chi^2$) was calculated as 2.10204 (df = 8; P-value = 0.9777; Power = 0.5236). Using a predetermined critical alpha ($\alpha$) of 0.05, we fail to reject the null hypothesis. There is no significant inter-annual difference between parent cyclonic eddies. Since a result of not statistically significant is obtained, it is important to consider the power of the chi-square test on these data with respect to committing a Type I ($\alpha$) or Type II ($\beta$) error.
The power (1-β error probability) of the chi-square goodness of fit test is calculated in relation to the specific degrees of freedom, sample size (n) and effect size (w = medium effect = 0.3). G*Power (version 3.0.3) software was used in this study. Using the standard α error probability of 0.05, a medium effect size (w) of 0.3, total sample size of 98 and degrees of freedom of 8, the following parameters in Table 5.4 may be calculated:

Table 5.4: Central and non-central parameters of parent cyclonic eddy annual formation frequency power test

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Non-centrality parameter λ</td>
<td>8.9200</td>
</tr>
<tr>
<td>Critical $\chi^2$</td>
<td>15.5073</td>
</tr>
<tr>
<td>Power (1-β error probability)</td>
<td>0.5236</td>
</tr>
</tbody>
</table>

Figure 5.14 shows a distribution plot of the critical chi-square value for these parent formation frequency data. This graphical representation clearly shows the expected increase in an α error with an increase in the critical chi-square value. Also noticeable is the possible β error range within this sample.

![Figure 5.14: Distribution plot of the power of a chi-square test on annual frequency of formation parent cyclonic eddy data](image)

From Figure 5.14 it is possible to observe that the critical $\chi^2$ of 15.507 lies much further along the x-axis than the calculated test parent $\chi^2$ value of 2.10204. Therefore there is a low probability that an α error has been committed within this test. Equally noticeable is that the calculated test parent $\chi^2$ value of 2.10204 is situated at the edge of the curve of a possible β error, implying a minimal risk that a Type II error has been committed. Therefore, with a test Power of 0.5236 (Table 5.4), it is possible to state that the chi-square goodness of fit test is an appropriate statistical test to determine annual formation frequency significance of parent cyclonic eddies.
When considering annual child cyclonic eddy formation, the calculated chi-square test statistic returned a result that was also not statistically significant ($df = 7, \chi^2 = 2.8933, P-value = 0.8947, Power = 0.7499$). Again we fail to reject the null hypothesis and are able to state that there is no significant inter-annual difference between child cyclonic eddies. Therefore, within parent and child annual formation, the observed values are not significantly different from the expected value, given the null hypothesis and random chance. This allows for the assumption that cyclonic eddies production rates are not enhanced by extraneous, infrequent environmental effects that may be apparent over a decade, such as El Niño or La Niña for example.

The incidence of child cyclonic eddy formation is expected to be higher than parent cyclonic eddy formation due to the very dynamic nature of the region in which this study takes place (Boebel et al., 2003), as well as the actual formation method of the child cyclonic eddies. This was found to be the case in that there were fifty-two more child cyclonic eddies identified than possible parent cyclonic eddies. This 1:1.5 ratio proves unequivocally that some parent cyclones produce more than one child.

Hence, there does not seem to be a correlation between the numbers of cyclonic eddies formed each year. Inter-seasonality may provide insight into any possible significant parent and child cyclonic eddy association.

As with annual formation frequency, 1992 formation data were excluded from this portion of the analysis, as were 2001 seasonal frequency data from the child cyclonic eddy data set. Formation frequency data were divided into the following southern hemisphere seasonal categories: spring (September – November); summer (December – February); autumn (March – May) and winter (June – August). Seasons were defined from Schouten et al. (2000) and de Ruijter et al. (1999).

With thirty parent cyclonic eddies, summer ranked as the season with the highest number of eddy formations between 1993 and 2001 (figure 5.15). Autumn and winter’s twenty-five and twenty-four followed closely behind. Spring recorded the lowest formation frequency at nineteen. With a formation mean of 24.5 parent cyclonic eddies, the seasonal frequency of parent cyclonic eddies seems to be constant.
As found off the west coast of the United States (Harney et al., 2001), the slight increase in production over the summers may potentially be attributed to coastal upwelling driven by the prevailing summer winds (Shannon and Nelson, 1996). This is particularly applicable to those cyclonic eddies formed in association with the Columbine and Cape Peninsula Upwelling cells, whereas the northern upwelling cells are far more consistent and less seasonally driven.

Statistical analysis shows a parent seasonal formation median of 24.5, with a standard error of 2.25 and a standard deviation of 4.51. To determine whether there is any significant variability in seasonal formation frequency, the chi-square goodness of fit test was deemed the appropriate test to use. Table 5.5 contains the seasonal parent and child formation frequency data. As before, the null hypothesis states that there is no difference between seasonal cyclonic eddy formations, while the alternative hypothesis suggests that there is a difference.

In the case of parent cyclonic eddy formation, the chi-square $\chi^2 = 2.1020 \ (df = 3, P-value = 0.4771, \alpha = 0.05, Power = 0.7015)$. This result is not statistically significant; therefore it is necessary to fail to reject the null hypothesis. Therefore, within parent seasonal formation, the observed values are not significantly different from the expected value, given the null hypothesis and random chance.
Table 5.5: Seasonal formation frequency of parent and child cyclonic eddies

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Spring (Sep - Nov)</td>
<td>19</td>
<td>35</td>
</tr>
<tr>
<td>Summer (Dec - Feb)</td>
<td>30</td>
<td>29</td>
</tr>
<tr>
<td>Autumn (Mar - May)</td>
<td>25</td>
<td>33</td>
</tr>
<tr>
<td>Winter (Jun - Aug)</td>
<td>24</td>
<td>47</td>
</tr>
<tr>
<td>Totals</td>
<td>98</td>
<td>144</td>
</tr>
</tbody>
</table>

Child cyclonic eddy conception seems to vary more per season than their parents. The 1993 to 2000 winters experienced forty-seven child cyclones, while thirty-five were recorded in spring. The autumn and summer seasons of the study period observed a drop in production rate down to the early thirties at thirty-three and twenty-nine respectively (figure 5.15).

Statistical analysis shows a child seasonal mean formation of 36 cyclonic eddies, with a standard error of 3.87 and a standard deviation of 7.75. Assuming the same hypothesis as above, the calculated chi-square test applied to the formation frequency data for seasonal child cyclonic eddies (table 5.5) returned a test result of not statistically significant \( (df = 3, \chi^2 = 5.000, P\text{-value} = 0.1718, \text{Power} = 0.8696) \). Again we fail to reject the null hypothesis. Therefore, within both parent and child seasonal formation, the observed values are not significantly different from the expected value, given the null hypothesis and random chance.

Hence both tests prove that there is no significant difference of formation frequency for parent or child cyclonic eddies between the seasons. However, the increase in child cyclonic eddy formation over the cooler winter and residual spring months may be possibly attributed to the effect of commonly occurring winter storms on the parent cyclonic eddies. These storms amplify the stirring and mixing of the region’s waters, creating more potential for eddy splitting. This hypothesis may be further explored by considering the formation frequency of cyclonic eddies on a monthly scale.
Figure 5.16: Monthly formation frequency of cyclonic eddies. As per standard monthly nomenclature, month 1 to 12 represent January to December respectively. The blue bars represent parent cyclonic formation frequency and the red diamonds represent child formation frequency.

Figure 5.16 represents the monthly formation frequency of the 1993 - 2001 parent and 1993 - 2000 child cyclonic eddy data set. What is apparent is the potential randomness of monthly formation frequency of both parent and child cyclonic eddies. Here it is possible to perceive that the month of September had the lowest creation rate for parent cyclonic eddies, while February months observed the most. Production rates dropped during May, August and September, rising dramatically in January and July. Child cyclonic eddy monthly formation frequency demonstrated the opposite pattern to their parents. As may be seen from the 1993 - 2000 data in figure 5.16, child eddy formation was clearly greatest (23) in the month of August, dipping to a low of seven in both April and November.

Statistically, parent cyclonic eddy formation had a mean formation of 8.08 per month, a standard error of 0.75 and a standard deviation of 2.61. Mean child cyclonic eddy monthly formation was higher at 13.0, with a standard error of 1.39 and a standard deviation of 4.82. A chi-square goodness of fit statistical test produces the following results from the data shown in table 5.6. As before, for both parent and child cyclonic eddies, the null hypothesis indicates that there is no difference between monthly formation frequencies, whereas the alternative hypothesis reports that there is.
Table 5.6: Monthly formation frequency of parent and child cyclonic eddies

<table>
<thead>
<tr>
<th>Month</th>
<th>Observed number of parent cyclones</th>
<th>Observed number of child cyclones</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>11</td>
<td>8</td>
</tr>
<tr>
<td>February</td>
<td>12</td>
<td>10</td>
</tr>
<tr>
<td>March</td>
<td>9</td>
<td>15</td>
</tr>
<tr>
<td>April</td>
<td>11</td>
<td>7</td>
</tr>
<tr>
<td>May</td>
<td>5</td>
<td>14</td>
</tr>
<tr>
<td>June</td>
<td>8</td>
<td>14</td>
</tr>
<tr>
<td>July</td>
<td>10</td>
<td>13</td>
</tr>
<tr>
<td>August</td>
<td>5</td>
<td>23</td>
</tr>
<tr>
<td>September</td>
<td>4</td>
<td>19</td>
</tr>
<tr>
<td>October</td>
<td>7</td>
<td>15</td>
</tr>
<tr>
<td>November</td>
<td>8</td>
<td>7</td>
</tr>
<tr>
<td>December</td>
<td>7</td>
<td>11</td>
</tr>
</tbody>
</table>

For monthly parent cyclonic eddy formation frequency (table 5.6), the chi-square test statistic was not statistically significant ($df = 11, \chi^2 = 9.2680, P-value = 0.5972, \textit{Power} = 0.4565$). A chi-square goodness of fit test applied to the monthly child cyclonic eddy formation data (table 5.6), however, returns a different result, allowing the null hypothesis to be rejected ($df = 11, \chi^2 = 19.6023, P-value = 0.04974$). Therefore, within child monthly formation, the observed values are significantly different from the expected value, given the null hypothesis and random chance.

With no extreme annual, seasonal or monthly formation frequency, the next logical step would be to look at location as a function of formation frequency. Figure 5.17 depicts the overall number of parent cyclonic eddies formed to the north and south of 30°S. There does seem to be a difference in formation location across the board, with the southern region emerging as consistently more productive than the north.

1993, 1995, 1996, 1999 and 2001 displayed a much larger disparity (figure 5.17) between regional formations. During 1993 and 1995 the southern region produced double the number of parent cyclones than the northern region. 1996, 1999 and 2001 saw a manufacture of more than twice in the south than in the north. Production in both regions was closest during 1994 and 1997, when northern production rates were slightly higher than during other years; and 2000, where southern formation of parent cyclonic eddies seemed to be fewer than during the rest of the study period.
Ultimately, at ninety-eight, more parent cyclonic eddies were formed in the southern region than in the north, where fifty-three were formed. It may be possible that this is due to the fact that the southern region is subjected to passing Agulhas Rings and other mesoscale eddies from the Agulhas Retroflection area. It has been repeatedly proven (Boebel et al., 2003) that this region is highly turbulent and prone to intense mixing and stirring, creating a platform for increased cyclonic eddy formation as shown in the results above.

There seemed to be large difference between northern and southern formation frequency for child cyclonic eddies, as depicted by the 1993 to 2000 data in figure 5.18. Due to the dynamic nature of the southern region of this study area (Boebel, 2003), it is to be expected that oceanic cyclones passing through this region will be subjected to a larger amount of turbulence and inter-mingling with other mesoscale features (Richardson and Garzoli, 2003) than in the northern region. This mixing and stirring is likely to result in cyclone splitting, the creation of child cyclonic eddies, and from figure 5.18, was obviously prevalent south of 30°S.
For every year bar 1999, child cyclonic eddy formation rate in the south was twice, if not triple the rate of formation in the north. In 1999, although the number of cyclonic eddies formed were not as dramatically different as in the other years, they still followed the same pattern as above.

![Frequency of cyclonic eddy child formation](image)

*Figure 5.18: Northern (north of 30°S) versus southern (south of 30°S) annual formation frequency of child cyclonic eddies (1993–2001). The blue bars represent northern formation frequency and the pink stems represent southern formation frequency.*

To determine the relationship between parent and child cyclones in the two regions, formation frequency was graphed as in figure 5.19 (northern region) and 5.20 (southern region). Interestingly, in the north fewer child cyclonic eddies were generally formed than parent cyclonic eddies. In fact, in some cases, parent cyclonic eddy formation was double that of child formation. During 2000, an even number of parent and child cyclonic eddies were recorded.

Indulging in speculation, this lack of child eddy formation could potentially indicate that the northern region of the study area is a more stable region west of the continental shelf than in the south, potentially due to fewer topographical obstacles, as well as being north of the turbulent Agulhas Retracking region (Boebel et al., 2003). In the north, the Walvis Ridge is also closer to the continental shelf than in the south. Perhaps this shorter distance allows the newly formed parent cyclones to reach...
the Walvis Ridge with a stronger structure than those in the south, potentially increasing their ability to cross the Ridge without too much loss of energy or child formation probability.

However, in 1993 and 1999, this was not the case. During 1993, two more child cyclones were formed than parent cyclones. 1999 represents a very obvious exception to the norm in that six more child cyclonic eddies were formed than parents. This exception was also represented in figure 5.18 above.

![Chart showing frequency of cyclonic eddy formation north of 30°S](image)

**Figure 5.19.** Northern (north of 30°S) parent versus child cyclonic eddy annual formation frequency (1993 - 2001). The blue bars represent parent formation frequency and the pink sticks represent child formation frequency.

As may be seen from figure 5.20, 56 more child cyclonic eddies were formed than parents in the south during the study period. In fact, 1993, 1994, 1997, 1998 and 2000 saw the production of three times the number of child cyclonic eddies than their parents. On average, 12.63 child cyclonic eddies were formed annually during the study period from a mean of 5 parent cyclonic eddies.
Figure 5.20: Southern (south of 30°S) parent versus child cyclonic eddy annual formation frequency (1993 - 2001). The blue bars represent parent formation frequency and the pink signs represent child formation frequency.

Figure 5.21: Northern versus southern annual formation frequency.
While the formation frequency of parent eddies remained relatively constant in both the northern and southern regions, figure 5.21 depicts the phenomenon whereby the formation frequency of child cyclonic eddies was drastically higher in the south than in the north. Of importance when regarding formation frequency within the two regions is the realisation that the number of child cyclonic eddies formed are directly related to the number of parent cyclonic eddies present. For example, fewer parent cyclonic eddies should produce fewer child cyclonic eddies. This realisation is perfectly graphed in figure 5.21.

What is particularly apparent in figure 5.21 is the interesting formation frequency event that occurred in the north in 1999, whereby four more child cyclonic eddies were formed than the next most productive year in that data series. Both of these associations are statistically tested below.

Table 5.7: Regional formation frequency of parent and child cyclonic eddies

<table>
<thead>
<tr>
<th>Year</th>
<th>Observed number of parent cyclones formed in the north</th>
<th>Observed number of parent cyclones formed in the south</th>
<th>Observed number of child cyclones formed in the north</th>
<th>Observed number of child cyclones formed in the south</th>
</tr>
</thead>
<tbody>
<tr>
<td>1993</td>
<td>5</td>
<td>5</td>
<td>7</td>
<td>15</td>
</tr>
<tr>
<td>1994</td>
<td>8</td>
<td>4</td>
<td>5</td>
<td>14</td>
</tr>
<tr>
<td>1995</td>
<td>6</td>
<td>6</td>
<td>3</td>
<td>13</td>
</tr>
<tr>
<td>1996</td>
<td>5</td>
<td>6</td>
<td>3</td>
<td>16</td>
</tr>
<tr>
<td>1997</td>
<td>9</td>
<td>4</td>
<td>6</td>
<td>13</td>
</tr>
<tr>
<td>1998</td>
<td>7</td>
<td>4</td>
<td>4</td>
<td>13</td>
</tr>
<tr>
<td>1999</td>
<td>5</td>
<td>6</td>
<td>11</td>
<td>6</td>
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<tr>
<td>2000</td>
<td>4</td>
<td>3</td>
<td>4</td>
<td>11</td>
</tr>
<tr>
<td>2001</td>
<td>4</td>
<td>7</td>
<td>n/a</td>
<td>n/a</td>
</tr>
<tr>
<td>Totals</td>
<td>53</td>
<td>45</td>
<td>43</td>
<td>101</td>
</tr>
</tbody>
</table>

Descriptive statistics of these data show that parent cyclonic eddies were formed at a mean rate of 5.89 and 5.00 in the north and south respectively, with those formed in the north producing a standard error (at 0.59) that is slightly higher than in those in the south (at 0.44). The standard deviations follow this trend with 1.76 in the north and 1.32 in the south. Northern child cyclonic eddies emulated their parents with a mean formation rate of 5.38, standard error of 0.94 and standard deviation of 2.67. Their southern cousins, however, showed a much higher average rate of formation at 12.63 per annum. A standard error of 1.08 and a standard deviation of 3.07 are associated with this mean.
Using a standard chi-square test for independence on the data presented in table 5.7 it is possible to ascertain that there is no significant association between parent cyclonic eddies formed within the north and the southern parts of the study area \((df = 8, \chi^2 = 4.595, P-value = 0.7999, Power = 0.5240)\). A chi-square test for trend shows that, with an \(\chi^2 = 0.5231, P-value = 0.4695\), there is not a significant linear trend among the annual categories, given the null hypothesis of no difference and random chance. Therefore, it is possible to state that there is no significant difference between the numbers of parent cyclonic eddies formed within the northern verses the southern portions of the study area.

When considering the child formation frequency per location (table 5.7), the same statistical tests are used. A chi-square test for independence returns a result of no significant association between child cyclonic eddies formed within the north and the southern parts of the study area \((df = 7, \chi^2 = 13.174, P-value = 0.0680, Power = 0.7499)\). A chi-square test for trend shows that an \(\chi^2 = 1.588, P-value = 0.2077\) also does not return a significant linear trend among the annual categories, given the null hypothesis of no difference and random chance. Again it is possible to qualify the statement that there is no significant difference between the number of child cyclonic eddies formed within the north compared to the south of the study area.

Nonetheless, cyclonic eddy formation may be encouraged by upwelling along the African continental shelf (Boebel et al., 2003). To examine this statement, seasonal frequency formation data were separated into a northern and southern portion of the study area at 30 °S. This latitude was chosen as the dividing line for a number of reasons. The first reason was the clear delineation between current speeds and directions in the South Atlantic along 30 °S (Fu, 2006). Fu found that north of 30 °S water moves at an average of 4 – 6km/day in a north-westerly direction, while water south of 30 °S moves at a sedate 0 – 2 km/day towards the east, potentially marking the north and south branches of the South Atlantic subtropical gyre.

Another motivation for the 30 °S data separation was that the 35 – 45 °S westerly winds is thought to effect wind rotation up to 30 °S (Shannon and Nelson, 1996). North of this latitude the perennial Central Namibia and Lüderitz upwelling cells are seen to peak in late winter (Shannon and Nelson, 1996). South of 30 °S, the seasonal
Columbine and Cape Peninsula upwelling cells are dominant in summer (Shannon and Nelson, 1996). Hence a northern versus southern pattern should emerge from the seasonal parent cyclonic eddy data if cyclonic eddy formation is affected by coastal upwelling.

![Seasonal frequency of northern versus southern parent cyclonic eddy formation (1993 - 2001)](image)

**Figure 5.22:** Seasonal frequency of northern versus southern parent cyclonic eddy formation (1993 - 2001).

As is visible from figure 5.22, formation frequency seemed to be the highest in summer during the study period. This phenomenon ties in nicely with the theory that the associated summer upwelling maximum in both the north and the south drives parent cyclonic eddy formation. Relatively high formation in the autumn and winter also corroborate the idea that passing anticyclonic eddies and winter storms encourage cyclonic eddy creation in the south, while the perennial upwelling cells of the north maintain development in the north.

Child cyclonic eddy formation, while showing a completely different seasonal pattern (figure 5.23), showed the validity of the above parent cyclonic eddy statement. Here, the lowest number of child cyclonic eddies were formed during summer in the north. This may potentially be due to the persistent offshore winds providing a stabilising effect on the surrounding water, minimizing mixing and stirring and therefore
reducing the potential formation of child cyclonic eddies. On the other hand, in the south, a far higher number of child cyclonic eddy were formed in the winter season. This may be directly attributed to the winter storms of the season, creating an environment that is so turbulent and unstable that cyclonic eddies split and merge at an amplified rate. However, a statistical approach is necessary to determine numerical significance.

![Seasonal frequency of northern versus southern child cyclonic eddy formation](image)

*Figure 5.23: Seasonal frequency of northern versus southern child cyclonic eddy formation (1993 - 2000).*

A chi-square test for independence on table 5.8's data determines that there is no significant correlation between seasonal parent formation frequency within the north and the south ($df = 3, x^2 = 3.077, P-value = 0.3800, Power = 0.7015$). A chi-square test for trend results in an $x^2 = 2.220, P-value = 0.1362$, meaning that there is no significant linear trend among the seasonal categories, given the null hypothesis of no difference and random chance. Therefore, it is possible to state that there is no significant difference between the seasonal formation frequencies of parent cyclonic eddies formed within the northern and the southern portions of the study area.
Table 5.8: Regional seasonal formation frequency of parent and child cyclonic eddies

<table>
<thead>
<tr>
<th>Season</th>
<th>Observed number of parent cyclones formed in the north</th>
<th>Observed number of parent cyclones formed in the south</th>
<th>Observed number of child cyclones formed in the north</th>
<th>Observed number of child cyclones formed in the south</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spring</td>
<td>7</td>
<td>12</td>
<td>14</td>
<td>21</td>
</tr>
<tr>
<td>Summer</td>
<td>17</td>
<td>13</td>
<td>6</td>
<td>23</td>
</tr>
<tr>
<td>Autumn</td>
<td>14</td>
<td>11</td>
<td>13</td>
<td>20</td>
</tr>
<tr>
<td>Winter</td>
<td>15</td>
<td>9</td>
<td>10</td>
<td>37</td>
</tr>
<tr>
<td>Totals</td>
<td>53</td>
<td>45</td>
<td>43</td>
<td>101</td>
</tr>
</tbody>
</table>

Child seasonal formation frequency per location (table 5.8) chi-square test for independence concludes that there is no significant association between child cyclonic eddies formed within the northern and southern parts of the study area \((df = 3, \chi^2 = 5.968, P-value = 0.1132, Power = 0.8696)\). A chi-square test for trend shows that an \(\chi^2 = 1.738\) and \(P-value = 0.1874\) does not constitute a significant linear trend among the seasonal categories, given the null hypothesis of no difference and random chance. Therefore there is no significant difference between the seasonal formation frequencies of child cyclonic eddies formed within the north and south regions.

As shown, there did not seem to be any significant signal in parent cyclonic eddy formation frequency on an annual, seasonal or monthly time scale. Apart from a significantly different monthly formation frequency, the same may be said for child cyclonic eddy formation. This relationship between parent and child eddy formation begs the question as to whether there is a time lag correlation between the formation of a parent and subsequent child cyclonic eddy.

A time lag present a complex correlation because child cyclonic eddies may be formed from parent cyclonic eddies; as well as sibling child cyclonic eddies. To explain, within one cyclonic eddy family a parent eddy may spawn numerous child eddies and these child cyclonic eddies may in turn spawn other child cyclonic eddies. Graphically these data groups are represented in figure 5.24, named with a nomenclature that indicates the order of their formation. Due to the multiple groups, complexities of cyclonic eddy formation and subsequent splitting events, overall groups as per order of formation have been included in figure 5.24. Individual group graphs are represented numerically below in table 5.8 and 5.9. Child cyclonic eddies formed from parent and sibling origin may be found in figures 5.25 to 5.29 below.
Figure 5.24: Time lag in days between the formation of cyclonic eddies and their subsequent splitting events.

Table 5.8 shows the descriptive statistics associated with all the child cyclonic eddies formed from parent eddies, with the subject nomenclature indicating the order in which the child cyclonic eddy was formed. Table 5.9 shows the same descriptive statistics but this time of the child cyclonic eddies formed from siblings, again with the subject nomenclature indicating the order in which the child cyclonic eddy was formed.

Table 5.8: Statistical overview of the time lag between the formations of parent cyclonic eddies and the splitting event that subsequently created a child cyclonic eddy (as measured per day). 1st child eddies represent those first formed from the parent. 2nd child eddies represent the second child eddy formed from a parent cyclonic eddy within a family etc.

<table>
<thead>
<tr>
<th>Statistics for parents and child cyclonic eddies</th>
<th>1st child eddies</th>
<th>2nd child eddies</th>
<th>3rd child eddies</th>
<th>4th child eddies</th>
<th>5th child eddies</th>
<th>6th child eddies</th>
<th>All 7th Child eddies</th>
<th>All 8th Child eddies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>133.05</td>
<td>217.36</td>
<td>247.45</td>
<td>418.25</td>
<td>574.06</td>
<td>445.07</td>
<td>351.75</td>
<td>n/a</td>
</tr>
<tr>
<td>Standard Error</td>
<td>15.44</td>
<td>19.19</td>
<td>16.87</td>
<td>35.74</td>
<td>44.89</td>
<td>41.89</td>
<td>217.93</td>
<td>122.96</td>
</tr>
<tr>
<td>Standard Deviation</td>
<td>12.15</td>
<td>16.09</td>
<td>16.87</td>
<td>17.47</td>
<td>19.80</td>
<td>19.80</td>
<td>377.50</td>
<td>245.32</td>
</tr>
<tr>
<td>Range</td>
<td>600</td>
<td>448</td>
<td>600</td>
<td>253</td>
<td>253</td>
<td>253</td>
<td>714</td>
<td>n/a</td>
</tr>
<tr>
<td>Minimum</td>
<td>7</td>
<td>21</td>
<td>28</td>
<td>28</td>
<td>28</td>
<td>28</td>
<td>28</td>
<td>119</td>
</tr>
<tr>
<td>Maximum</td>
<td>616</td>
<td>469</td>
<td>798</td>
<td>490</td>
<td>588</td>
<td>714</td>
<td>644</td>
<td>244</td>
</tr>
<tr>
<td>Count</td>
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<td>23</td>
<td>11</td>
<td>4</td>
<td>2</td>
<td>3</td>
<td>4</td>
<td>1</td>
</tr>
<tr>
<td>Confidence Level (95.0%)</td>
<td>30.69</td>
<td>44.14</td>
<td>113.34</td>
<td>177.27</td>
<td>177.27</td>
<td>937.77</td>
<td>390.27</td>
<td>n/a</td>
</tr>
</tbody>
</table>
Table 5.9: Statistical overview of the time lag between the formations of child cyclonic eddies and their splitting (i.e., created another child cyclonic eddy) as measured per days. 2nd child eddies represent those formed from the first child cyclonic eddy within a family of eddies. 3rd child eddies represent the third child eddy formed from another sibling child in sequence etc.

<table>
<thead>
<tr>
<th>Statistics for child and child cyclonic eddies</th>
<th>2nd child eddies</th>
<th>3rd child eddies</th>
<th>4th child eddies</th>
<th>5th child eddies</th>
<th>6th child eddies</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>94.18</td>
<td>109.10</td>
<td>87.40</td>
<td>104.83</td>
<td>114.33</td>
</tr>
<tr>
<td>Standard Error</td>
<td>21.01</td>
<td>28.54</td>
<td>22.26</td>
<td>18.60</td>
<td>22.26</td>
</tr>
<tr>
<td>Standard Deviation</td>
<td>69.67</td>
<td>90.25</td>
<td>70.40</td>
<td>45.35</td>
<td>38.35</td>
</tr>
<tr>
<td>Range</td>
<td>210</td>
<td>273</td>
<td>259</td>
<td>140</td>
<td>77</td>
</tr>
<tr>
<td>Minimum</td>
<td>7</td>
<td>7</td>
<td>14</td>
<td>28</td>
<td>77</td>
</tr>
<tr>
<td>Maximum</td>
<td>217</td>
<td>280</td>
<td>273</td>
<td>168</td>
<td>154</td>
</tr>
<tr>
<td>Count</td>
<td>11</td>
<td>16</td>
<td>16</td>
<td>6</td>
<td>3</td>
</tr>
<tr>
<td>Confidence Level (95.47%)</td>
<td>46.81</td>
<td>64.56</td>
<td>50.36</td>
<td>47.80</td>
<td>95.77</td>
</tr>
</tbody>
</table>

From the ranges of all groups in both tables 5.8 and 5.9, as well as the large standard deviations, it is possible to assume that there is no direct correlation between the time lag of formation between cyclonic eddies. Due to data non-normality, a non-parametric Kruskal-Wallis (KW) test is applied to determine whether there is any significance between the mean time lags of each group of child cyclonic eddy. The Kruskal-Wallis test results in a P-value of 0.0609 (H = 13.496, Power = 90.11), which is considered not statistically significant, given a null hypothesis of no difference. Therefore it is not possible to reject the null hypothesis, allowing for the assumption that there is no significant difference between the means of each group of child eddy formation, beyond that expected by chance. It is important to note that the Kruskal-Wallis test approximated the P-value (from the chi-square distribution) because some groups had duplicate time lag values. However, some interesting patterns emerge from figures 5.25–5.29 below.

![Time lag in days of 2nd child eddies from parents](image1)

![Time lag in days of 2nd child eddies from other children](image2)

*Figure 5.25: Time lag in days (to-date) between formation of the second child cyclonic eddy from parent and other child cyclonic eddies respectively.*

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As seen from table 5.8, there are more child cyclonic eddies formed as first formations (sixty-one) than from secondary splitting events (thirty-six in total for both parent and sibling splitting groups). This condition is applicable throughout the sequential formations. Another rather obvious pattern to emerge was that in the higher groups in the formation sequence (i.e. first to second child cyclonic eddies) more child cyclonic eddies were formed from parent cyclonic eddies than from siblings. In this example, 25 child cyclonic eddies emerging from second splitting events were created by parent cyclonic eddies against 11 child cyclonic eddies formed from siblings.

The third child cyclonic eddy group seemed to show similar formation rates between parent and sibling child origins. For groups four onwards more child cyclonic eddies were formed from sibling child cyclonic eddies than original parent cyclonic eddies. For example, within group five (figure 5.28), six child cyclonic eddies were formed from siblings, with only two formed from parents during the study period. This would seem logical due to the greater possibility of formation from the higher proportion of child cyclonic eddies available for splitting.
Another interesting revelation is that child cyclonic eddies formed from other child cyclonic eddies do not have as long a lifespan as those formed from parent cyclonic eddies, regardless of their sequence in formation from the original parent cyclonic eddy. This phenomenon could be due to the splitting process, whereby parent cyclonic eddies release energy into a child cyclonic eddy as it coalesced from its parent. Unless the energy was introduced from an outside source (other eddies), subsequent child cyclonic eddies splitting from a sibling would potentially divide that energy again (depending on the size of the original child cyclonic eddy). As seen from figure 5.28 and 5.29, child cyclones formed from siblings had a smaller average lifespan than child cyclonic eddies formed directly from the parent cyclone.

The most interesting result to emerge from the data presented above is the impact on formation that location has. It appears that both the Continental Slope and wind-driven upwelling cells that predominate along the West Coast of Southern Africa appear to be related to the formation of cyclonic eddies in the Cape Basin.
To recap, there does not appear to be any significant difference in frequency of formation between parent and child cyclonic eddy annual and seasonal frequency. Monthly parent cyclonic eddy frequency is also not statistically significant, however, child cyclonic eddies do appear to differ in frequency rates on a monthly scale. When separated into those formed in the northern and southern regions of the study area, results that were not statistically significant were returned for parent and child formation on an annual and seasonal basis. Due to the nature of formation of a child cyclonic eddy, the time lags between the formations of the parent cyclonic eddies and subsequent splitting events (child cyclonic eddies) were analysed for possible correlations. However, all events returned a result of not statistically significant, allowing for the assumption that there is no statistical difference in time lags between child formation events.

Following on from location and frequency of formation of the cyclonic eddies under discussion would be their size at formation. The next section attempts to determine the necessary size, if any, of formation needed for these cyclonic eddies to move away from their formation location.
5.3 Formation size of cyclonic eddies

An important point to mention whilst considering the sizes of cyclonic eddy at formation, is that parent cyclonic eddies were only tracked once they had exceeded the mean sea level by more than -10 cm in the altimetric data. Hence, while it may be possible that cyclonic eddies require a certain diameter to achieve a sea surface height anomaly deeper than -10 cm, those cyclonic eddies that were smaller in diameter and therefore more shallow than -10 cm in height, were potentially overlooked in the data set.

Most parent cyclonic eddies were identified as forming with the same size and had varying life spans (covered in chapter 6.2, page 95). Additionally, the consistent size at formation (or first identification within the satellite altimetry) of cyclonic eddies removed any potential formation location effects, allowing for the assumption that parent cyclonic eddies were formed at the same size in all location within the study area.

Table 5.9: Parent and child cyclonic eddy approximate size on formation

<table>
<thead>
<tr>
<th>Latitude versus longitude (degrees)</th>
<th>Equivalent nautical miles</th>
<th>Parent cyclonic eddies</th>
<th>Child cyclonic eddies</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.5 - 2.5</td>
<td>150 x 150</td>
<td>1</td>
<td>0</td>
</tr>
<tr>
<td>2 - 2</td>
<td>120 x 120</td>
<td>5</td>
<td>6</td>
</tr>
<tr>
<td>1.5 - 1.5</td>
<td>90 x 90</td>
<td>1</td>
<td>5</td>
</tr>
<tr>
<td>1.25 - 1.25</td>
<td>75 x 75</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>1 - 1</td>
<td>60 x 60</td>
<td>70</td>
<td>82</td>
</tr>
<tr>
<td>0.75 - 0.75</td>
<td>45 x 45</td>
<td>0</td>
<td>5</td>
</tr>
<tr>
<td>0.5 - 0.5</td>
<td>30 x 30</td>
<td>3</td>
<td>24</td>
</tr>
<tr>
<td>0.25 - 0.25</td>
<td>25 x 25</td>
<td>1</td>
<td>11</td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td>83</td>
<td>135</td>
</tr>
</tbody>
</table>

Upon formation, 70% of surveyed cyclonic eddies were approximately one degree of latitude by one degree of longitude (see table 5.9) in diameter, converting to approximately 60 x 60 nautical miles (n.mi) in size. Out of 83 parent cyclonic eddies 70 were of those dimensions, whereas 82 out of 135 child cyclonic eddies complied with that statement. One parent cyclonic eddy had a diameter as large as 150 x 150 n.mi on formation. Eleven child cyclonic eddies were formed with a diameter of 25 n.mi, although six were formed as large as 120 x 120 n.mi, showing the range of formation size of child cyclonic eddies.
Regardless of their sea surface height, child cyclonic eddies were recorded as they split from their formation cyclonic eddy. During these instances they were documented with similar, as well as smaller horizontal sizes to parent cyclonic eddies, denoting the probability that a parent cyclonic eddy with a smaller horizontal size and a sea surface height of -10 cm could also have been tracked if present in the data set. Therefore it may be possible to deduce that 60 n.mi is an accurate size of diameter for the formation of most parent and child cyclonic eddies. This diameter potentially provides the eddy with sufficient strength to begin its translocation away from its formation location.

Chapter 5 has considered three aspects of formation of cyclonic eddies within the Cape Basin. Section one considered their formation location, where it was possible to determine that the majority of parent cyclonic eddies were formed along the West Coast continental shelf, while most of the child cyclonic eddies formed to the west of the 4000 m isobath. Additionally, there may be possible links between parent formation locations and known upwelling cells within the region.

In section 2 the formation frequency of these cyclonic eddies was considered. Whereas most frequencies returned results that were not statistically significant, it appears that there is a significant difference between the monthly formation frequencies of child cyclonic eddies. Section 3 showed that the size of formation of cyclonic eddies was generally a degree of longitude and a degree of latitude (60 x 60 n.mi), although the altimetric data source and its resolution accuracy should be taken into account when considering formation size of cyclonic eddies. Within chapter 6 the trajectories, life spans, physical properties and behaviour of these identified cyclonic eddies will be investigated.
CHAPTER 6

Cyclonic eddy lifestyles

6.1 Cyclonic eddy trajectories

Figure 6.1 depicts the trajectories of the 105 parent and 157 child cyclonic eddies identified between October 1992 and December 2001. As mentioned above, most cyclonic eddies appeared to form on the continental slope. Overall trajectory, as shown in figure 6.1, was generally in five directions: south-southwest, southwest, west-southwest, west, and west-northwest.

![Trajectories of 1992-2001 parent and child cyclonic eddies](image)

*Figure 6.1: Trajectories of cyclonic eddies formed between 1992-2001.*

However, the intricate nature of cyclonic eddy movement is too complex to allow for an in-depth look at cyclonic trajectories when displayed in their entirety as above. Therefore the trajectories of 95 parent cyclonic eddies have been re-plotted (figures 6.2, 6.6) between their formation location and their demise to show their overall trajectory.
Seven parent cyclones formed in the most southern region of the study area translocated in a south-southwest direction, as shown in Figure 6.2. Eighteen moved directly southwest, as shown in Figure 6.3.
Thirty of the ninety-six overall trajectories moved in a west-southwest direction (figure 6.4), meaning that the majority of parent cyclonic eddies translocated in a direction concurrent with results published by Morrow et al. (2004). However, this west-southwest direction only accounts for 31.58% of the eddies considered during this study.
The next most popular direction was westerly, with 26 cyclones as shown in figure 6.5. This direction accounts for 27.37% of overall eddy translocation. Fourteen parent cyclonic eddies followed a west-north-westerly track as shown in figure 6.6, perhaps carried along by neighbouring anticyclonic eddies.

![Parent cyclonic eddies with west-northwest trajectory](image)

Figure 6.6: Parent cyclonic eddies with west-northwest trajectories between 1992 and 2001. Stars represent the final location of individual cyclonic eddies.

Typically most cyclonic eddies seemed to migrate south-westwards when within the Cape Basin. Once across the Walvis Ridge they tended to maintain more of a westerly trajectory. However, cyclonic eddies in the Cape Basin did not travel in a straight direction, but meandered along seemingly random paths. It is therefore necessary to contemplate individual cyclonic eddy passage per year to fully understand their movement.

Figure 6.7 shows a representation of cyclonic eddies formed between October and December 1992. All eddies were formed on the continental slope and ultimately propagated in the proposed five overall directions. Cyclonic eddy 92A (in red) formed at the edge of the continental slope, just west of Lambert’s Bay. After travelling southwest, it turned north and remained within that area for a number of weeks, as an anticyclonic eddy built in intensity to the southeast. Upon resuming its journey it travelled directly south, before veering southwest again. 92A then split into two equal child cyclones, 92Ai and 92Aii.
92AII proceeded to travel in a south-westerly direction until its signal faded from the sea surface height altimetric data. 92Al did the opposite. It turned northwards and shot up two degrees of latitudes, before again veering south-westwards. Its signal disappeared from the data as it was heading in a westerly direction and it merged with another cyclonic eddy. In total, cyclonic eddy 92A and its associated children covered approximately five degrees of latitude (300 n.mi) and eleven degrees of longitude (660 n.mi), travelling in an overall southwest direction.

Cyclonic eddy 92A is representative of cyclonic eddies that followed a southwest and south-southwest overall trajectory, interspersed with northward movement as stated above. Cyclonic eddies following this translocation direction may generally be found within the southern regions of the study area, where their direction is thought to be influenced by the dynamic area in which they occur (Morrow et al., 2004; Boebel et al., 2003). In this region, they may be essentially pulled or pushed along by warm core Agulhas Rings and the ever-shifting Agulhas Current Retroflection (Lutjeharms, 2007).
Parent and associated child cyclonic eddies found to emulate 92A’s trajectory may be located in figures 6.8 – 6.16 and include the following: 93A; 94A; 94H; 95B; 95C; 95D; 95W; 96A; 96S; 97X; 98B; 99B; 99E; 99S; 99W; 00A; 01A; 01F and 01L. Overall trajectories for these cyclones were south-south-westerly, however, they often followed a convoluted route to achieve this direction.

Cyclone 93A in figure 6.8 (again in red) is a perfect example of an eddy produced within the turbulent region that borders the Agulhas Retrotection and ring shedding zone. After forming over the continental slope alongside Lamberts Bay, it remained within that location for a couple of weeks, growing in intensity. Its next recorded movement shows it travelling quickly south before veering slightly southwest and into deeper water, where 93A1 was occluded. 93A continued to head southwest until 36 °S where, within a week, it covered almost 1.5 degrees of latitude (90 n.mi) before turning 90 °E. Within another week it had turned south again before appearing to rebound off the Wyandot Seamount (1895 m) in a south-south-westerly direction. In an action reminiscent of a pinball game, it was seen to collide with the Erika Seamount (1187 m), sending 93A on a south-easterly course before bouncing off the Agulhas Ridge in a north-westerly direction. Its signal disappeared from the altimetry data once it again came into contact with the Schmidt-Ott Seamount (1122 m).

Figure 6.8: Trajectories of cyclonic eddies formed in 1992
After its formation, 93AI proceeded in a north-easterly direction. Upon reaching the upper limits of the continental slope, it followed the contour of the slope in a south-easterly manner. After travelling across approximately two degrees of latitude and two degrees of longitude (120 x 120 n.mi) it merged with a massive sea surface depression formed within the Agulhas Retroflection region.

93AI split from 93A at 35 °S 14 °E, potentially due to pressure from an adjoining Agulhas Ring. Its initial course was due north, covering approximately two degrees of latitude (120 n.mi) in a week. At 33 °S it turned south again for a period of a few weeks, before rapidly moving to 35.5 °S. From there it took weeks to reach 36 °S before sauntering off in a north-westward direction. 93AI was the last and most short-lived child cyclonic eddy of the 93A family. It split from 93AI, apparently forced by an adjacent Agulhas Ring, around which it arced in a semi-circle before dissipating.

![Diagram](image)

**Figure 6.9: Trajectories of cyclonic eddies formed in 1994**

Figure 6.9 represents an exception to the west-southwest generalisation. Cyclonic eddy 94N (in blue) was also formed on the continental slope, yet made its way south. Other cyclonic eddies to emulate this slower, southward motion are 95N (figure 6.10), the partial trajectories of 96Q (figure 6.11), 97J, 97K, 97X (figure 6.12), 98X (figure 6.13) and 99A (figure 6.14). This southward progression may possibly be linked to the southward flowing Benguela Undercurrent (Shannon and Nelson, 1996).
95P in figure 6.10 (in yellow) is a very clear example of those cyclonic eddies that moved in a more linear direction upon formation. Its trajectory was almost directly westward from 12 °E to 5 °E before encountering the Walvis Ridge. There a smaller cyclonic eddy merged with 95P, allowing it to grow from a square degree in size (60 x 60 n.mi) to an area larger than three-square degrees (180 x 180 n.mi). The merger potentially provided 95P with enough energy to traverse the Walvis Ridge, as it reacted with only a slight northward vector to its original course.

Figure 6.10: Trajectories of cyclonic eddies formed in 1995.

Once across the Walvis Ridge, 95P continued in a north-westerly direction before the \( \beta \)-effect (figure 2.3; page 21) played a part in redirecting the cyclonic eddy meridionally towards the southwest. As it neared the Mid-Atlantic Ridge, 95Pl was shed at 28 °S 3.5 °W, before 95P continued west. Upon reaching the highest portion of the Mid-Atlantic Ridge, 95P's signal disappeared from view in the data. 95Pl headed directly north after creation, before turning west to follow the fate of its parent.
Within figure 6.11, 96W (in black) symbolises the most commonly found cyclonic eddy trajectory within the 1992 to 2001 satellite altimetry data set. With a formation location on the continental slope and a west-southwest trajectory for the majority of its track, 96W traversed the Cape Basin. During its lifetime it occluded two child cyclonic eddies, 96Wi and 96Wii. The west-southwest course represented by cyclonic eddy 96W agrees with research completed by Morrow et al. (2004), in which they state that 66% of their observed cyclonic eddies formed within the Cape Basin were seen to travelled in this direction (figure 2.4; page 22).

Figure 6.11: Trajectories of cyclonic eddies formed in 1996

Figure 6.12 provides a good overall view of the south-westerly progression of cyclonic eddies in the Cape Basin. In addition, cyclonic eddy 97X (in yellow) highlights the effects of topography on cyclonic eddy movement. After 97X split numerous times due to potential anticyclonic eddy and Agulhas Ring interactions, two of the child cyclonic eddies reached the lower limits of the Walvis Ridge. Here they twisted and turned in various directions as they made their way further into the shallower water, eventually dissipating.
Figure 6.12: Trajectories of cyclonic eddies formed in 1997.

Figure 6.13: Trajectories of cyclonic eddies formed in 1998. Stars indicate cyclonic eddies still present but leaving the study area.
Figure 6.13 and figure 6.14 both show representative examples of cyclonic eddies that retained sufficient energy to translocate across both the Walvis and Mid-Atlantic Ridge. 98C in yellow (figure 6.13) and 99S in red (figure 6.14) were both formed on the continental slope. 98C followed a westerly track to cross the Walvis Ridge relatively quickly, occluding only one child cyclonic eddy while still in the Cape Basin. 99S, on the other hand, remained within the Cape Basin for long enough to shed a number of eddies, two of which were able to cross both the Walvis and Mid-Atlantic Ridges.

Figure 6.14: Trajectories of cyclonic eddies formed in 1999. The size represents the last position of cyclonic eddy 98SIII, which was still present at the end of the 1992 – 2002 data set.

Figure 6.15 depicts another feature of the topography and its effect on cyclonic eddy translocation. Although also apparent in other years, figure 6.15's 001A (pale pink), 001 (yellow), 00S (red) and 00Z (purple) clearly show how the deeper part of the Walvis Ridge allowed cyclonic eddies with sufficient energy to leave the Cape Basin. Further south, cyclonic eddies of similar size and strength during formation (00A in green and 00B in black), were not able to transverse their topographical obstacle. Instead they remained within the Cape Basin, where they were subjected to more frequent splitting and merging.
Figure 6.15: Trajectories of cyclonic eddies formed in 2000. Stars indicate cyclonic eddies still present at the end of the 1992–2002 data set.

01D (in black) and 01E (in blue) in figure 6.16 were unusual in that they were very short-lived. Although both grew to normal size, neither translocated a great distance before being reduced in size again. Interestingly, in both cases an anticyclonic eddy was present in the region. 01D was adjacent to one but 01E was not. However, most of the other cyclones shown here are representative of the five major directional groups to be found within these 1992 – 2001 satellite altimetry data. The trajectories in figure 6.16 were shorter than in previous figures as this was the final year of the data set. White stars indicate cyclonic eddies that continued past February 2002.

It is therefore clear that change in propagation direction is common in cyclonic eddies. Topography, background water movement and interactions with other cyclonic and anticyclonic eddies may play a large role in the translocation of the Cape Basin cyclonic eddies.

To summarise, cyclonic eddies within the Cape Basin follow a multi-directional path, where the majority of only 31.58% translocated in the expected west-southwest overall direction. In fact, 27.37% travelled in a west-northwest direction. Additionally, these bearings only constitute an overall path, the cyclonic eddies were seen to meander along many differing routes during their life spans.

Since the trajectories of these cyclonic eddies have now been considered, the next logical action would be to investigate their life spans over these distances. Determining how long they are present within the region may provide insights to their impact on their surroundings.
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...
Child cyclonic eddies (figure 6.17) had an overall shorter lifespan than parent cyclonic eddies, with the 141 surveyed child cyclonic eddies surviving for 118 days on average. Although they were more numerous than their parents, they were generally smaller and weaker and may drop below the $< 10$ cm SSHA criterion sooner than parent cyclonic eddies. The shortest living child cyclonic eddies were 93Viii, 94Tiii, 97Fi, 97Xi, 97Xv, 98Ai and 00Avi, all evident in the satellite altimetry for only seven days. The longest living child cyclonic eddy was 99Sii at 707 days.

Annually, there is very little difference between the average life spans of approximately 237 days for parent cyclonic eddies formed during 1993 and 1996. Lifespan dropped to 219 days for those formed in 1992, rising to double that at 428 days for parent cyclonic eddies created in 2000. In total, the range between average lifespan over the study period was 226 days. Appendix C (page 146) contains the original lifespan data for parent and child cyclonic eddy, represented annually in figures 9.16 and 9.24.

Table 6.1: Parent and child cyclonic eddy average lifespan.

<table>
<thead>
<tr>
<th>Year</th>
<th>Parent cyclonic eddy mean (days)</th>
<th>Child cyclonic eddy mean (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1992</td>
<td>219.00</td>
<td>245.00</td>
</tr>
<tr>
<td>1993</td>
<td>235.80</td>
<td>98.00</td>
</tr>
<tr>
<td>1994</td>
<td>245.00</td>
<td>59.50</td>
</tr>
<tr>
<td>1995</td>
<td>241.50</td>
<td>56.00</td>
</tr>
<tr>
<td>1996</td>
<td>229.09</td>
<td>63.00</td>
</tr>
<tr>
<td>1997</td>
<td>202.46</td>
<td>52.50</td>
</tr>
<tr>
<td>1998</td>
<td>315.00</td>
<td>49.00</td>
</tr>
<tr>
<td>1999</td>
<td>268.55</td>
<td>28.00</td>
</tr>
<tr>
<td>2000</td>
<td>428.00</td>
<td>115.50</td>
</tr>
</tbody>
</table>

A one-way analysis of variance (ANOVA) is deemed the appropriate test to determine whether any inter-annual variability is statistically detectable within the lifespan data of parent cyclonic eddies (table 6.1). As ANOVA assumes that the data are sampled from populations with identical standard deviations (Zar, 1984), Bartlett’s test is used to meet this criterion. Additionally, each annual category was subjected to a Kolmogorov-Smirnov normality test, with all but 1994 passing, allowing for the use of ANOVA. Bartlett statistic (corrected) equalling 6.244 ($P$-value = 0.6199) suggested that the differences among the standard deviations (SD) are not significant.
For all future lifespan data analysis, assume that the null hypothesis states that there is no variation between annual lifespan, while the alternative hypothesis states that there is. With a \( P\text{-value} = 0.3719 \), the data are considered not statistically significant \((df = 93, f = 1.099, \text{Power} = 0.3192)\), therefore it is not possible to reject the null hypothesis. Variation among parent cyclonic eddy yearly means is not significantly greater than expected by random chance, therefore there is no extreme difference between the average lifespan of parent cyclonic eddies on an annual basis.

Annual child cyclonic eddy lifespan mirrored that of the parents for 1994 to 1998 in that the annual mean remained constant at more or less 56 days. 1992 and 2000 produced child cyclones that survived for far longer than the majority 1994 – 1998 group. 1992 produced child cyclones that lasted for approximately 217 days longer than those formed in 1999, the year containing the shortest mean lifespan for child cyclones.

Unlike annual parent lifespan, child cyclonic eddy (table 6.1) lifespan data does not pass the Kolmogorov-Smirnov normality test. Attempts to transform the data to a lognormal distribution (normally used for measurement); a Poisson distribution using a square root of a count; or a binomial distribution using a arcsine of square root of proportion function did not succeed in converting the data into a Gaussian (normal) distribution.

Therefore it is necessary to use a non-parametric ANOVA statistical test. The Kruskal-Wallis (KW) test is applied to the data, resulting in a \( P\text{-value} \) of 0.2513 \((H = 10.199, \text{Power} = 0.5022)\), which is considered not statistically significant. Therefore it is not possible to reject the null hypothesis. Variation among child cyclonic eddy yearly medians is not significantly greater than expected by chance. It is important to note that the Kruskal-Wallis test approximates the \( P\text{-value} \) (from the chi-square distribution) because some years had duplicate lifespan values.

When contemplating lifespan on a seasonal scale, there did not appear to be a difference between the lifespan of parent cyclonic eddies formed between seasons (table 6.2). While those formed in summer lasted on average 288 days, those formed in winter only lasted approximately 233 days.
Child cyclonic eddies, on the other hand, did seem to have notably different life spans. Those that split from their parents in autumn tended to survive for an average of 153 days, while those formed in spring only 89 days.

<table>
<thead>
<tr>
<th>Seasonal lifespan</th>
<th>Parent (days)</th>
<th>Child (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer (Dec – Feb)</td>
<td>288</td>
<td>107</td>
</tr>
<tr>
<td>Autumn (Mar – May)</td>
<td>258</td>
<td>153</td>
</tr>
<tr>
<td>Winter (Jun – Aug)</td>
<td>233</td>
<td>121</td>
</tr>
<tr>
<td>Spring (Sep- Nov)</td>
<td>240</td>
<td>89</td>
</tr>
</tbody>
</table>

A one sample $t$-test is used to determine whether the seasonal formations of parent cyclonic eddies are significantly different. Results shows a $P$-value = 0.9850 ($df = 3, t = 0.02037, \text{Power} = 1.0000$), which is considered not significant, given a null hypothesis of no difference and random chance. Therefore there is no significant difference between mean life spans of parent cyclonic eddies per season.

Using the same test to investigate possible associations within the child cyclonic eddy data, the two tailed $P$-value > 0.9999 ($df = 3, t = 0.000, \text{Power} = 0.9999$), resulting in a failure to reject the null hypothesis. Therefore the mean values are not significantly different from the expected mean, given the null hypothesis and random chance. With no significant differences between annual or seasonal life spans, perhaps location may play a part in the longevity of these cyclonic eddies.

Potential lifespan limiting factors are clearly visible in figure 6.1 (page 82) at the beginning of this chapter, where a clearly defined Walvis and Mid-Atlantic Ridge rise dramatically from the ocean floor. Other probable hazards to Cape Basin cyclonic eddies are positive sea level anomalies (SLA), such as anticyclonic eddies or Agulhas Rings, which may degrade cyclonic eddies in passing. Both topography and other SLA's affect the northern and southern portions of the study area to differing extents. An interesting consideration in this regard is whether cyclonic eddies formed in the north lived longer than those formed in the south.
Figure 6.18. Lifespan of parent cyclonic eddies formed to the north and south of 30°S between 1992 and 2001 (in days).

Figure 6.19. Lifespan of child cyclonic eddies formed to the north and south of 30°S between 1992 and 2001 (in days).
As used previously in this study, the 30 °S line of latitude was identified as the split between the northern and southern portions of the study area. Figures 6.18 and 6.19 show the relationships between parent and child cyclonic eddies of both regions respectively.

Parent cyclonic eddies (table 6.3) formed within the northern reaches of the study area lived for a mean of 238 days, approximately 18 days shorter than the average for their southern cousins. Child cyclonic eddies (table 6.4) formed in the north had a survival rate of 99 days, lower than the mean of 119 days for those in the south. In spite of that, the longest surviving parent cyclonic eddy (at 840 days) was formed within the northern region of the study area. This region was also responsible for producing the shortest parent cyclonic eddy identified during the study period. The north created the longest child cyclonic eddy recorded during this time frame, although both regions produced child cyclonic eddies with a one-week lifespan.

A $t$ test is selected to determine whether the means of parent cyclonic eddies formed in the north and the south differ significantly. However, a $t$ test assumes that the data are sampled from a Gaussian distribution. After subjection to a Kolmogorov-Smirnov normality test, the lifespan data for the parent cyclonic eddies formed in the northern region of the study area did not fit a normal distribution. To acquire normality, the data were transformed using a lognormal function. This transformation allows for the use of an unpaired $t$ test with Welch correction (as the two groups have different standard deviations).

**Table 6.3: Northern verses southern parent cyclonic eddy annual lifespan data summary (before a lognormal distribution correction was applied)**

<table>
<thead>
<tr>
<th></th>
<th>North (days)</th>
<th>South (days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>238</td>
<td>256</td>
</tr>
<tr>
<td># of points</td>
<td>56</td>
<td>49</td>
</tr>
<tr>
<td>Std deviation</td>
<td>188.147</td>
<td>181.729</td>
</tr>
<tr>
<td>Std error</td>
<td>25.142</td>
<td>25.961</td>
</tr>
<tr>
<td>Minimum</td>
<td>21</td>
<td>28</td>
</tr>
<tr>
<td>Maximum</td>
<td>840</td>
<td>742</td>
</tr>
<tr>
<td>Median</td>
<td>171.5</td>
<td>231</td>
</tr>
<tr>
<td>Confidence limit</td>
<td>50.386</td>
<td>52.199</td>
</tr>
</tbody>
</table>

(95.0 %)
The unpaired \( t \) test applied to the parent lifespan data (represented in table 6.3) returned a two-tailed \( P\)-value of 0.4866 (Welch’s approximate \( t = 0.6982 \), not assuming equal variance, \( df = 102 \), \( Power = 0.7164 \)), which is considered not statistically significant. Therefore it is not possible to reject the null hypothesis. There does not appear to be a significant difference between the means of parent cyclonic eddies formed in the northern and the southern regions of the study area.

Table 6.4: Northern verses southern child cyclonic eddy annual lifespan data summary

<table>
<thead>
<tr>
<th></th>
<th>North</th>
<th>South</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>99.458</td>
<td>118.796</td>
</tr>
<tr>
<td># of points</td>
<td>56</td>
<td>49</td>
</tr>
<tr>
<td>Std deviation</td>
<td>145.828</td>
<td>139.254</td>
</tr>
<tr>
<td>Std error</td>
<td>21.049</td>
<td>13.338</td>
</tr>
<tr>
<td>Minimum</td>
<td>7</td>
<td>7</td>
</tr>
<tr>
<td>Maximum</td>
<td>707</td>
<td>672</td>
</tr>
<tr>
<td>Median</td>
<td>42</td>
<td>63</td>
</tr>
<tr>
<td>Confidence limit (95.0 %)</td>
<td>42.344</td>
<td>26.438</td>
</tr>
</tbody>
</table>

Northern and southern child cyclonic eddy lifespan data does not appear to have a normal distribution. Therefore a Mann-Whitney test is selected to determine whether the medians of the north and the south data (represented in table 6.4) differ significantly. The Mann-Whitney test applied to the child lifespan data returned a two-tailed \( P\)-value of 0.2263 (\( U = 2298.0 \), \( U' = 2934.0 \), \( n_1 = 3474 \), \( n_2 = 8929.0 \)), which is considered not statistically significant. The \( P\)-value is an estimate based on a normal approximation, as the ‘exact’ method would not be exact due to tied ranks. Therefore it is not possible to reject the null hypothesis. There does not appear to be a significant difference between the medians of child cyclonic eddies formed in the north and south.

Figure 6.20 shows that there was no noticeable connection in lifespan between parent and child cyclonic eddies formed north of 30 °S. Figure 6.21 represents life spans for those formed to the south of 30 °S. What is very apparent in figure 6.20 is that the lifespan of child cyclonic eddies was generally less than that of the parent cyclonic eddies in the north. For a second time it is possible to see that there is no significant association between the life spans of the parent and child cyclonic eddies in both the north and south.
Figure 6.20: Lifespan of parent and child cyclonic eddies formed north of 30°S between 1992 and 2001 (in days).

Figure 6.21: Lifespan of parent and child cyclonic eddies formed south of 30°S between 1992 and 2001 (in days).
Due to a non-Gaussian data distribution, a non-parametric Mann-Whitney test is chosen to consider any potential significant difference between the mean lifespan of parent and child cyclonic eddies in the northern part of the study area. The null hypothesis states that there is no difference between the mean lifespan of the two groups. The resulting two-tailed P-value of $< 0.0001$ ($U = 556.0; U' = 2132.0$) is considered extremely significant and is detailed in table 6.5. Therefore it is possible to reject the null hypothesis and state that the means of parent and child cyclonic eddy lifespan in the north do differ significantly.

The same statistical test is applied to the lifespan data from the southern cyclonic eddies (table 6.5). Again the null hypothesis states that there is no difference in lifespan between parent and child cyclonic eddies within the southern region. A similar result of an extremely significant P-value of $< 0.0001$ is returned ($U = 1233.0; U' = 4108.0$), allowing for the rejection of the null hypothesis. Therefore, there is a significant difference between the lifespan of parent and child cyclonic eddies within the southern region of the study area. This means that there is a difference in mean lifespans of parents and child cyclonic eddies in the north and the south. Contemplating parent and child cyclonic eddy relationships, it may be feasible that the time it takes for a parent cyclonic eddy to produce a child plays a part in their overall lifespan.

### Table 6.5: Data summary of the lifespan of parent and child cyclonic eddies in the northern and southern regions of the study area

<table>
<thead>
<tr>
<th></th>
<th>North Parent</th>
<th>North Child</th>
<th>South Parent</th>
<th>South Child</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>238.00</td>
<td>99.46</td>
<td>256.12</td>
<td>118.71</td>
</tr>
<tr>
<td># of points</td>
<td>56</td>
<td>48</td>
<td>49</td>
<td>109</td>
</tr>
<tr>
<td>Std deviation</td>
<td>188.15</td>
<td>145.83</td>
<td>181.73</td>
<td>139.25</td>
</tr>
<tr>
<td>Std error</td>
<td>25.142</td>
<td>21.049</td>
<td>25.961</td>
<td>13.338</td>
</tr>
<tr>
<td>Minimum</td>
<td>21</td>
<td>7</td>
<td>28</td>
<td>7</td>
</tr>
<tr>
<td>Maximum</td>
<td>840</td>
<td>707</td>
<td>742</td>
<td>672</td>
</tr>
<tr>
<td>Median</td>
<td>171.5</td>
<td>42</td>
<td>231</td>
<td>63</td>
</tr>
<tr>
<td>Lower CI (95.0 %)</td>
<td>187.58</td>
<td>57.074</td>
<td>203.87</td>
<td>92.244</td>
</tr>
<tr>
<td>Upper CI (95.0 %)</td>
<td>288.42</td>
<td>141.84</td>
<td>308.37</td>
<td>145.17</td>
</tr>
</tbody>
</table>

As shown in the section on potential time lag difference (chapter 5.2; page 74), one parent cyclonic eddy may spawn numerous child eddies and these child eddies may in turn spawn other sibling eddies. The life spans of these data groups are represented in figure 6.22, where each group is named with a nomenclature that indicates the order of their formation, from consecutive first to eighth string.
Due to a non-Gaussian data distribution, the non-parametric Kruskal-Wallis ANOVA test used returns a result of extreme significance, with a \( P \)-value of less than 0.0001 (\( H = 56.233 \)). Variation among column medians is significantly greater than expected by chance. Therefore, child cyclonic eddies formed from consecutive groups do not have the same lifespan means. This is an interesting result as a Kruskal-Wallis test that looked at the overall lifespan of child cyclonic eddies did not produce a significant result. Therefore, a positive result indicates a subsequent trend in the group formation order of child cyclonic eddies, as shown in table 6.6.

Table 6.6: Lifespan statistics of consecutive groups of cyclonic eddies

<table>
<thead>
<tr>
<th>Group</th>
<th>Sample number</th>
<th>Median</th>
<th>Minimum</th>
<th>Maximum</th>
</tr>
</thead>
<tbody>
<tr>
<td>Parent</td>
<td>101</td>
<td>217</td>
<td>21</td>
<td>840</td>
</tr>
<tr>
<td>Child 1</td>
<td>61</td>
<td>56</td>
<td>7</td>
<td>530</td>
</tr>
<tr>
<td>Child 2</td>
<td>36</td>
<td>29</td>
<td>14</td>
<td>307</td>
</tr>
<tr>
<td>Child 3</td>
<td>21</td>
<td>49</td>
<td>7</td>
<td>666</td>
</tr>
<tr>
<td>Child 4</td>
<td>14</td>
<td>70</td>
<td>7</td>
<td>588</td>
</tr>
<tr>
<td>Child 5</td>
<td>8</td>
<td>56</td>
<td>7</td>
<td>448</td>
</tr>
<tr>
<td>Child 6</td>
<td>6</td>
<td>112</td>
<td>7</td>
<td>511</td>
</tr>
<tr>
<td>Child 7</td>
<td>4</td>
<td>42</td>
<td>14</td>
<td>112</td>
</tr>
<tr>
<td>Child 8</td>
<td>1</td>
<td>63</td>
<td>63</td>
<td>63</td>
</tr>
</tbody>
</table>

Figure 6.22: Lifespan, in days, of consecutive groups of child cyclonic eddies.
Within this study a lot of attention has been paid to the formation of the cyclonic eddies. The next few paragraphs address the demise of these features in closer detail. Reverting to figures included earlier in this chapter (section 6.1: Cyclonic eddy trajectories), figures 6.7 to 6.16 (pages 86 – 93) show that quite a number of individual parent and child cyclonic eddies were no longer visible on the satellite altimetry after they had come into contact with a bathymetric feature. Certain cyclonic eddies, such as 92Aii (figure 6.7; page 86); 94Aii (figure 6.9; page 88) and 98Biv (figure 6.13; page 91) were lost to the satellite imagery after they came into contact with a seamount. 93A (figure 6.8; page 87) provided a comprehensive example of how a cyclonic eddy may collide with a seamount and change its direction, before rebounding from another seamount. Its new reverse track headed it straight for the original seamount again, where it disappeared from the satellite trace.

Other cyclonic eddies ceased to exist upon contact with the Walvis Ridge. 12 cyclonic eddies (4.6 % of those studied here) disappeared upon entering the more shallow waters leading up to the Walvis Ridge, namely 93D; 94Aiv; 94M; 95F; 95L; 97B; 98Ci; 98Ciii; 99Fii; 00Avii; 00Aviii and 00Bii (as seen in figures 6.8 – 6.15; pages 87 – 93). Seven (2.7 % of the cyclonic eddies considered here) expired on top of the Walvis Ridge and were identifiable as 92Bi; 93V; 94J; 97K; 97Xv; 97Xvi and 97Xvii (traceable in figures 6.7 – 6.12). Eleven other cyclonic eddies (4.2 % of these data) managed to traverse the shallowest areas of the Walvis Ridge, only to terminate upon reaching the deeper waters on the other side. 92R; 92Ri; 95J; 95Wiv; 95Cv; 96Ai; 97Pi; 98Civ; 98E; 99Fiv and 99Siv (figures 6.7 – 6.14) were examples of cyclonic eddies that followed this trend.

Cyclonic eddies that traversed all of these obstacles on their westward trajectory were then faced with an even more unforgiving hydrographical feature, the Mid-Atlantic Ridge. Not many managed to cross the oceanic mountain range that forms the divergent tectonic boundary between the African and South American Plates. Eleven cyclonic eddies (4.2 % of those studied here), as reflected within these data, that were unable to negotiate this feature were 94A; 94Yii; 94Tiv; 95P; 95Pi; 96F; 96Fvi; 96Fvii; 98Ci; 00D and 00Di (figures 6.9 – 6.15).
A few cyclonic eddies did cross both the Walvis and the Mid-Atlantic Ridge. These were 92Dii; 98C; 98Cvi and 99Siii, only four eddies in eight consecutive years of data (1.5% of these data). 92Dii (figure 6.7; page 86) lived for 672 days before fading from the altimetry. 98C and 98Cvi (figure 6.13; page 91) survived for 840 and 196 days respectively before moving westward beyond the study area. 99Siii (figure 6.14; page 92) existed for at least 686 days after it split from its parent and was still visible as a shallow sea surface anomaly at the end of the study period. Although only 17% of the studied parent and child cyclonic eddies interacted with a topographical feature, upon considering the above eddies, it is safe to assume that the topography of the ocean basin has an effect on the lifespan of certain cyclonic eddies.

Therefore it is possible to say that the parent cyclonic eddies studied here had a mean lifespan of 254 days, although one cyclonic eddy survived for 742 days overall. Child cyclonic eddies lived for approximately 188 days, although their lifespan was found to range between 7 and 707 days. When considering lifespan associations between individual parent and child cyclonic eddies, as well as between the parent and child groups, over annual and seasonal timescales, northern and southern regions, no significant difference between mean life spans was apparent. There was a significant relationship between the mean lifespan of parent and child cyclonic eddies formed within the north and south of the study area.

This association led to the investigation of whether the time lag between parent formation and the occlusion/splitting of a child cyclonic eddy was related. Overall time difference was not significant, although there was a significant difference between each subsequent child occlusion group, suggesting a direct relationship between the youth of a parent cyclonic eddy and a child's potential for longevity. However, other variables were seen to influence that longevity, such as the topography and other mesoscale features within the region. Hence a look at the physical properties of these eddies in the next section 6.3 may shed some light on their life styles.
6.3 Physical properties of cyclonic eddies

Altimetric data were used to determine sea level anomalies from the mean sea surface height, which would indicate cyclonic and anticyclonic eddies within the study area. As mentioned in chapter 5, section 5.3 (page 80), containing formation size parameters of cyclonic eddies, eddies were considered formed and traceable once they exhibited a signal of 10 cm deeper than the mean sea surface height. This criterion is noticeable in figure 6.23 below, which contains a representative sample of accumulated sea surface height data.

Figure 6.23 shows that there is no set depression height for these cyclonic eddies. Sea surface height depths may range from -8 to -43 cm’s during the lifespan of an individual cyclonic eddy. For a greater understanding of the apparent randomness of cyclonic eddy sea surface height, it is necessary to review the formation location and overall trajectory of the cyclonic eddies represented below.

![Cyclonic eddy sea surface height over time](image)

*Figure 6.23: The sea surface height of cyclonic eddies 92A, 92B, 93A and 93B over time (7 day increments), as determined using AVISO altimetric data.*
To recap, cyclonic eddies 92A (figure 6.7; page 86) and 93A (figure 6.8; page 87) were formed to the immediate north of the Agulhas Retroreflection, within the region of the Columbine Upwelling Cell. After moving in a southwest direction for 133 days, cyclonic eddy 92A divided to form two child cyclones of equal size. Child cyclonic eddy, 92Ai, translocated in a mean westerly direction for another 252 days, while 92Aii shot down the original south-westward path for another two weeks, before colliding with the Vema Seamount.

Cyclonic eddy 93A produced triplets and mirrored the pattern of the 92A cyclonic eddy family. Both sets of cyclonic eddies remained within the known region of aggressive mixing and stirring (Boebel et al., 2003).

Cyclonic eddy 92R (figure 6.7), on the other hand, was formed within the Lüderitz Upwelling cell. 92R progressed in a north-westward direction for 84 days before occluding a child cyclonic eddy, 92Ri. After heading in opposite poleward directions, both 92R and 92Ri ultimately followed a true westward direction. Both encountered the Walvis Ridge and disappeared from altimetric trace upon traversing the Walvis Ridge and encountering deeper water.

Cyclonic eddy 93B (figure 6.8) was formed between those cyclones mentioned above, within the meridian of the Namaqua Upwelling cell. 93B did not appear to split at any point during its lifespan of 497 days, nor did it appear to have deviated greatly from its westerly trajectory.

Unfortunately, an in-depth look at the sea surface height of numerous cyclonic eddies is unrealistic due to the nature of this study. Not only were individual cyclonic eddies manually tracked in a time consuming fashion, but the figures used to display the AVISO sea surface height data of the cyclonic eddies provided a non-numerical, colour-bar legend for the height depression level (as visible in chapter 4, figure 4.4; page 32). Therefore any results obtained for sea surface height in this manner would be non-analytical and subjective, considered inconclusive for a statistical study. An objective review that allows for quantification and statistical analysis of these data is necessary.
However, after an intense, visual inspection of the data set, a few patterns became apparent. Indulging in some postulation, it seems that the area of formation plays a large role in the achieved intensity of these cyclonic eddies. Those formed in the southern, highly volatile region of the Agulhas Retroflection tended to acquire a far deeper and far more varied sea surface height than those formed further north. In the north they were seen to remain relatively constant in overall sea level height during their life spans. Additionally, it was observed that cyclonic eddies were intensified by a passing SSH High, as evident by the number of child cyclonic eddies formed in association with anticyclonic eddies (covered statistically in chapter 7; page 119).

Again speculating, the cyclonic eddies formed above the influences of the Agulhas Retroflection (northerly formation locations) tended to remain at the same intensity for a number of weeks before increasing or decreasing in sea surface height, whereas those formed in the south near the Agulhas Retroflection oscillated between deeper and shallower heights within a number of days. It is possible to reason that these behavioural trends may be accredited to interactions with anticyclonic eddies found within the Ring Corridor (Garzoli et al., 1999), as well as the encapsulating influence of the dynamic region that is the Cape Basin (Boebel et al., 2003).

Cyclonic eddies 92A; 93A and 93B were seen to become deeper in sea surface height prior to dissipating, as seen in figure 6.23 (page 107). This behaviour may be linked to topography, as 92A encountered the Vema Seamount (figure 6.7; page 86). 93A collided with the Agulhas Ridge and its surrounding topographical features (figure 6.8; page 87), while 93B crossed the Walvis Ridge at approximately 31 °S 0 °E, before veering southwards and fading from the altimetry trace (figure 6.8).

92R and 92Ri both crossed the Walvis Ridge at approximately 25.8 °S 6 °E and 27.8 °S 3.5 °E respectively, without drastically affecting their sea surface height. However the topographical feature was seen to weaken the cyclonic eddies to a slight degree. These case studies provide a reasonable representation of the most commonly seen effects that topography has on the cyclonic eddies found within this study. As mentioned, a future analytical approach should be undertaken to fully quantify these postulations.
Combining hydrographical and altimetric data, table 6.7 shows the varying properties of the cyclonic eddies present in both data sets. As these cyclonic eddies are not inclusions of water from other oceans (such as Agulhas Rings bringing Indian Ocean water into the Atlantic Ocean) but are formed from the background water of the region, it is not possible to determine their signature from sea surface temperature traces. Therefore the only indicators of their presence are the direction of water flow, sea surface height and elevation of the thermohaline within the water column.

Table 6.7: Physical properties for those cyclonic eddy present in both hydrographical and altimetric data sets. Values represent depth in meters

<table>
<thead>
<tr>
<th>Variable depth in meters</th>
<th>92E</th>
<th>99H</th>
<th>99S</th>
<th>00A</th>
<th>00Ai</th>
<th>00B</th>
<th>01E</th>
<th>Retroflection cyclone A</th>
<th>Retroflection cyclone B</th>
</tr>
</thead>
<tbody>
<tr>
<td>15 °C isotherm depth</td>
<td>20</td>
<td>25</td>
<td>135</td>
<td>110</td>
<td>140</td>
<td>120</td>
<td>10</td>
<td>200</td>
<td>145</td>
</tr>
<tr>
<td>Background 15 °C isotherm depth</td>
<td>170</td>
<td>50</td>
<td>180</td>
<td>270</td>
<td>270</td>
<td>300</td>
<td>40</td>
<td>375</td>
<td>325</td>
</tr>
<tr>
<td>10 °C isotherm</td>
<td>370</td>
<td>197</td>
<td>360</td>
<td>200</td>
<td>240</td>
<td>405</td>
<td>190</td>
<td>375</td>
<td>300</td>
</tr>
<tr>
<td>Background 10 °C isotherm depth</td>
<td>490</td>
<td>235</td>
<td>425</td>
<td>580</td>
<td>580</td>
<td>625</td>
<td>205</td>
<td>745</td>
<td>730</td>
</tr>
<tr>
<td>35 psu salinity depth</td>
<td>Above the surface</td>
<td>97.2</td>
<td>N/a</td>
<td>210</td>
<td>200</td>
<td>290</td>
<td>100</td>
<td>275</td>
<td>250</td>
</tr>
<tr>
<td>Background 35 psu salinity depth</td>
<td>360</td>
<td>150</td>
<td>N/a</td>
<td>480</td>
<td>480</td>
<td>500</td>
<td>140</td>
<td>625</td>
<td>490</td>
</tr>
</tbody>
</table>

Additionally, thermohaline signals might be weaker than needed for identification. This is due to the probability that the fortuitous hydrographical platforms, dedicated to Agulhas Ring and other routine hydrographic studies, did not cross the center of the cyclonic eddy, but rather merely grazed the edge of the feature. Within table 6.7, all the named cyclonic eddies were formed within the study area, whereas the two Retroflection cyclonic eddies, A and B, were not. The Retroflection cyclonic eddies, A and B, were included here solely due to their appearance in both data sets, providing additional proof that confidence may be placed in the altimetric signals.
Cyclonic eddy, 92E, is represented graphically in chapter 4. Figure 4.4 (page 32) shows the altimetric trace and figure 4.5 (page 33) shows its hydrographic thermohaline properties. Appendix D (page 151; figures 9.25 – 9.37) contains the geographical representation of the rest of the cyclonic eddies listed in table 6.7.

From table 6.7 it is possible to observe the anticipated rise in depth of the thermoclines and haloclines of the cyclonic eddies. For example, within cyclonic eddy 92E the 15 °C isotherm depth rose from 170 m to 20 m, while the 10 °C isotherm rose from 490 m to 370 m. Throughout the cyclonic eddies shown above the 15 °C isotherm was seen to rise by varying degrees; from 25 m in 99H to 180 m in 00B and Retroflection cyclone B. The 10 °C isotherm displayed the same variability, fluctuating in difference between 15 m in 01E and 430 m in Retroflection cyclone B.

The same trend is noticeable when considering the 35 psu line of constant salinity. Within 92E, the salinity was less than 35 psu at the surface. However the surrounding water recorded this salinity level at 360 m. Of the examples listed in table 6.7, the differences in the halocline ranged between 40 m in 01E and 350 m around Retroflection cyclonic eddy A, showing the large fluctuations visible within these parameters. Hence the use of only one station of hydrographic data to distinguish a cyclonic eddy is not a viable method of identification.

Unfortunately, due to the nature of the collection of these hydrographical data, as well as the formation location that ensues that the cyclonic eddies contain the same thermohaline signal as the background water, it is difficult to discern an identifiable physical pattern or template for these cyclonic eddies. Future hydrographic studies that target these Cape Basin cyclonic eddies are recommended to increase thermohaline knowledge of these features. Still, with a plethora of altimetric data available there are many facets of these cyclonic eddies that may be investigated, such as section 6.4, which considers their behaviour during their lifespan.
6.4 Cyclonic eddy behaviour

As mentioned in the previous chapters, cyclonic eddies are dynamic features that are subject to changes in shape and size. Cyclonic eddies are known to split and form child cyclonic eddies, as covered in chapter 5.1 (page 45). Figure 6.24 shows an overview of cyclonic eddy split locations. It is noticeable that there is no distinct pattern of annual splitting.

![Figure 6.24: Location of cyclonic eddy splits between 1992 and 2001.](image)

Cyclonic eddies were often seen to merge with one another, as shown in figure 6.25. Descriptive statistics applied to the data show an average merge rate of 16.375 per year (standard deviation = 8.123; standard error = 2.872) for parent cyclonic eddies and 14 per year (standard deviation = 9.181; standard error = 3.246) for child cyclonic eddies. To test whether there is any variability in their annual merge frequency, a chi-square goodness of fit test is used on both the parent and child cyclonic eddy merge data (table 6.8). As throughout this work, the null hypothesis states that there is no difference in merge frequency between parent and/or child cyclonic eddies, whereas the alternative hypothesis states that there is.
Applying a chi-square statistical test to the annual parent cyclonic eddy merge frequency data (table 6.8), the resulting P-value = 0.000202 ($\chi^2 = 28.206, df = 7$) suggests that there is a significant association between the annual parent merge frequency, allowing for the rejection of the null hypothesis. The child cyclonic eddy merge frequency data also proves to be significant ($\chi^2 = 39.5082, P$-value = 1.5632 e-06, $df = 7$), enabling the rejection of the null hypothesis. Therefore there is a significant difference between the merge frequency within both parent and child cyclonic eddy groups.

Table 6.8. Parent and child cyclonic eddy merge frequency events

<table>
<thead>
<tr>
<th>Year</th>
<th>Parent cyclonic eddy merge frequency</th>
<th>Child cyclonic eddy merge frequency</th>
</tr>
</thead>
<tbody>
<tr>
<td>1992</td>
<td>24</td>
<td>5</td>
</tr>
<tr>
<td>1993</td>
<td>10</td>
<td>12</td>
</tr>
<tr>
<td>1994</td>
<td>5</td>
<td>6</td>
</tr>
<tr>
<td>1995</td>
<td>24</td>
<td>14</td>
</tr>
<tr>
<td>1996</td>
<td>6</td>
<td>9</td>
</tr>
<tr>
<td>1997</td>
<td>23</td>
<td>18</td>
</tr>
<tr>
<td>1998</td>
<td>18</td>
<td>14</td>
</tr>
<tr>
<td>1999</td>
<td>18</td>
<td>14</td>
</tr>
<tr>
<td>2000</td>
<td>21</td>
<td>34</td>
</tr>
</tbody>
</table>
Again a standard chi-square test for independence is used to test the relationship between parent and child cyclonic eddy merge frequency data presented in table 6.8. The resulting $P$-value $\approx 0.0034 \left(x^2 = 21.257, df = 7, Power = 0.9587\right)$ suggests that there is a significant association between the two groups, allowing for the rejection of the null hypothesis. Therefore there is a significant difference between the merge frequency of parent and child cyclonic eddy, in that child cyclonic eddies merge more frequently with other eddies than the parent cyclonic eddies. This may be due to the more temporary nature of child cyclonic eddies as they were prone to merge with their own parents and others in the vicinity.

![Location of cyclonic eddies linked with other oceanic cyclones (1992-2001)](image)

*Figure 6.26. Location of cyclonic eddy links with other oceanic cyclones (1992-2001).*

Cyclonic eddies were also noted to form links with other cyclonic eddies during their lifetime. In this case, a link refers to the process whereby two adjacent cyclonic eddies merge their outer boundaries, without either cyclonic eddy becoming completely absorbed by the other. Linkages such as these may occur for weeks at a time before these cyclonic eddies again disengaged from each other, negating the ability to quantify these events.
Figure 6.26 represents cyclonic eddy links per year during 1992 and 2001. It is important to note that this figure represents the entire time, not individual events, that cyclonic eddies were linked with another mesoscale feature. This continued association partly accounts for the closely associated locations of linking events, as depicted in figure 6.26. Overall, links were seen to be as random as other cyclonic eddy behaviour, equally affected by topographical barriers and other cyclonic and anti-cyclonic eddy features.

After considering cyclonic eddy splitting and merging behaviour it is logical to investigate their translocation speeds. This may be done in two ways: calculating just their poleward trajectory (as per Morrow et al., 2004 and Boebel et al., 2003) or both components of directional translocation. Morrow et al. (2004) and Boebel et al. (2003) both derived a poleward translocation speed of 0.3 km/day for cyclonic eddies within the Cape Basin. To compare their results with those found during this study, the question would be whether the median of the daily poleward translocation speeds of cyclonic eddies differs significantly from the 0.3 km/day calculated by Morrow et al. (2004) and Boebel et al. (2003).

A Mann-Whitney statistical test returns a two-tailed \( P\)-value of < 0.0001, which is considered extremely significant (\( W\)-statistic = -9880.0; corrected for ties), and is detailed in table 6.9. Therefore there is a difference between the mean daily speed of poleward translocation between parent and child cyclonic eddies within these data and those studied by Morrow et al. (2004) and Boebel et al. (2003).

Table 6.9: Descriptive statistics of all cyclonic eddies mean daily speed of poleward translocation

<table>
<thead>
<tr>
<th>Descriptive Statistics</th>
<th>All cyclonic eddy poleward translocation speeds/day</th>
<th>Parent eddy poleward translocation speeds/day</th>
<th>Child eddy poleward translocation speeds/day</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>0.0246</td>
<td>0.3577</td>
<td>-0.1983</td>
</tr>
<tr>
<td>Sample size</td>
<td>252</td>
<td>101</td>
<td>151</td>
</tr>
<tr>
<td>Standard deviation</td>
<td>2.026</td>
<td>0.6999</td>
<td>2.533</td>
</tr>
<tr>
<td>Standard error</td>
<td>0.1276</td>
<td>0.0696</td>
<td>0.2061</td>
</tr>
<tr>
<td>Minimum</td>
<td>-15.429</td>
<td>-0.8571</td>
<td>-15.429</td>
</tr>
<tr>
<td>Maximum</td>
<td>12.00</td>
<td>3.571</td>
<td>12.00</td>
</tr>
<tr>
<td>Median</td>
<td>0.1004</td>
<td>0.2679</td>
<td>-0.0476</td>
</tr>
<tr>
<td>Lower 95 % CI</td>
<td>-0.2256</td>
<td>0.2193</td>
<td>-0.6023</td>
</tr>
<tr>
<td>Upper 95 % CI</td>
<td>0.2747</td>
<td>0.4960</td>
<td>0.2057</td>
</tr>
</tbody>
</table>
However, when the poleward translocation speeds of only the parent cyclonic eddies are compared with the Morrow et al. (2004) and Boebel et al. (2003) results, a Mann-Whitney statistical test returns a two-tailed \( P-value = 0.7297 \). This result is considered not statistically significant (\( W \)-statistic = -205.0; corrected for ties), again represented in table 6.9. Therefore there is no difference between the results found in Morrow et al. (2004) and Boebel et al. (2003) and the poleward speeds of only the parent cyclonic eddies within this study.

The translocation relationship between parents and child cyclonic eddies requires closer scrutiny. Due to the large zonal translocation of cyclonic eddies within the Cape Basin, both latitudinal and longitudinal distances covered by the cyclonic eddies are considered here, instead of just the poleward component.

This dual-vector measurement was achieved by considering the distance covered over time between the formation and demise location of the cyclonic eddies, as determined by a simple geometric triangle equation. Descriptive statistics of parent and child cyclonic eddy speeds of translocation are shown in table 6.10, where parent cyclonic eddies averaged a distance of 2.15 km/day, while child cyclonic eddies moved at approximately 2.98 km/day. These speeds were less than those calculated by Fu (2006), who established that north-westward eddies (anticyclones) progress at a speed of 3 – 4 km/day once they have left the Agulhas Current region.

<table>
<thead>
<tr>
<th>Descriptive Statistics</th>
<th>Parent cyclonic eddy speed of translocation per day (km/day)</th>
<th>Child cyclonic eddy speed of translocation per day (km/day)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean</td>
<td>2.153</td>
<td>2.975</td>
</tr>
<tr>
<td>Sample size</td>
<td>101</td>
<td>151</td>
</tr>
<tr>
<td>Standard deviation</td>
<td>0.8881</td>
<td>2.762</td>
</tr>
<tr>
<td>Standard error</td>
<td>0.08837</td>
<td>0.2247</td>
</tr>
<tr>
<td>Minimum</td>
<td>0.5214</td>
<td>0.1515</td>
</tr>
<tr>
<td>Maximum</td>
<td>6.401</td>
<td>19.714</td>
</tr>
<tr>
<td>Median</td>
<td>2.046</td>
<td>2.235</td>
</tr>
<tr>
<td>Lower 95 % CI</td>
<td>1.977</td>
<td>2.535</td>
</tr>
<tr>
<td>Upper 95 % CI</td>
<td>2.329</td>
<td>3.416</td>
</tr>
</tbody>
</table>
As before, the null hypothesis states that the medians of parent and child cyclonic eddy formation do not differ significantly, whereas the alternative hypothesis shows that they do. Using a Mann-Whitney non-parametric test, the two-tailed P-value is 0.0744, which is considered not quite significant (U = 6613.0; U' = 8638.0; Sum of ranks in parent cyclonic eddies = 11764 and child cyclonic eddies = 20114). Of note is that the P-value is estimated based on a normal approximation and the ‘exact’ method would not be exact, due to tied ranks. Therefore, there does not appear to be a significant difference between the overall translocation speeds of parent and child cyclonic eddies within the Cape Basin.

To summarise, section 6.1 considered the multi-directional trajectory of cyclonic eddies within the Cape Basin. Not only was their overall translocation in different directions but they were also subjected to external influences such as regional topography and mesoscale features, for example Agulhas Rings, which introduced a meandering quality to their translocation.

Section 6.2 covered the lifespan of these cyclonic eddies, where cyclonic eddy 98C clocked the longest lifespan at 840 days before moving out of the study area. Within the boundaries of the study, cyclonic eddy 94A was considered to have the longest lifespan as it survived for 742 days before being lost from altimetric trace. Overall, parent cyclonic eddies were seen to survive for a mean of 254 days, while child cyclonic eddies averaged 188 days. Time lag between formations of child cyclonic eddies and topographical features were also seen to have an impact on the lifespan and subsequent demise of these cyclonic eddies.

Section 6.3 investigated the physical properties of cyclonic eddies, using fortuitous hydrographic data in conjunction with satellite altimetry. A definite trend in the rising of the thermohaline signatures was identifiable as a cyclonic eddy passed through a survey track. However, dedicated in situ measurements are needed to increase the knowledge of cyclonic eddies within this region.

The behaviour of these cyclonic eddies was discussed in section 6.4, where the frequency of merging and linking between eddies was shown to be high. Translocation speeds were similar to results obtained from Morrow et al. (2004) and Boebel et al. (2003) at 0.3 km/day for poleward translocation of parent cyclonic
eddies. However, translocation speeds appeared greater when both directional components were measured, with a mean overall speed of 2.153 km/day for parent cyclonic eddies and 2.975 km/day for child cyclonic eddies.

Having covered the formation and lifestyles of the Cape Basin cyclonic eddies in chapters five and six, chapter seven will discuss possible cyclonic eddy associations with other mesoscale features found within the region.
CHAPTER 7

Cyclonic and anticyclonic eddy association

One of the most interesting questions to arise from a study of these cyclonic eddies is whether the formation of a cyclonic eddy requires the presence of an Agulhas Ring. Before contemplating that association, one needs to be clear on Agulhas Ring identification. As mentioned previously, numerous studies have considered Agulhas Rings (Lutjeharms, 2007; Schmid et al., 2003; Arhan et al., 1999; Gründlingh, 1995; Lutjeharms and van Ballegooijen, 1988; to name but a few). Boebel and Baron, 2003, state that "cyclones, together with Agulhas Rings, manifest themselves in alternating mesoscale patches of positive and negative sea-surface height anomalies", as shown in figure 7.1.

![Figure 7.1: 23 December and 16 December 1997 AVISO altimetric data showing clear anticyclonic and cyclonic eddy signals, represented by white and black arrows respectively (2003).](image)

Therefore, the blatant difference between the negative and positive sea level anomalies (SLA) allows for a confident identification of both cyclonic and anticyclonic eddies. Within this study, anticyclonic eddies that originated from the region of the Agulhas Retrolflection and achieved a large diameter in size, were labelled as Agulhas Rings. Those obvious anomalies with less of a positive difference
from the mean and smaller in diameter were categorised within a generalised group called Positive Sea Level Anomalies (+ SLA), allowing for the potential miss-identification of Agulhas Rings.

Returning to the question of whether it was necessary for an anticyclonic eddy to be present during the formation of cyclonic eddies; table 7.1 looks at the relationship between the two. During the formation of 82 parent cyclonic eddies, positive SLA’s were found in the immediate vicinity in 68 cases. This means that for 82.93% of the time there was an association between the two types of anomalies. Of those 68 cases, 27 were possible Agulhas Rings, while 41 were definite positive SLA’s. Considering the 135 child cyclonic eddies, 121 were formed adjacent to positive SLA’s. Of those 89.63%, 82 were identified as potential Agulhas Rings and 39 were generalised as positive SLA’s. Ironically, there were 14 cases for both parent and child cyclonic eddies where formation occurred without a nearby anticyclonic eddy presence. This confirms that while positive SLA’s may stimulate the formation of cyclonic eddies; they were not critical to the creation process.

Table 7.1: Cyclonic and anticyclonic eddy formation association

<table>
<thead>
<tr>
<th></th>
<th>+ SLA</th>
<th>Agulhas Ring</th>
<th>No association</th>
<th>Percent</th>
</tr>
</thead>
<tbody>
<tr>
<td>Parent cyclonic eddies</td>
<td>41</td>
<td>27</td>
<td>14</td>
<td>82.93</td>
</tr>
<tr>
<td>Child cyclonic eddies</td>
<td>39</td>
<td>82</td>
<td>14</td>
<td>89.63</td>
</tr>
</tbody>
</table>

With such a large percentage there is ample proof that Agulhas Rings and positive SLA’s do influence the formation of cyclonic eddies, negating the need for further statistical analysis. To investigate the individual effects of the two groups, the study area was again split into a northern and southern portion. The theory behind the split was that Agulhas Rings would be more prevalent in the southern portion of the study area, due to the proximity of the Agulhas Retroflection. The northern portion is subject to perennial upwelling and frequent SLA highs, and is generally removed from the influence of the Agulhas Ring Corridor (Garzoli et al., 1999).
Table 7.2: Cyclonic and anticyclonic eddy formation association within the northern and southern portions of the study area

<table>
<thead>
<tr>
<th></th>
<th>+ SLA</th>
<th>Agulhas Ring</th>
<th>No association</th>
<th>+ SLA &amp; Agulhas Ring</th>
<th>Percent</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern parent CE</td>
<td>32</td>
<td>6</td>
<td>11</td>
<td>38</td>
<td>77.55</td>
</tr>
<tr>
<td>Southern parent CE</td>
<td>9</td>
<td>21</td>
<td>3</td>
<td>30</td>
<td>90.91</td>
</tr>
<tr>
<td>Northern child CE</td>
<td>19</td>
<td>12</td>
<td>10</td>
<td>31</td>
<td>75.61</td>
</tr>
<tr>
<td>Southern child CE</td>
<td>20</td>
<td>70</td>
<td>4</td>
<td>90</td>
<td>95.74</td>
</tr>
</tbody>
</table>

From table 7.2 it may be observed that in the south 90.91% of the parent cyclonic eddies and 95.74% of the child cyclonic eddies considered were associated with anticyclonic eddies and Agulhas Rings. However, 77.55% and 75.61% of the northern parent and child cyclonic eddies respectively also interacted with anticyclonic features. Of interest was the dramatic increase in association of the southern child cyclonic eddies with positive SLA’s, with 94 cases of interaction verses only 4 instances with no association. However, the number of cyclonic eddies available for interaction in the region also played a part in this equation, hence the more accurate use of percentages for this result.

Therefore it is possible to determine that anticyclonic features play a very important role in the formation of both parent and child cyclonic eddies, although a few cases exist where formation occurred with no anticyclonic presence. As mentioned, Agulhas Rings form in the southern region of the study area. Hence, when contemplating whether there was a regional impact on anticyclonic association, the study area was split at 30°S. Results showed that there was only a 13% rise in interaction with parent cyclonic eddies between the northern and southern regions. Correlation of anticyclonic features and child cyclonic eddies increased by 20% from north to south, not to dramatic a rise in association at already high levels of 75.61% in the north. Therefore it is safe to state that anticyclonic eddies played an important role in the formation of both parent and child cyclonic eddies.
CHAPTER 8

Conclusions

Cyclonic eddies formed between 1992 and 2001 within the Cape Basin were studied to determine their formation location and frequency, lifestyles and associations with other mesoscale eddies. Satellite altimetry and hydrographic data were utilised in tandem to provide a comprehensive database for analysis.

8.1 Cape Basin cyclonic eddy formation

During the study period, 105 parent and 157 child cyclonic eddies were identified and tracked. The majority of parent cyclonic eddies were seen to form along the West Coast continental shelf, while most of the child cyclonic eddies were formed to the west of the 4000 m isobath. Additionally, there may be possible links between these parent formation locations and known upwelling cells within the region.

The formation location and frequency of these cyclonic eddies were considered on an annual, seasonal and monthly timescale. Results showed that an average of 11 parent and 17 child cyclonic eddies were formed annually; translating into an average seasonal formation of 25 parents and 36 child cyclonic eddies between 1993 and 2001. On a calendar month timescale, a mean of 8 parent and 13 child cyclonic eddies were created. Potential variation within all three timescales was investigated. Most results were considered not statistically significant, except for the monthly formation frequencies of child cyclonic eddy ($\chi^2 = 19.6023$, $df = 11$, $P$-value = 0.04974), which did show a variation between months. Here August, September and October produced the largest number of child cyclonic eddies, but not enough to influence the seasonal statistical results.

The study area was divided into a northern and southern portion at 30°S. This division enabled the determination of whether there was a discernable difference between the influence of perennial upwelling in the north and seasonal upwelling in the south, as well as the documented Ring Corridor originating from the Agulhas Retroflection. Although there was a slight drop in spring formation rates for parent cyclonic eddies and summer for child cyclonic eddies in the northern region of the study area, no results proved to be statistically significant.

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Delving more deeply into the actual formation process of child cyclonic eddies, these eddies were seen to form as parent cyclonic eddies split or occluded a portion of themselves to form a separate entity. In addition, child cyclonic eddies split to form other siblings. To determine any possible triggers for these events, the time lag between the formation of the parent cyclonic eddy and the subsequent splitting event was considered. In total, 61 child cyclonic eddies were formed as first formation cyclonic eddies between 1992 and 2001, followed by 36 child cyclonic eddies formed from secondary splitting events. Logically, more child cyclonic eddies were formed from parent cyclonic eddies within the first few formation categories. After that, the number of sibling splitting events increased with respect to parent splitting events. This is possibly due to the fact that the number of child cyclonic eddies present in the region increased the probability of potential splitting events. Additionally, child cyclonic eddies formed from parent cyclonic eddies were seen to have an average lifespan that was longer than the child cyclonic eddies formed from siblings. However, no significant time lag was determined in any of the above categories.

8.2 Cape Basin cyclonic eddy lifestyles

Once formed, the cyclonic eddies moved along multidirectional paths. Contrary to previous beliefs that cyclonic eddies generally move in a west-southwest direction, only 31.58 % of the cyclonic eddies went in that overall direction. The next most popular direction was a west-northwest overall trajectory, which 27.37 % of the cyclonic eddies followed. Other prominent directions were determined to be westwards, southwest and south-southwest. However, all the cyclonic eddies were seen to meander along their paths, never moving in one direction for too long. This meandering trackline is thought to be influenced by various external factors such as the topography of the region and other mesoscale features in the near proximity.

The lifespan of the cyclonic eddies was examined and results showed that parent cyclonic eddies existed for a mean of 254 days; whereas child cyclonic eddies averaged 188 days. 98C was the longest surviving parent cyclonic eddy during the study period, as it appeared within the data for a period of 840 days before moving westwards out of the study area. Cyclonic eddy 94A remained within the boundaries of the study area and survived for 742 days before being lost from altimetric trace.
Annual and seasonal variation in the life spans of parents and child cyclonic eddies was not found to be statistical significant, nor was any division in the north and south regions of the study area. However, there was a significant variation between parent and child cyclonic eddy life spans for those formed in the north and the south (North: non-parametric Mann-Whitney test returned a two-tailed P-value of < 0.0001; $U = 556.0; U' = 2132.0$. South: Mann-Whitney test: a two-tailed P-value < 0.0001; $U = 1233.0; U' = 4108.0$).

One of the variables affecting the lifespan of cyclonic eddies is thought to be topographical features such as the Walvis Ridge. Overall, 11.5% of the studied cyclonic eddies expired either within the shallow waters before the Walvis Ridge, upon it, or once they had traversed it. 4.2% were unable to cross the next major obstacle, the Mid-Atlantic Ridge, while 1.5% negotiated both ridges. In total, 17% of the cyclonic eddies tracked between 1992 and 2001 were affected negatively by the region's topography, allowing for the assumption that topographical features had an impact on the lifespan and subsequent demise of these cyclonic eddies.

Having considered the formation and lifestyle of these cyclonic eddies; their physical properties were investigated. This was attempted using satellite altimetry in conjunction with fortuitous hydrographic data, collected in pursuit of other projects. Cyclonic eddies were seen to depress between −8 and −42 cm from the mean sea surface height and a definite trend in the rising of the thermohaline signatures was identified when a cyclonic eddy passed through a survey track. Unfortunately this trend was un-measurable due to the incidental assembly of the cyclonic eddy hydrographic data. Therefore, few conclusions may be drawn from these data regarding the completion of a template for the hydrographical identification of cyclonic eddies in the Cape Basin. Future recommendations definitely include dedicated in situ measurements of these cyclonic eddies.

Another facet of their lifestyle that was investigated was the behaviour of these cyclonic eddies. The frequency of merging and linking between eddies was high, with an average merge rate of 16.38 parent and 14 child cyclonic eddies per year. The annual merge frequency was seen to be significantly different within both parent ($x^2 = 28.206, P-value = 0.000202, df = 7$) and child cyclonic eddy groups ($x^2 = 39.5082, P-value = 1.5632 \times 10^{-06}, df = 7$).
The speed of translocation of these cyclonic eddies was measurable and results were shown to be similar to the 0.3 km/day for poleward translocation of parent cyclonic eddies obtained by Morrow et al. (2004). However, when considering both directional components, the translocation speeds appeared greater, with a mean overall speed of 2.153 km/day for parent cyclonic eddies and 2.975 km/day for child cyclonic eddies, similar to results found by Fu (2006).

8.3 Cape Basin cyclone associations

Finally, a possible association between the cyclonic eddies within this study and other mesoscale features, such as anticyclonic eddies, was explored. The results were conclusive. 82.93 % of the parent cyclonic eddies and 89.63 % of the child cyclonic eddies identified during this study were formed adjacent to a positive sea level anomaly (anticyclonic eddy) or Agulhas Ring. In an attempt to account for the propensity of Agulhas Rings in the south, the region was split into northern and southern portions, where association results returned an average of 77 % for northern parent and child cyclonic eddies (77.55 % and 75.61 % respectively) and 93 % for southern parent and child cyclonic eddies (90.91 % and 95.74 % respectively). Causality cannot be proven from these data, but may be explored theoretically or with an appropriate numerical model.

So concludes the descriptive analysis of cyclonic eddies found within the Cape Basin. Future recommendations include a targeted in situ hydrographical study to investigate the physical properties of these cyclonic eddies. Such information will highly benefit the next recommendation, which is a three-prong cyclonic eddy study. An expanded time-series study should include in situ hydrographic data, altimetry and an output from a regional predictive model. Automatic identification and tracking tools, although still in development, may be used to remove individual user bias, as well as expediate the data collection process. A study such as that may allow for a more in-depth analysis of the heat budget and water transport of these cyclonic eddies in the Cape Basin, hence adding to what is known about the essential thermohaline circulation within the oceans of the world.
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APPENDIX A: Daily log of cyclonic eddy 94A

<table>
<thead>
<tr>
<th>Position</th>
<th>Date</th>
<th>Activity</th>
</tr>
</thead>
<tbody>
<tr>
<td>34°S; 16°E</td>
<td>05.01.94</td>
<td>94A begins as ~10 -20cm SSH low, which moves westwards. There is a SSH high directly south of 94A (which reduces in size until it is joined by another SSH high), as depicted in figure 9.1a.</td>
</tr>
<tr>
<td>34°S; 15°E</td>
<td>09.02.94</td>
<td>94A maintains position, while SSH high passes south.</td>
</tr>
<tr>
<td>33.5°S; 14.5°E</td>
<td>02.03.94</td>
<td>94A moves northwards as it is pushed by the SSH high.</td>
</tr>
<tr>
<td>33.5°S; 13.5°E</td>
<td>16.03.94</td>
<td>94A gathers strength ~20 -30 cm.</td>
</tr>
</tbody>
</table>

![Figure 9.1. Altimetric snapshots highlighting cyclonic eddy 94A, as indicated by a white arrow: a) a newly formed cyclonic eddy 94A (5 Jan 1994); b) 94A's position on the 23 March 1994 (AVISO) data displayed using Matlab.](image-url)
<table>
<thead>
<tr>
<th>Date</th>
<th>Longitude</th>
<th>Latitude</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>33.5 °S; 13 °E</td>
<td>23.03.94</td>
<td></td>
<td>94A gathers strength and strength to 20 cm, and merges with small coastal cyclonic eddy. A change in overall shape is noticeable in figure 9.1b.</td>
</tr>
<tr>
<td>33.5 °S; 14 °E</td>
<td>30.03.94</td>
<td></td>
<td>94A stops and remains in place.</td>
</tr>
<tr>
<td>33.5 °S; 14 °E</td>
<td>04.05.94</td>
<td></td>
<td>94A grows in diameter (1 x 1 degree of latitude and longitude).</td>
</tr>
<tr>
<td>33.5 °S; 14.5 °E</td>
<td>11.05.94</td>
<td></td>
<td>94A incorporates another SSH low and deepens to approximately 20 to 35 cm, as shown in figure 9.2a.</td>
</tr>
<tr>
<td>33.5 °S; 13.5 °E</td>
<td>18.05.94</td>
<td></td>
<td>94A strengthens in depth to 20 cm and is flattened sideways by an approaching SSH high. Small cyclonic eddy, 94A, is occluded off the southwest side of 94A, as shown in figure 9.2b.</td>
</tr>
</tbody>
</table>

![Figure 9.2: Altimetric snapshots highlighting cyclonic eddy 94A, as indicated by a white arrow: a) 94A's position on the 11 May 1994, b) 94A's position on the 18 May 1994 (AVISO data displayed using Matlab).](image-url)
33.0°S; 13.0°E 15.06.94 94A is pushed north-westwards by SSH high. Diameter of the main 94A body is 31°S to 34°S, while the sea surface height is approximately 20 to -40cm. Nearby 94A1 occludes a small eddy, named 94Aii.

33.0°S; 13.0°E 29.06.94 94A joins another SSH low found to the south, while the SSH high still chases from the south-east.

33.0°S; 13.0°E 06.07.94 94A plus new SSH low form dipole cyclone spanning 31.5°S to 37.5°S, moving in westerly direction.

33.0°S; 13.0°E 20.07.94 Dipole SSH low system joins smaller 94Aii (33.5°S 9.5°E, at 20cm SSH) from the west. Accompanying SSH high reduces in size, as represented by figure 9.3a.

Figure 9.3: Navigable snapshot highlighting cyclonic eddy 94A, as indicated by a white arrow. a) 94A's position on the 20 July 1994. b) 94A's position on the 24 August 1994 (AVISO data displayed using Matlab).
33°S; 12°E  27.07.94  Dipole SSH low splits, while still retaining links with a
SSH low. SSH high moves southwards.

33°S; 11°E  03.08.94  94A joins westerly SSH low.

33°S; 11°E  10.08.94  94A led by lower original dipole SSH low and moves
south-westward around SSH high (possible Agulhas
Ring).

33°S; 11°E  17.08.94  Small cyclonic eddy 94Aiii is occluded from the west of
a linked 94A and 94Aii.

33°S; 11°E  24.08.94  Main SSH low still in position but adjoining a SSH
high. Both move in a south-westerly direction, as shown
in figure 9.3b.

34°S; 10°E  07.09.94  Northwest SSH low splits off, while another SSH low
joins 94A from the south.

Figure 9.4: Mosaic snapshots highlighting cyclonic eddy 94A, as indicated by a white arrow. a)
94A's position on the 14 September 1994. b) 94A's position on the 21 September 1994 (AVISO data
displayed using Matlab).
34°S; 10°E  14.09.94  94A starts consolidating (back into a circular shape) and, reducing in depth, moves westward behind a SSH high system. 94Aiv begins to occlude from the northeast corner of the 94A system, as shown in figure 9.4a.

33.5°S; 10°E  21.09.94  94Aiv splits off the northwest portion of 94A (32°S 12.5°E) and 94Aiii moves away from 94A, as shown in figure 9.4b.

34°S; 7°E  19.10.94  94A deepens with presence of 3 SSH highs (northwest, west, southwest positions).

34°S; 7°E  26.11.94  94A occludes 94Av from its southwest corner.

34°S; 7°E  09.11.94  94A re-joins with 94Av (34.5°S 5.5°E), with a large SSH high directly above the joining.

Figure 9.5: Arealic snapshots highlighting cyclonic eddy 94A, as indicated by a white arrow: a) 94A's position on the 7 December 1994. b) 94A's position on the 4 January 1995 (AVISO data displayed using Matlab).
34.5°S; 6°E 23.11.94 94A moves around SSH high in northwest direction (-10 to -20 cm), reducing in size.

34°S; 5°E 07.12.94 94A reduces in size as 94Aiv moves closer. Both SSH lows are wrapped around a SSH high, as shown in figure 9.5a.

34°S; 4.5°E 14.12.94 94A (-10 cm+) merges with 94Aiv (previously split on 21.09.94). Dipole forms around SSH high (32°S 5.5°E).

32.5°S; 4°E 04.01.95 94A starts to split from 94Aiv, as shown in figure 9.5b.

32.5°S; 4°E 11.01.95 94A completely split from 94Aiv and moves westward (-10 cm), as depicted in figure 9.6a.

Figure 9.6: A multi-year plot highlighting cyclonic eddy 94A, as indicated by a white arrow. (a) 94A's position on the 11 January 1995. (b) 94A's position on the 12 July 1995 (AVISO data displayed using Matlab).
33 °S; 4 °E 25.01.95 -10cm 94A resembles a fried egg as it propagates westwards.
33 °S; 1 °E 08.03.95 94A almost less than 10cm as it encounters a SSH high.
32 °S; 4 °W 12.07.95 94A resembles fried egg shape again as it continues on its westward path, as shown in figure 9.6b.
33 °S; 10.5 °W 24.01.96 94A finally reduces in SSH height to less than 10cm, as shown in figure 9.7. The satellite altimetric trace of 94A is lost after this point.

Figure 9.7: 24 January 1996 altimetric snapshot highlighting the ending of cyclonic eddy 94A, as indicated by a white arrow (AVISO data displayed using Matlab).
APPENDIX B: Annual Formation Locations

Figure 9.8: Cape Basin cyclonic eddy parent and child formation location during October - December 1992.

Figure 9.9: Cape Basin cyclonic eddy parent and child formation location during January - December 1994.
Figure 9.10: Cape Basin cyclonic eddy parent and child formation location during January - December 1995.

Figure 9.11: Cape Basin cyclonic eddy parent and child formation location during January - December 1996.
Figure 9.12: Cape Basin cyclonic eddy parent and child formation location during January - December 2000.

Figure 9.13: Cape Basin cyclonic eddy parent and child formation location during January - December 1998.
Figure 9.14: Cape Basin cyclonic eddy parent and child formation location during January - December 1999.

Figure 9.15: Cape Basin cyclonic eddy parent and child formation location during January - December 2001.
APPENDIX C: Lifespan of cyclonic eddies within the Cape Basin

Figure 9.16: Cape Basin cyclonic eddy parent and child lifespan in 1992.

Figure 9.17: Cape Basin cyclonic eddy parent and child lifespan in 1993.
Figure 9.18: Cape Basin cyclonic eddy parent and child lifespan in 1994.

Figure 9.19: Cape Basin cyclonic eddy parent and child lifespan in 1995.
Figure 9.20: Cape Basin cyclonic eddy parent and child lifespan in 1996.

Figure 9.21: Cape Basin cyclonic eddy parent and child lifespan in 1997.
Figure 9.22: Cape Basin cyclonic eddy parent and child lifespan in 1998.

Figure 9.23: Cape Basin cyclonic eddy parent and child lifespan in 1999.
Figure 9.24: Cape Basin cyclonic eddy parent and child lifespan in 2006.
APPENDIX D: Corresponding hydrographic and altimetric data

Cyclonic eddy 99II, as identified by black arrows:

Figure 9.25. Cyclonic eddy 99II. a) temperature section through the region; b) thermlaline (TS) plot of the section; c) map showing hydrographical stations as gathered by SAOCO; and d) salinity section through the region.

Figure 9.26. Altimetry image 30 and 21 June 1999 (AVISO, 2005). Cyclone 99II is indicated by a black arrow.
Cyclonic eddy 99S (white arrows):

Figure 9.27: Cyclonic eddy 99S: a) temperature section through the region, b) thermohaline (TH) plot of the section, and c) map showing hydrographical NADP/D and SAAE stations. Salinity information was not collected.

Figure 9.28: Altimetry image of Dec and 24 November 1999 (AVISO, 2003). Cyclone 99S is indicated by a white arrow.
Cyclonic eddy 00Λ (black arrows) and 00Λi (white arrows):

Figure 9.29. Cyclonic eddy 00Λ and 00Λi: a) temperature section through the region; b) thermohaline (TH) plot of the section; c) map showing hydrographical SAVW stations; and d) salinity section through the region.

Figure 9.30. Altimetry image 28 and 21 February 2001 (AVISO, 2005). Cyclones 00Λ and 00Λi are indicated by a black and white arrow respectively.
Cyclonic eddy 00B (white arrows):  

![Image of temperature and salinity sections through the region](image)

**Figure 9.31:** Cyclonic eddy 00B: a) temperature section through the region, b) thermohaline (TH) plot of the section, c) map showing hydrographical SAV stations, and d) salinity section through the region.

![Image of satellite image](image)

**Figure 9.32:** Altimetry image 02 February and 26 January 2000 (AVISO, 2005). Cyclones 00B is indicated by a white arrow.

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Cyclonic eddy U1E (white arrows):

Figure 9.33: Cyclonic eddy U1E: a) temperature section through the region; b) thermohaline (TS) plot of the section; c) map showing hydrographical SADCO stations; and d) salinity section through the region.

Figure 9.34: Altimetry image 28 and 21 February 2001 (AVISO, 2003). Cyclone O1E is indicated by a white arrow.

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Retroflection area cyclonic eddy A (black arrows) and B (white arrows):

Figure 9.35: Retroflection cyclonic eddy A; a) temperature section through the region, b) thermohaline (T-S) plot of the section, c) map showing hydrographical WPO stations, and d) salinity section through the region.

Figure 9.36: Retroflection cyclonic eddy B; a) temperature section through the region, b) thermohaline (T-S) plot of the section, c) map showing hydrographical SAVE stations, and d) salinity section through the region.
Figure 9.37: Altimetry image 28 and 21 February 2004 (MISCO, 2005). Retroreflection cyclonic eddies A and B are indicated by a black and white arrow respectively.