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WIND-STRESS VARIABILITY OVER THE BENGUELA UPWELLING SYSTEM

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“There is only one good, knowledge, and one evil, ignorance.”

- Socrates

Abstract

Regional wind-stress variability over the Benguela Upwelling System is described using 16 months (01 August 1999 – 29 November 2000) of satellite derived QuikSCAT wind data. The QuikSCAT data are compared to the climatologies presented by Kamstra (1985) and Bakun and Nelson (1991), as well as the long-term climatology (1968-1996) of the surface vector wind speed field off the coast of southern Africa, as derived from the 2.5° resolution NCEP/NCAR reanalysis dataset. Broad scale similarities are found between the QuikSCAT and the long-term NCEP/NCAR climatology (1968-1996) data sets. This allows one to have confidence in using this scatterometer data to investigate details of spatial and temporal variability over the Benguela System.

During summer, wind-stress maxima are found at approximately 17, 29 and 34°S. These maxima strengthen in late summer. The seasonal northward migration of the South Atlantic Anticyclone becomes apparent in late autumn, when the strongest wind-stress occurs north of 28°S. A significant wind-stress minimum is observed to develop slightly north of Cape Columbine (33°S) during autumn. To the north (10-23.5°S) the Benguela is characterised by relatively strong south-easterly wind-stress during winter. To the south (24-35°S) the Benguela is characterised by relatively weak westerly to south-westerly wind-stress during winter. A southward migration of south-easterly wind-stress is observed during early spring. By November the entire Benguela Upwelling System is once again characterised by southerly to south-easterly wind stress.

Wind-stress variability is investigated using both a type of artificial neural network, known as the Kohonen Self Organising Map (SOM), as well as a wavelet analysis. Two independent SOM studies are conducted. The first study produced a 6x4 SOM output array, which is used to examine seasonal variability as well as the temporal evolution of two synoptic-scale wind events. For the second study both a SOM and a wavelet analysis are applied to an extracted data set to find that the system can be divided into six discrete wind regimes, 10-15°S; 15.5-18.5°S; 19-23.5°S; 24-28.5°S; 29-32.5°S; and 33-35°S. The wavelet power spectra for these wind cells span a range of frequencies from 4 to 64 days, with each region appearing to contain distinct periodicities. To the north, 10-23.5°S, the majority of the power occurs during winter, with a 6-16 day periodicity. Further south, 24-35°S, the majority of the power occurs in the summer. Here a bi-modal distribution occurs, with peaks of 6-16 and 35-40 days.

Lastly a case study sequence of the spatial distribution of wind-stress, wind-stress curl and SST, at a location off the west coast of southern Africa (25-30°S and 12-17°E), is discussed in relation to an intense, upwelling favourable, wind event that occurred from 11-20 February 2000.

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CHAPTER 1

INTRODUCTION

The Benguela Current, which is the eastern boundary current of the South Atlantic subtropical gyre, is one of four major eastern boundary current regimes of the World Ocean. The oceanography off the western coast of southern Africa is dominated, like the regions off the west coasts of South America (Strub *et al.*, 1998), North America (Hickey, 1998), and north-west Africa (Barton, 1998), by a wind-driven coastal upwelling system. Although Dengler (1985) estimated that the combined surface for all four upwelling systems accounts for only 0.1% of the total surface area of the world oceans, these regions account for almost 30% of the world's total fish catches. Thus the Benguela Ecosystem is extremely important, economically speaking, to the countries of southern Africa, and as a result a large amount of marine research has been directed towards the analysis and understanding of the ecosystem and its variability.

The local wind systems of eastern boundary regions are controlled by semi-permanent subtropical high-pressure systems, an example of which is the South Atlantic Anticyclone. These systems account for the predominantly equatorward component of wind-stress found within these regions. At the poleward margins of these zones, however, meridional shifts in the subtropical high-pressure systems result in the equatorward wind forcing reversing seasonally to become poleward during the winter months (Hill *et al.*, 1998).

The Benguela Upwelling System is situated on the wide south-western shelf of Africa, from Cape Agulhas (34.5°S) in the south (Lutjeharms and Stockton, 1987) to the Angolan-Benguela Front situated at approximately 14-18°S in the north (Shannon *et al.*, 1987). The Benguela system is unique among the four major upwelling systems of the world, as it contains a warm-water boundary at both its southern most and its northern most extremity. Due to the fact that the tip of Africa extends only to approximately 35°S, the Agulhas Current system is able to extend far enough westwards that eddies and rings, shed by the current, are then able to interact

with the southern boundary (Gordon *et al.*, 1987) as well as the offshore boundary (Shillington *et al.*, 1992).

The semi permanent high-pressure system, found over the subtropical South Atlantic, otherwise known as the South Atlantic Anticyclone (SAA), in conjunction with the low-pressure system that develops over southern Africa during the summer months, exerts a strong atmospheric influence over the Benguela Upwelling System. The prevailing winds in the region blow from the south and southeast. This in turn drives an offshore Ekman transport which results in the upwelling of cold, nutrient-rich water onto the continental shelf (Peterson and Stramma, 1991). This upwelling exists with much spatial and temporal variability, all along the southwest coast of Africa, from Cape Agulhas to Cape Frio (Lutjeharms and Meeuwis, 1987). The dominant south-easterly wind regime is modulated south of the Orange River, at 28.5°S, by the passage of east-moving mid-latitude cyclones, and associated frontal systems, south of the continent. During the autumn months the effect of this is usually seen in a weakening of the South Atlantic Anticyclone and an abatement of south-easterly winds along the coast. During the winter months, however, these mid-latitude cyclones can, at times, result in gale force north-westerly and south-westerly winds. Preston-Whyte and Tyson (2000) suggest that this modulation occurs in cycles of 2-8 days.

It is the intention of this dissertation to contribute to a better understanding of the spatial and temporal variability of wind-stress and wind-stress curl over the Benguela Upwelling System, at synoptic to intra-seasonal time scales of 4–64 days. This period range of 4–64 days is close to, but not identical with, the definition of 6–70 days adopted by Kiladis and Mo (1998).

The objectives for this dissertation are two fold. Firstly, an investigation into the variability of wind-stress and wind-stress curl over the Benguela upwelling system for the period of the 01 August 1999 to the 29 November 2000, using a data set derived from the QuikSCAT Level 3 data set. This derived data set consisted of the u and v wind-stress components as well as the derivative wind-stress fields, namely curl and divergence. These data were sub sampled to a spatial domain of 10°S to 35°S and 5°E to 20°E, and were smoothed temporally, to 2-day composites, and spatially, to a one-

half degree resolution. The less traditional statistical techniques of Self Organising Maps and wavelet analysis are used to conduct this investigation. The second objective is to investigate the atmospheric evolution of an intense, upwelling favourable, wind event that occurred from 11–20 February 2000. The temporal succession and magnitude of such events has been shown in previous work (e.g. Roy *et al.*, 2001) to have direct links to phytoplankton abundance, primary production and fish recruitment in the southern Benguela Ecosystem

With this in mind, the dissertation proceeds in the following manner; **Chapter 2** consists of a literature review of the Benguela upwelling regime. This chapter includes an overview of both the atmospheric forcing and hydrography of the region. It also includes an overview of the observed variability, from synoptic to interannual time scales, found within the Benguela system.

Chapter 3 is concerned with the SeaWinds instrument, the original NASA scatterometer data, as well as the derived data set, investigated here. It commences with a brief history of scatterometers, which is then followed by an introduction to the technical specifications of both the SeaWinds scatterometer and the QuikSCAT Satellite. Lastly this chapter concentrates on the data produced by the SeaWinds instrument, the processing thereof, and work done in terms of calibration by authors such as Chelton *et al.* (2001).

Chapter 4 discusses the rationale behind both Wavelet Analysis and a type of artificial neural network otherwise known as the Kohonen Self Organising Map.

The results and discussion are contained within **Chapter 5**, while all conclusions derived from this dissertation are found within **Chapter 6**.

CHAPTER 2

A REVIEW OF THE BENGUELA UPWELLING REGIME

GEOGRAPHY & BATHYMETRY

The southwest coastline of Africa is relatively straight with only a few major capes and embayments, most of which occur in the southern region. Examples of these features are False Bay, the Cape Peninsula, Cape Columbine and Saldanha Bay, Lüderitz, Walvis Bay, and Cape Frio. The region contains only four river systems of any significance, namely the Orange and Cunene Rivers to the north and the Olifants and Berg Rivers to the south. Because of the extremely arid nature of the region, the fresh water input into the coastal ocean is relatively small, with exceptions occurring during times of summer flooding in the interior (Shillington, 1998).

The most marked feature of the bathymetry of the South Atlantic is the Mid-Atlantic ridge. This ridge runs from north to south and between 10 and 20°W. In doing so it roughly divides the South Atlantic into two large parallel troughs. Secondary ridges segment the eastern trough into the Angola, Cape, and Agulhas Basins. The most significant of these secondary ridges is the Walvis ridge (Figure 2.1). This ridge runs along a northeast-southwest axis and connects the Mid-Atlantic ridge to the African continent. The Walvis ridge serves to separate the Angola Basin from the Cape Basin. Further south, the Cape Rise separates the Cape Basin from the Agulhas Basin, which straddles the division between the Atlantic and Indian Oceans.

The Agulhas Bank occupies the area to the south of South Africa. Here the continental shelf is approximately 230km wide. As one moves north, towards Cape Town and Cape Columbine the shelf begins to narrow to be roughly 40km wide off the Cape Peninsula. West of the Orange River the shelf widens to approximately 180km, before once again narrowing to approximately 30km north of 17°S (Shannon and Nelson, 1996). Double shelf breaks are common along the southwest coast of Africa. An example of one of these double shelf breaks is found off the coast at Walvis Bay. Here, there are inner and outer breaks beginning at depths of about 140 and 400m respectively (Shannon and Nelson, 1996). Several submarine canyons exist between

31 and 35°S, the most significant of which is the Cape Canyon. This canyon runs from a head 60km west of Cape Columbine in a southerly direction (Shillington, 1998).

All eastern margins exhibit particular coastal features, similar to those mentioned above. These features are important in the local forcing of any one particular upwelling regime. For example capes and promontories act to trigger instabilities within upwelling jets, which may in turn result in filament formation. Localised or intensified coastal upwelling can also often be attributed to particular promontories thus making them important with respect to spatial variability within upwelling regimes (Hill *et al.*, 1998). Canyons, such as the Cape Canyon, that intersect the continental slope can have a number of significant influences and can modify the local circulation (Hill *et al.*, 1998). Dynamically, their most important effect is to allow the geostrophic flow on steep slopes to be broken locally, thus resulting in an increase in cross-slope motion. These canyons also provide possible sites for intensified upwelling, which typically occurs on their equatorward side (Hill *et al.*, 1998).

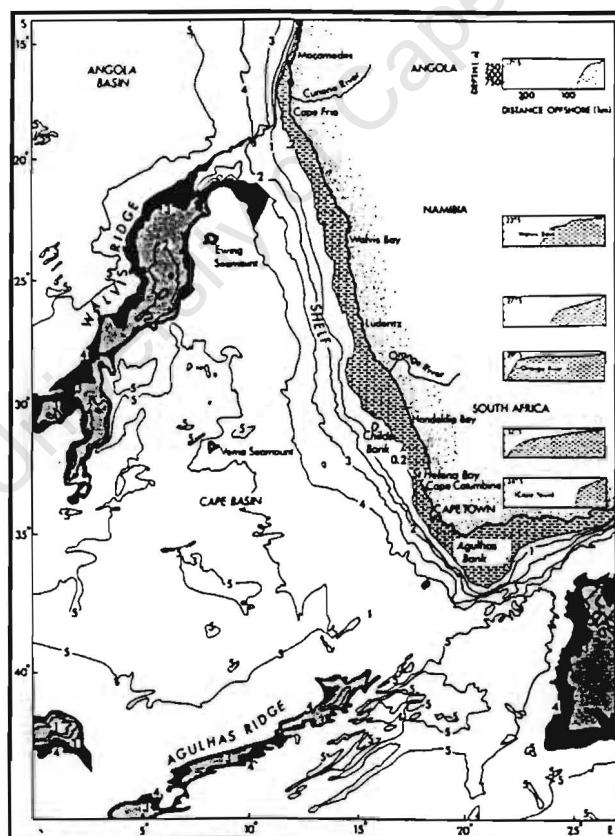


Figure 2.1 Bathymetry of the Southeast Atlantic Ocean with profiles of the shelf at selected locations (after Shannon and Nelson, 1996.)

ATMOSPHERIC FORCING

The Subtropical South Atlantic Anticyclone

The curved anticyclonic airflow over the Benguela region is associated with the South Atlantic Anticyclone (SAA). Strong heating of the continent in summer leads to relatively lower pressures in the interior, viz. a heat low. During the summer, when this heat low contrasts sharply with the SAA, the pronounced zonal pressure gradients produce strong, upwelling favourable, southerly and south-easterly winds, along the south-west coast of Africa. These winds are laterally confined by both the topography of the continental escarpment and the thermal barrier of the heat low (Kamstra, 1985). The SAA is maintained throughout the year, with seasonal differences in pressure being of the order of 3-4hPa. Schell (1968, in Shannon, 1985) found the SAA to be centred at approximately 30°S:5°E in summer and 26°S:10°E in winter, a shift of 4° of latitude and 5° of longitude. In contrast to this, Tyson (1986) found the SAA to have a seasonal 6° shift of latitude, reaching its northern and southern extremities in May and February respectively, and a shift of 13° of longitude, reaching an extreme westward position in August. The pressure changes substantially over the African subcontinent from the well-developed continental heat low during the summer months, to a weak high for the duration of the winter months, as the Intertropical Convergence Zone (ITCZ), and the heat low, move northward. Consequently the pressure gradient along the west coast is seasonally variable.

The monthly shifts in position are up to double those of its latitudinal variation and do not materially affect the weather of southern Africa (Preston-Whyte and Tyson, 2000). On the scale of days, however, the SAA may ridge eastward and to the south of the continent. Extended ridging of the SAA may lead to the budding off of eastward tracking migratory anticyclones. These anticyclones drift eastwards, south of South Africa, into the Indian Ocean prior to being subsumed in to the South Indian High. In doing so they may strongly affect the southern Benguela by producing, for example, strong south-easterly winds over the Western Cape.

Mid-Latitude Westerly Disturbances

The dominant south-easterly wind regime is modulated within the southern Benguela by the passage of east-moving mid-latitude cyclones, and associated frontal systems, south of the continent. These cells of low-pressure are formed within the westerly wind belt, between 35°S and 60°S, ahead of planetary waves in the subtropical jet stream. During the summer months the effect of this is usually seen in a weakening of the SAA and an abatement of south-easterly winds along the coast. During the winter months, however, these mid-latitude disturbances can, at times, result in gale force north-westerly and south-westerly winds, which tend to occur in cycles of 2-8 days (Preston-Whyte and Tyson, 2000). Another type of westerly disturbance that significantly affects the weather over South Africa is the cut-off low. These cold-cored depressions start as troughs in the upper westerlies prior to deepening into closed circulations extending downward to the surface and which become displaced equatorward of the westerly wind belt. Cut-off lows are unstable, baroclinic systems that slope to the west with increasing height and are associated with strong convergence and vertical motion, particularly while they are deepening. These disturbances account for many of the flood-producing rain events observed over South Africa. Such was the case in Laingsburg in 1981 and in KwaZulu-Natal in 1987 (Preston-Whyte and Tyson, 2000).

Coastal Lows

Often associated with the trailing edge of eastward tracking migratory anticyclones, is the appearance of shallow low-pressure cells, which form near Lüderitz. These coastally trapped disturbances (CTD), or coastal lows, are shallow, mesoscale disturbances that are trapped vertically by a strong subsidence inversion and horizontally, through Coriolis effects, against the steep escarpment that fringes the subcontinent (Reason and Jury, 1990). As well as acting to constrain these perturbations horizontally, the escarpment, which forms a roughly semi-circular barrier of radius 1000km with elevations exceeding 1000m, acts to steer the coastal lows around the coast of southern Africa (Jury *et al.*, 1990b). Apart from the southern African example, CTD have been noted to occur off the coasts of California and south-eastern Australia. In contrast to the CTD found off southern Africa, which propagate as coastal lows, all of the Californian and Australian CTD cases reported to

date have propagated as coastal ridges (Reason and Steyn, 1990). Reason and Steyn (1990) showed that even though the fundamental dynamics were the same for all three regions, the regional differences in forcing and boundary conditions lead to diversity in the manifestation of the disturbances in each area. These differences include variations in the width of the coastal mountains, the strength and height of the inversion and variations in the scale of forcing.

The Coastal Low Workshop (CLW, 1984; in Taljaard, 1995) defined a coastal low as “a small area of relatively lower pressure which appears in the lower levels of the atmosphere (below 700hPa) along the coast... the essential features of the coastal low are the pressure minimum and the shallowness of the system”. According to this definition, wave depressions on cold fronts and the surface reflections of deep troughs and cyclonic vortices, such as cut-off lows, are ruled out. Taljaard (1995) identified a number of types of pressure minima found along the west coast of Southern Africa, as indicated in Figure 2.2. Of these, the most common type of coastal low along the south and south-east coasts in summer and the almost exclusive type in winter is shown in Figure 2.2 C&D. This is the characteristic system that precedes cold fronts and some intercell troughs between anticyclones, when they advance eastwards at the latitude of the south coast of South Africa. These Coastal lows tend to have a general seaward extent of between 50 and 150km (Jury *et al.*, 1990b and Reason and Jury, 1990). Oilrig data reported by Hunter (1987) indicate that coastal lows tend to grow in size from a diameter of approximately 100km on the west coast to 150km on the south coast. Walker (1984) found the mean propagation speed of coastal lows to be approximately 6m.s^{-1} , while Schumann (1983) reported speeds of over 20m.s^{-1} on the east coast. Heydenrych (1987) studied the climatology of coastal lows along the west coast between Langebaan and Cape Agulhas. Heydenrych was able to select 13 fairly clear-cut examples of coastal low passages at Cape Town for the 12-month period of November 1984 to October 1985. From these cases he was able to derive mean values of climatological parameters for the period starting 36 hours before passage at Cape Town till 24 hours after passage. In this way he was able to construct time and space sections for coastal lows as they approached, passed and receded from the vicinity of Cape Town.

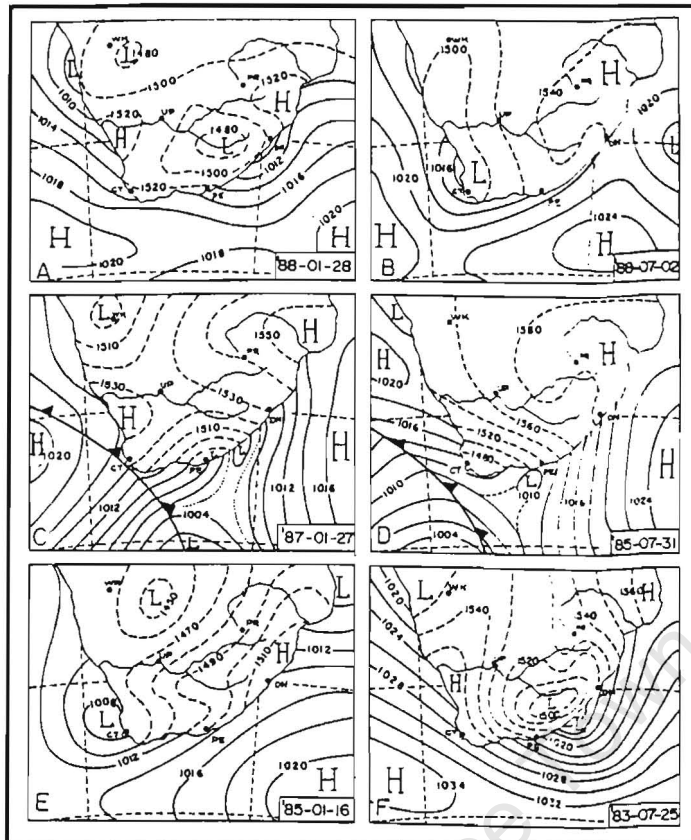


Figure 2.2 Types of coastal pressure minima. Types A and B have decidedly lowest pressure along the coast. Types C and D are the most common and are characteristic of the systems that precedes cold fronts. Types E and F are associated with upper lows. (after Taljaard, 1995)

The primary cause of coastal low formation is considered to be the adiabatic warming of the air as it descends from the interior plateau to sea level, or adjacent low ground, and the entrainment of this buoyant air mass with its associated cyclonic vorticity into the marine layer. Warming in the lower coastal atmosphere reaches a maximum when offshore drainage of the air from the plateau reaches a peak, typically associated with strong offshore adiabatic “Berg” winds and a lowering of the inversion towards the surface. The reverse process occurs when the cool maritime air behind the coastal low displaces the preceding warm continental air mass causing the inversion to once again rise (Taljaard, 1995). Also important to the formation of coastal lows is conservation of potential vorticity. In basic terms this implies that anticyclonic vorticity of an air parcel increases on the upward side of a mountain range or high ground when it is forced to rise and accelerate over the obstruction. On the other hand cyclonic vorticity is generated on the leeward side where the air parcel decelerates and its depth increases. If the flow is from a flat plateau to low ground or to sea level, the cyclonic

vorticity will increase, thus resulting in the generation of a leeside low or trough. With this in mind, the fact remains that the pressure at the lower level is determined almost entirely by the vertical density distribution as prescribed by the hydrostatic equation (Taljaard, 1995). A series of studies have been carried out to determine whether a coastal low fits the theory of Kelvin waves propagating on a temperature inversion. Reason and Jury (1990) concluded that coastal lows, at least the type associated with ridging anticyclones and frontals systems, passing south of South Africa, essentially propagated around the coast as an internal Kelvin wave trapped by the southern African escarpment.

Berg Winds

Another large-scale feature of the meteorology of virtually the entire Benguela region is the occurrence of “Berg” winds, the effects of which are most obvious during autumn and winter (Shannon, 1985). The term Berg winds was first used by Steward (1901) to refer to the hot winds, which occasionally blew in winter from the landward side over the coastal belt of the Cape Colony. In later years it became known that the same phenomenon also occurs all along the Namibian coastal zone, the southern and eastern Cape, as well as in Natal and even in southern Mozambique. These winds tend to be closely associated with coastal lows, however they can also occur when no coastal low is present (Taljaard, 1995). There are a number of other misconceptions associated with Berg winds, one of which is the idea that Berg winds are confined to a narrow strip of land along the coast. Temperatures will tend to rise well above normal wherever downslope seaward flow, off the escarpment, exists. Places that experience Berg winds inland from the coast are, for example, Beaufort West, Graaf-Reinet, Somerset East, and Worcester. Adiabatic warming will reach a maximum where both elevation changes as well as air mass contrasts are greatest. This is usually, but not necessarily, at the coast.

Land-Sea Breezes

Land-sea breezes are common along the coast north of Cape Columbine. To the south, these local winds are often masked by larger scale winds, such as the strong southeasterly winds often experienced over Cape Town. Hart and Currie (1960) have very

adequately described the diurnal modulation of the coastal wind. The contrast between the morning and afternoon winds, in particular those at Walvis Bay, is illustrated in Figure 2.3. In this figure, the marked intensification of the coastal winds during the day, with the backing of the northerly winds to a more south-south-westerly direction, being evident. In addition to land-sea breezes, the central Namib Desert experiences both valley and plain-mountain and mountain-plain winds. These local to regional wind phenomena occur frequently during summer and winter, respectively (Preston-Whyte and Tyson, 2000). The diurnal pulsing of local and regional winds and the seasonal variation exhibited therein have a marked influence on the coastal upwelling dynamics of much of the Benguela region, the northern part in particular.

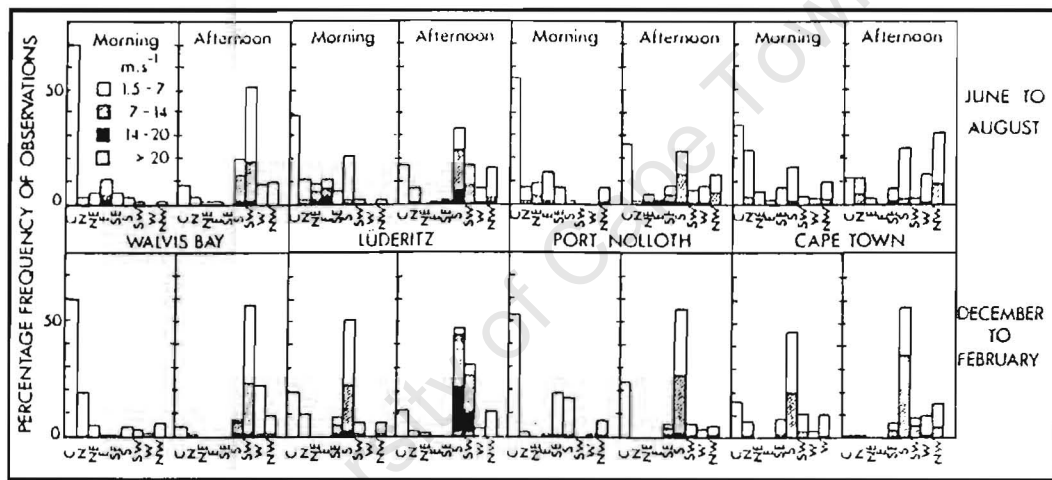


Figure 2.3 Percentage frequency and speeds of coastal winds in winter and summer. (after Hart and Currie, 1960)

Wind-Stress and Wind-Stress Curl

Wind-stress is the force exerted by the wind on the sea surface and is responsible for wave and current generation and vertical mixing. Wind-stress can be calculated using the bulk aerodynamic formula

$$\underline{\tau} = \rho C_d \underline{V}_{10} |\underline{V}_{10}|,$$

where τ is the stress vector, ρ is the density of air, C_d is a dimensionless drag coefficient, $\underline{V}_{10}$ is the wind velocity at a height of 10m above the sea surface, and $|\underline{V}_{10}|$ is the wind speed (Large and Pond 1981). Wind-stress curl, a derivative of wind-stress, is a fundamental forcing agent for dynamic ocean processes. It stands out as the external input term in ocean model formulations expressed in terms of vorticity balance. The same process when considered from vertical structure in the flow field appears as the divergence of surface Ekman transport. Wind-stress curl thus drives “Ekman pumping” and associated vertical advection in regions seaward of the immediate effects of coastal boundaries (Bakun and Nelson, 1991). Using historical wind reports extracted from the NOAA National Climatic Data Centre’s (NCDC) global file of surface marine observations, Bakun and Nelson (1991) attempted to define characteristic seasonal distributions of wind-stress curl over the four major eastern boundary current regions of the World Ocean. From their investigation, these authors were able to conclude that eastern boundary current regions are characterised by anticyclonic wind-stress curl in the offshore portions. Near the coastal boundary, the wind-stress curl tends to be cyclonic, particularly during the coastal upwelling seasons. These areas of cyclonic curl typically extend 200 to 300 km offshore, and are particularly intense adjacent to coastal promontories. The coastal cyclonic curl regions tend to have their greatest alongshore extents during the summer seasons in the respective hemispheres. For the Benguela Current region in particular, the coastal area of wind-stress curl appears as a wedge extending from the vicinity of 20°S, narrowing in offshore extent toward the south (Figure 2.4). The wedge has its most limited poleward extent during the winter months, when it reaches approximately Alexandra Bay (28°S). By early spring the wedge has extended south to Cape Columbine, and beyond Cape Point by late spring, thus reflecting a maximum in upwelling-favourable winds. The wedge tends to remain in this position until it begins to retreat northward

in late autumn. Equatorward of 20°S , the distributions are less distinct. However, one persistent feature is an area of anticyclonic curl, some hundreds of kilometres in diameter lying about 200 km offshore, and extending northward of Cape Frio (Bakun and Nelson, 1991).

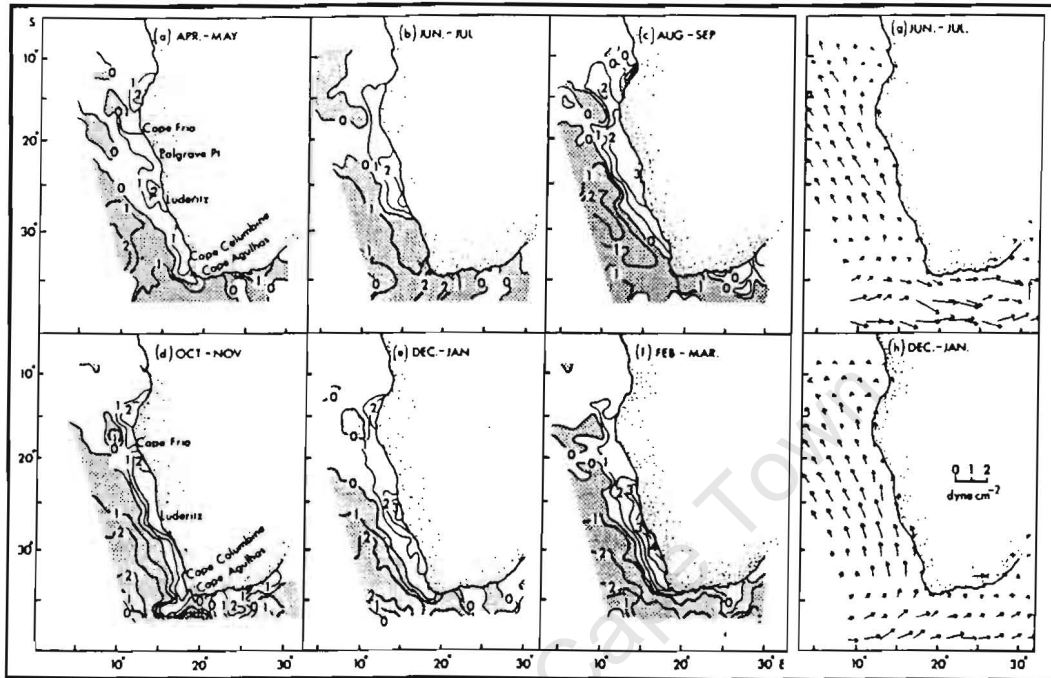


Figure 2.4 Wind-stress curl ($10^{-8} \text{ dyn.cm}^{-3}$) (a-f) and surface winds (g-h) for the Benguela region.
(modified from Bakun and Nelson, 1991)

HYDROGRAPHY

There are many physical oceanic processes in operation on a continental shelf. Those relevant to the Benguela Upwelling System are depicted in Figure 2.5. Some of these processes are internal in the sense that they are generated within the shelf boundaries and their effects are local or are trapped there. Examples of these are coastal-trapped waves, upwelling, frontal formation, thermocline formation and internal wave generation. Other processes are external in the sense of being externally applied forcing functions. The wind field and the tidal Kelvin wave are examples of external forcing functions. The wind field, bottom topography and coastal boundary, and stratification are the three factors which couple to produce the very complex nature of the shelf zone of the Benguela Upwelling Regime (Shannon *et al.*, 1990b).

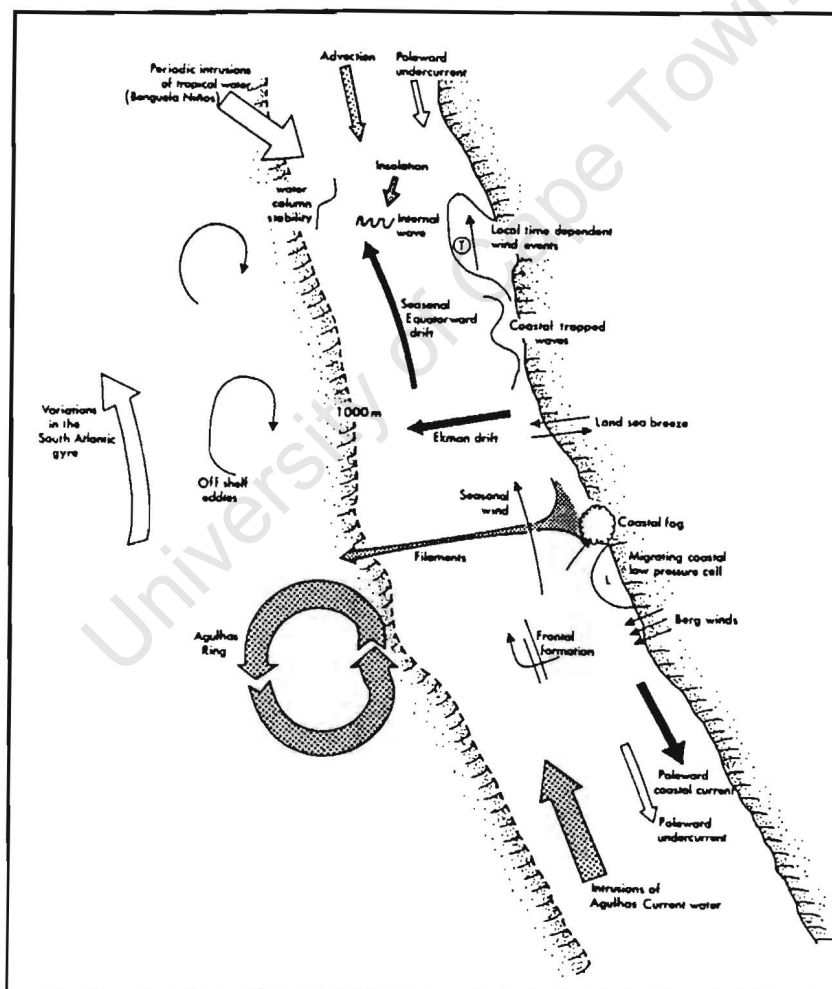


Figure 2.5 Schematic illustrating the complexity of the processes to be considered in the Benguela Upwelling System. (after Shannon *et al.*, 1990.)

Currents and Upwelling

As was mentioned above, the Benguela Current is the eastern boundary current of the South Atlantic subtropical gyre. Charts by Reid (1989) and Stramma and Peterson (1989) show the Benguela Current as beginning with a northward flow off of Cape Point (34.35°S) before bending toward the northwest to separate from the African coast at approximately 30°S. Primarily the South Atlantic Current feeds the Benguela current, however it can and does receive both Agulhas Current water (Duncombe Rae *et al.*, 1992) as well as Subantarctic Surface water, the latter coming from perturbations in the Subtropical Front (Shannon *et al.*, 1989).

As stated above, the semi permanent high-pressure system found over the subtropical South Atlantic, in conjunction with the low-pressure system that develops over southern Africa during the summer months, exert a strong atmospheric influence over the Benguela Current region. The prevailing winds in the region blow from the south and south-east. This in turn drives an offshore surface drift and in turn the upwelling of cold, nutrient-rich water onto the continental shelf (Peterson and Stramma, 1991). This upwelling exists with much spatial and temporal variability all along the southwest coast of Africa, from Cape Agulhas to Cape Frio. Lutjeharms and Meeuwis (1987) used uncalibrated infrared Meteosat SST images (1982-1985) to establish that the Benguela Upwelling System consisted of eight localised upwelling cells, namely the Cunene (17°S), Namibia (19°S), Walvis (22°S), Lüderitz (25°S), Namaqua (29°S), Columbine (32°S), Peninsula (34°S), and Agulhas (35°S) cells, the last three being more seasonal than the cells found in the central and northern parts of the system (Figure 2.6). The upwelling cells mentioned above tend to be located near regions of cyclonic wind-stress curl, and are in most cases in regions where there is a change in orientation of the coastline (Shannon and Nelson, 1996). Lutjeharms and Meeuwis (1987) found the Lüderitz cell to be the centre of the upwelling regime, as the cell was found to have the lowest average SST values, the greatest seaward extent, the highest frequency of occurrence, and the most intense wind-stress. Lutjeharms and Stockton (1987) concluded that the Benguela upwelling regime could be divided into two regions, namely the coastal upwelling regime and the filamentous mixing area. They found the coastal regime to be 150 to 200km wide on average, while the filamentous mixing area, which contained cold water filaments, plumes, and frontal eddies, was

found to extend up to 625km offshore. The widest filamentous regime was found to occur during the winter months, and the most extensive seaward penetrations were found off the major upwelling cells of Cape Frio, Lüderitz, and Hondeklip Bay.

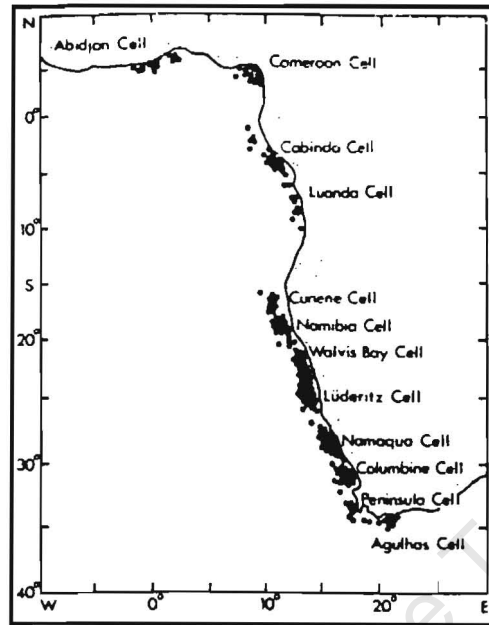


Figure 2.6 Distribution of the well-defined upwelling cells in the Southeast Atlantic Ocean. Each dot represents the centre of an upwelling event observed over a period of 156 weeks. (after Lutjeharms and Meeuwis, 1987.)

Bang and Andrews (1974) were the first to directly measure a shelf-edge jet to the west of Cape Town. Since then, this shelf-edge jet has been observed on a number of different occasions. In 1978, using a Neil Brown acoustic current meter, as well as advanced ship positioning, Nelson (1985) repeated observations of the shelf-edge jet. He found a near surface jet, with speeds of 40cm.s^{-1} , situated over the 300m isobath and in close proximity to the surface front. Nelson and Polito (1987) found the jet to be further inshore, with the jet being situated over the 230m isobath. A similar shelf-edge jet has been observed, on a number of different occasions, near Cape Columbine (Nelson and Hutchings, 1983). Bang (1973) found that trace amounts of Agulhas Current water, during the summer months, are almost always present in the region offshore of the southern Benguela Upwelling system, warm water which would presumably act to intensify the frontal jets.

Common to all but one of the World's eastern boundaries is the presence of an equatorward surface flow, and a poleward undercurrent, the exception being the Leeuwin Current situated off of Western Australia. Some of these poleward flows are due to region-specific local forcing, such as buoyancy-driven coastal currents. However, the extensive nature of eastern boundary poleward currents, which often flow against prevailing equatorward winds, suggests that a larger-scale, nonlocal forcing may be at work. These poleward flows tend to be located at the shelf break, with a width of between 20 and 100km and speeds of approximately 0.1m.s^{-1} (Hill *et al.*, 1998). Within the Benguela system, the poleward flow sometimes penetrates the shelf to occupy the lower layers (Nelson, 1989). The Sea Fisheries Research Institute has been deploying current meters on the shelf since 1978. These current meters have revealed that the subtidal currents on the shelf are dominated by the presence of coastally trapped waves with periods of 3-10 days. After removing low-frequency fluctuations from the signal, Nelson and Polito (1987) found that the predominant flow on the shelf is in a poleward direction. The mean speed of this flow is approximately $10\text{-}20\text{cm.s}^{-1}$, with higher values occurring in the region of steep topography near Cape Point.

The Leeuwin Current in the Indian Ocean off Western Australia differs from the other major eastern boundary currents as it contains a vigorous surface poleward flow of warm, low salinity water from the north-west shelf region of north-west Australia, against the prevailing equatorward winds, around Cape Leeuwin (34.5°S) in the south, and eastwards into the Great Australian Bight. An equatorward undercurrent exists at approximately 500m depth (Smith *et al.*, 1991). The explanation for the anomalous behaviour in this eastern boundary region seems to lie in the strength of the poleward steric height gradient in the eastern Indian Ocean. The strength of the steric height gradient along western Australia is related to the connection between the Pacific and Indian Oceans. This connection results in the steric height off north-west Australia being essentially the same as that of the western equatorial Pacific Ocean (Tomczak and Godfrey, 1994). This steric height is approximately 0.5m larger than that found off the coast of Peru, and drives a warm surface flow south towards the south-west tip of Australia where the anomalously warm water loses a great deal of heat to the atmosphere, overturns and sinks to intermediate depths. It is believed that this

convective overturning at mid-latitudes drives the equatorward Leeuwin Undercurrent.

Coastally Trapped Waves

The existence of coastal-trapped waves (CTW's) with periods of a few days to weeks has been demonstrated along various coastlines around the world. Cutchin and Smith (1973) described continental shelf waves and sea-level variation in the Oregon region, while Brink (1982) compared coastal-trapped wave theory with observations off Peru.

There have been relatively few investigations for CTW's off the coast of Southern Africa. Using tide gauge data collected along the west coast Southern Africa for the period of 1979-1983, Brundrit *et al.* (1984) gave evidence of progressive waves of sea-level change. These waves were found to have amplitudes of approximately 30 cm and typical phase speeds of 60km.h^{-1} (1440km.day^{-1}). The origin of such disturbances was speculated to be somewhere north of Walvis Bay, possibly as far as Luanda in Angola. Schumann and Brink (1990) used coastal sea level variations from 6 sites around South Africa to characterise CTW propagation. The authors were able to demonstrate the anti-clockwise progression of waves along the South African coastline. These waves had periods ranging from 4-14 days, speeds of approximately 3m.s^{-1} on the west coast (Port Nolloth to Cape Town), between 5 and 9m.s^{-1} on the south coast (George to Port Elizabeth), and between 2 and 5m.s^{-1} on the east coast (East London to Durban), and amplitudes of $>50\text{cm}$ on the south coast. In addition to this, the authors were also able to establish that the speeds of the wind systems moving along the coast fell into the same range as those of the CTW's, thus possibly leading to a resonance condition, and an explanation for the large CTW amplitudes. Jury *et al.* (1990b) correlated sea-level data with atmospheric coastal low-pressure cells. Their study suggested that a continuous wave-like propagation in sea level may occur in phase with a sharp reversal in longshore winds set up by transient coastal lows. These shelf waves were found to grow and decay in amplitude during their passage along the coast, reaching maximum amplitude near 22°E . Phase speeds evolved along the coast from 200km.day^{-1} south of Walvis Bay to over 1000km.day^{-1} north of Cape Town. Along the east coast these waves were found to decelerate to

approximately $100\text{km}\cdot\text{day}^{-1}$ at Durban, thus falling out of phase with their atmospheric counterparts.

Jury *et al.* (1990a) presented detailed ship data for a single event spanning the period 5-11 February 1981, so as to illustrate a sharp environmental pulse coinciding with the passage of a shelf wave. In the pre-wave period from the 5-7 February 1981, southerly winds of $20\text{m}\cdot\text{s}^{-1}$ induced a well-mixed shelf environment. Sea temperatures were $10\text{-}11^{\circ}\text{C}$ with no appreciable vertical gradient in the upper 50m. Salinities were 34.9-35.0psu, chlorophyll a concentrations were $10\text{mg}\cdot\text{m}^{-3}$ and concentrations of nutrients such as nitrate, silicate and phosphate were 10, 5 and $1\text{mmol}\cdot\text{m}^{-3}$ respectively in the upper 40m of the water column. Following the passage of the wave, the shelf environment became stratified, even though north-westerly winds blew at $8\text{m}\cdot\text{s}^{-1}$. Surface temperatures rose to $>13^{\circ}\text{C}$ while temperatures 40-50m deep decreased to $<10^{\circ}\text{C}$. Concentrations of chlorophyll-a and nitrate decreased within the surface waters, while levels of other nutrients in the lower layer (40-100m) increased, causing sharp vertical gradients. Vertical motions associated with the shelf wave caused the lower layer to decline from 40 to 60m and then to rise suddenly to 20m at the time of wave passage. Similarly, the character of the upper layer changed. Warmer water was advected from the northwest as winds and current reversed on 8 February. The wind reversal lasted for 24 hours, while surface currents remained poleward for over three days, according to a drogue track. Such environmental change associated with the regular passage of shelf waves implies that environmental pulsing of the Benguela Upwelling System is related to the passage of coastal lows (Jury *et al.*, 1990a).

Water Masses

The water that upwells to the sea surface along the continental shelf of southwest Africa consists of both South Atlantic Central Water, and South Indian Central Water (Shannon and Nelson, 1996). Central Water originates at the Subtropical Convergence and is formed through the sinking and partial mixing of Subantarctic Surface Water. Due to the fact that South Atlantic Central Water has very similar T/S characteristics to that of South Indian Central Water, it makes it very difficult to determine or to quantify leakage of South Indian Central Water into the southeast Atlantic Ocean

(Lutjeharms and Valentine, 1987). Nelson and Hutchings (1983) suggested that the inflow of South Indian Central Water is intermittent and that it is rapidly mixed into the intense upwelling zone. Above the South Atlantic Central Water lies the Atlantic Subtropical Surface Water. This water mass forms the upper layer of the southeast Atlantic that is not influenced by the coastal upwelling regime. The temperatures of this particular water mass tend to range from 15-23°C while the salinities are generally in the range of 35.4-36.0 psu (Lutjeharms and Valentine, 1987).

The Angolan Front

The Benguela upwelling system tends to be controlled through a number of important boundaries. These boundaries include the upper ocean surface, the open fluctuating boundary to the south, the offshore boundary, the bottom boundary, and lastly the northern boundary, which is represented by the Angola- Benguela front at approximately 16°S (Shillington, 1998).

The existence of a sharp thermal front at approximately 16°S, which separates the nutrient-poor warm Angolan Current water from the nutrient rich cold water of the Benguela system, was first documented by Hart and Currie (1960). Using a combination of data from the South African Centre for Oceanography, satellite thermal infra-red imagery, and airborne radiation thermometry surveys, Shannon *et al.* (1987) were able to highlight a number of large- and mesoscale features associated with the Angolan-Benguela frontal zone. On the large-scale, they were able to identify a largely zonal front which separated the waters of the Benguela and Angolan systems, with persistent tongue-like features to the north of the region. On the mesoscale, the authors found a complex system that, at times, consisted of two fronts between the area of 14 and 18°S. They concluded that the southernmost front was associated with the upwelling dynamics and that it could be viewed as an east-west expression of the main upwelling front. This then was seen to mark the northern boundary of the Benguela system proper. The second front appeared to be associated with the Angola tongue, and its position approximated to the southern most limit of penetration of tropical surface waters in the eastern Atlantic. Although the Angola-Benguela front is a permanent feature between 14 and 18°S, it does show some variability in time and

space. The main front appears to shift seasonally through about 2° of latitude, whereas a migration of 150km is possible on a time scale of days (Shannon *et al.*, 1987).

Factors that could determine the position of the front include coastline orientation, bathymetry, stratification, and wind-stress. Between 15 and 19°S, the coastline changes direction at both Cape Frio and Porto Alexandre, and this has important consequences for the coastal upwelling. Likewise, the width of the shelf and the steepness of the continental slope change dramatically between 15 and 20°S. In addition to these two factors, one finds a dramatic change in equatorward wind-stress within this region. North of 15°S, wind speeds are relatively low throughout the year, whereas near Cape Frio (18°S), winds are strongly upwelling-favourable throughout the year. Shannon *et al.* (1987) have thus suggested that a combination of the above four factors are likely to maintain the Angola-Benguela front.

Using satellite-derived weekly maps of SST for the period 1982-1985, Meeuwis and Lutjeharms (1990) found the northern frontal boundary to be located at approximately 14°S with an average water temperature of 23.6°C and the southern boundary at 16°S with a temperature of 20.6°C. Both boundaries were found to experience small fluctuations. These frontal movements occurred through a series of weak, pulse-like movements. Unlike Shannon *et al.* (1987), who only speculated as to the factors which may determine the position of the front, Meeuwis and Lutjeharms (1990) found the position of the front to be almost exclusively controlled by the opposing flows of the Angola Current and Benguela system. They found the Angola-Benguela front to be farthest south, widest, and most clearly developed during the summer months. This then correlates to when the south-flowing Angola Current, and the associated flow of warm water, is thought to be at a maximum, thus resulting in sharp temperature gradients over relatively small stretches of ocean. During late autumn and winter, the warm water retreats northwards, allowing cooler water to take its place. This results in weaker temperature gradients, causing the front to narrow, become less distinct and to lie much closer inshore.

Lass *et al.* (2000) investigated the structure and flow of the water column, in the vicinity of the Angola-Benguela front, down to 1200 dbar. The hydrographic pattern observed by Lass *et al.* (2000) in the surface mixed layer, agreed well with the

seasonal observations of Shannon *et al.* (1987) and Meeuwis and Lutjeharms (1990). The front, as determined by the 20°C isotherm, was located at 16°S at the sea surface, and was found to separate a 30-40m thick tongue of warm saline Angola Current water from cold and less saline Benguela water. The temperature gradient within the area of the front was approximately 5K per 2° of latitude, which, according to Shannon *et al.* (1987), is weaker than usual.

Rings and Filaments

In the mid-1970s, large-scale filaments of cold coastal water were observed in satellite infrared images extending 50-300km offshore in the California upwelling system (Bernstein *et al.*, 1977). Measurements of the kinematic characteristics of such filament features in the California Current system have shown that, under certain conditions, the flow configuration consists of a rapid offshore flow balanced by a slower, parallel return flow that converges at the upwelling front on the equatorward side of the filament (Flament *et al.*, 1985). Despite this meandering, some irreversible mixing of filament and ocean waters will occur. Thus filaments tend to play an important role in the exchange of water properties between the shelf and ocean (Huthnance, 1995). These filaments may extend down to a depth of over 100m and carry a volume flux in excess of $10^6 \text{m}^3 \cdot \text{s}^{-1}$ (Flament *et al.*, 1985).

Filaments are found in all major eastern boundary upwelling regimes, with the Benguela being no exception (van Foreest *et al.*, 1984). Lutjeharms and Stockton (1987) found the time scales of such filaments to be in the range of 5 days to 5 weeks. They were also able to identify the three centres associated with the main filaments; these occurred near 20, 26-27, and 30°S, the northern most position being over the Walvis Ridge. These locations are in substantial agreement with the key upwelling cells recognised by Lutjeharms and Meeuwis (1987). However, it should be noted that the Meteosat II data they used were not atmospherically corrected and that the spatial resolution at the latitude considered is of the order of 10km per pixel.

Many processes can produce filaments and several competing hypotheses have been advanced to explain their dynamical origin. Nelson (1985) found filaments forming as part of the Cape Peninsula upwelling cell to be a function of bottom topography,

while Jury (1985) found them to be a function of coastal orography. In another study, Taunton-Clark (1985) implicated bathymetric, orographic and meteorological effects in the formation of upwelling-tongues off the Cape Peninsula, Cape Columbine and Hondeklip Bay. Within the southern Benguela, Agulhas rings shed by the Agulhas Current Retroflexion have been known to interact with the Benguela frontal region (Duncombe Rae *et al.*, 1992). Lutjeharms *et al.*, (1991) made use of 10 years of thermal infra-red images for the southeast Atlantic Ocean to identify extreme upwelling filaments within the Benguela regime. They identified two mechanisms that could act to create these filaments of exceptional length, namely the interaction of the upwelling front with Agulhas rings and, secondly, the effect of intense berg winds. One such filament was identified on 5 July 1982 (Figure 2.7). This particular filament had a seaward extent of over 1300km. The surface expression for this particular filament was persistent for approximately ten days, 5-14 July 1982. These authors estimated the offshore volume flux of such a filament to be of the order of $1.5 \times 10^6 \text{ m}^3 \cdot \text{s}^{-1}$.

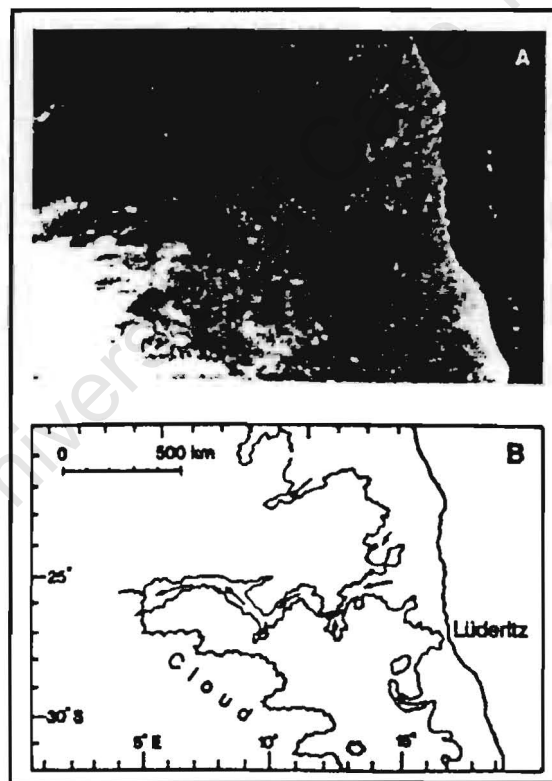


Figure 2.7 “A” is a thermal infrared image from the METEOSAT II satellite for 5 July 1982. The lighter hues are representative of colder water. An extreme, meandering upwelling filament is evident off Lüderitz. “B” is an interpretive sketch of the same scene with lines of latitude and longitude, as well as an approximate kilometre scale. (after Lutjeharms *et al.*, 1991.)

The first in situ measurements of the physical and biological properties of a filament of the Benguela upwelling system near Lüderitz were made in February 1988 by Shillington *et al.* (1990). Although the filament they surveyed had been in existence for approximately two weeks and was in a state of decay, they were able to identify four different water bodies in the region. The authors also speculated as to the nature of the mesoscale eddy that was located to the south of the filament. They suggested that this southern eddy might have been an Agulhas ring, which originated in the region of the Agulhas Retroflection. Russian ship measurements and satellite measurements of the northern Benguela frontal region in May 1985 (Kaz'min and Sutyurin, 1990) showed an eddy in a similar position to the one discussed by Shillington *et al.* (1990). The Russian results revealed that, even though the eddy was rotating anti-clockwise and was of a similar diameter to that of an Agulhas ring, approximately 140km, it was a relatively shallow feature with the 11°C isotherm situated at 250m. In comparison, the 11°C isotherm for Agulhas rings tends to be situated at approximately 550m (Duncombe Rae, 1991). This information suggests that the mesoscale may have been created through instabilities in the flow of the upwelling jet (Strub *et al.*, 1991), rather than being of Agulhas origin.

A number of investigations in regions other than that of the southeast Atlantic Ocean have revealed that localised maxima in productivity biomass are found at the borders of, or just outside of, upwelling fronts (e.g. Dengler, 1985). Simpson *et al.* (1979) concluded that, not only is the standing crop of phytoplankton several times greater at fronts, but in many cases it is considerably healthier. Lutjeharms and Walker (1985) have shown peaks in chlorophyll concentration at the borders of a warm Agulhas filament being advected into the South Atlantic Ocean. In the upwelling cells off southern Africa, it has furthermore been shown by Shannon *et al.* (1983) that phytoplankton blooms occur offshore and not in the centre of active upwelling cells. Lutjeharms and Stockton (1987) have suggested that an extensive field of upwelling filaments, with an abundance of long frontal lines and eddies, could be an area of considerable primary production in its own right. They further suggested that a substantial or even a major portion of the total primary production attributable to upwelling in this region could take place in the filamentous and not in the coastal upwelling regime.

VARIABILITY

From the above evidence it is clear that the Benguela Upwelling System is both a very dynamic and a very complex system. This upwelling system is forced at a variety of time and space scales, an example of which would be the wind events that have a variability that is dominated by the atmospheric synoptic scale of 3-10 days (Shillington, 1998). Intra- to interannual variability in the local marine environment manifests itself partly through changes in sea level, sea surface temperature, salinity, frontal structure, upwelling and nutrient supply, thermocline structure and mixed layer depth, currents, distribution and abundance of biota, community structure, dissolved oxygen concentrations and local weather (Shannon *et al.*, 1990b).

Aspects of Low Frequency Variability

There are few long-term data sets for the Southern African marine environment. The longest is for SST, dating from the early part of this century. Generally, however, the marine databases tend to be fragmented in both time and space. Consequently, much of the understanding of regional ocean variability has been deduced from case studies of extreme events (Shannon *et al.*, 1990b). Walker (1987), Taunton-Clark and Shannon (1988), and Taunton-Clark (1990), amongst others, have investigated variability in SST and surface winds this century in the Southeast Atlantic using data from voluntary observing ships (VOS). Using a principal component analysis, Walker (1987) revealed coherency of SST events in tropical and sub-tropical areas within the Southeast Atlantic from 14-30°S. Coastal events north of 30°S were found to correspond to offshore SST events but exhibited slightly altered timing and larger SST anomalies. A notable exception to the nearshore/offshore coherency was the nearshore cool event of 1981/83. This cool anomaly had its maximum expression near Lüderitz where negative anomalies of -2.5 to -3°C occurred. SST variability in the southern Benguela was found to differ markedly from that farther north as a result of mid-latitude atmospheric forcing, Agulhas current influences and an upwelling season out of phase with that of the northern Benguela. Taunton-Clark and Shannon (1988) analysed a long time series (1906-1984) of SST and pseudo wind-stress (τ), for six areas in the Southeast Atlantic (Figure 2.8). These areas encompassed both oceanic and coastal regimes. The principal findings of these authors were as follows. Both SST and wind-stress indices suggest that major changes have occurred this century in

the Southeast Atlantic. Periods prior to 1911 and after 1974 are characterised by significantly higher equatorward wind-stress, with lower values during the 1920's, 1930's and immediately after World War II. Comparable changes are evident in SST. An increasing trend in SST this century is evident, the post World War II era being typically 1°C warmer than earlier periods.

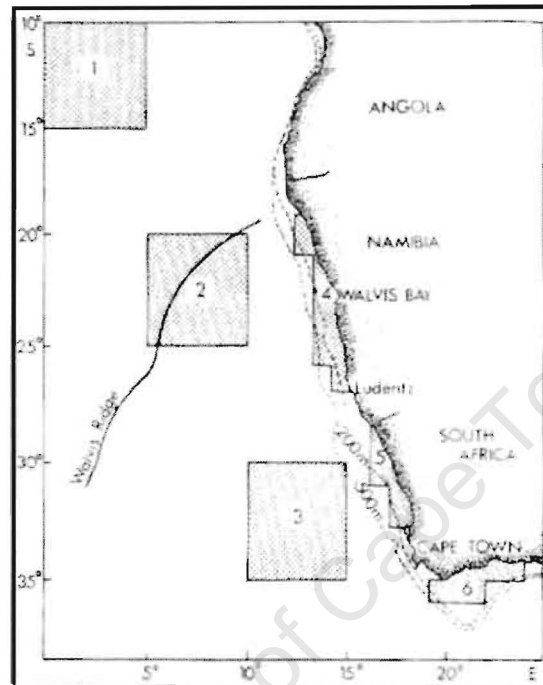


Figure 2.8 Locator map showing the six areas chosen by Taunton-Clark and Shannon (1988) for analysis of SST and wind variability (shaded and numbered).

Reason (2000) presented observations of multidecadal variability in SST, surface air temperatures and winds over the Southern Hemisphere for the period of 1900-1983. Using data derived from the GISST 2.2 data set, he concluded that the southeast Atlantic underwent a general cooling in SST from 1900-1941. This pattern was then reversed for the years of 1963-1983 where the southeast Atlantic was found to contain warm anomalies. Wind speed anomalies, derived from the COADS data set, show mainly negative anomalies for the Southern Hemisphere (including the southeast Atlantic) oceans for the period of 1921-1941 while from 1963-1983 they tended to be the reverse, with the southeast Atlantic following this trend. The patterns for the other two periods are a bit more complex. From 1900-1920, it appears that the southern Benguela region contained positive wind speed anomalies, while the northern Benguela was influenced by negative wind speed anomalies. For the period of 1942-

1962 it broadly appears that the reverse was true. Thus these findings are similar to those of Taunton-Clark and Shannon (1988), who were also able to identify similar variability in wind speed, for the southeast Atlantic, over the last century.

One of the most prominent examples of climatic variability is that associated with the El Niño - Southern Oscillation (ENSO). Most scientific research on ENSO has focused on the Pacific Basin where the core of the physical processes underlying the phenomenon occur (Reason *et al.*, 2000). Reason *et al.* (2000) focussed on ENSO signals in the Indian Ocean basin; however, their global plots of SST, MSLP and wind anomalies composited for strong El Niño and La Nina events during the 1877-1993 period suggest that there may also be a coherent ENSO signal in the South Atlantic Ocean. The data analysed by Reason *et al.* (2000) was at a relatively coarse resolution, which makes it difficult to assess the potential ENSO signal in the Benguela upwelling system. Nevertheless, some of the largest South Atlantic anomalies shown by Reason *et al.* (2000) for the austral summers during strong events tend to be found in the southeast of this basin.

Both cool and warm events are apparent from the SST record. Benguela Niños (Shannon *et al.*, 1986) are the important interannual warm events in the northern Benguela region, and are superimposed on seasonal changes in the structure and location of the Angola-Benguela front. To the south, the only well documented Agulhas Current intrusion into the Benguela region was one which peaked around 1986 (Shannon *et al.*, 1990a). This intrusion resulted in 1986 being the warmest year in the past century in the Southeast Atlantic. The cool events, which are possibly as important as the warm events, have received little attention. Taunton-Clark (1990) suggested that the cooling in 1928 and 1955 might have been associated with changes in the Subtropical Convergence and in the easterly winds.

Benguela Niños

Shannon *et al.* (1986) termed the warm events in the south-east Atlantic “Benguela Niños”, because of their apparent similarities with the warming in the Peruvian coastal upwelling and anomalous coastal rainfall during El Niño events. These warm events appear to arise as warm water of equatorial Atlantic origin propagates

poleward as a Kelvin wave along the south-west coast of Africa, at times reaching as far south as 25°S. It is thought that the warm events are forced by equatorial wind variability since they follow ENSO-like warming in the equatorial Atlantic associated with a sudden relaxation of the tradewinds near Brazil. This relaxation leads to the generation of downwelling equatorial Kelvin waves as well as a strengthening of the South Equatorial Counter Current, thereby producing a depression of the thermocline along the equator (Servain *et al.*, 1982). Warm waters then accumulate in the eastern equatorial south Atlantic and downwelling coastal Kelvin waves propagate the anomalies southward along the south-west African coast (Carton and Huang, 1994).

Shannon *et al.* (1986) made use of temperature, salinity, sea level, rainfall and NOAA satellite data to investigate three major warm events in the northern Benguela. The major events investigated occurred during the years of 1934, 1963 and 1984. These authors found these three events to share a number of common features. In each case, the warm water persisted for 6 months or longer, with anomalous conditions in 1963 spanning a full year. The 1963 and 1984 anomalies were associated with above-average sea level, with positive anomalies persisting for at least 12 months. During both 1963 and 1984, equatorward wind-stress over the northern Benguela was stronger than normal, yet upwelling was suppressed by the southward progression of warm, highly saline water from the Angolan region. The warm water intrusions were accompanied by abnormal rainfall and flooding of northern Namib Desert rivers. In all warm years, the origin of the warm water appeared to have been from the western equatorial Atlantic. During 1963 and 1984, Angolan water penetrated at least as far south as 24-25°S, while they found a strong indication from the surface drift in 1934 and the SST distribution in 1984 that the central part of the Benguela (24-30°S) may also have been affected.

Gammelsrød *et al.* (1998) reported the upper ocean temperatures in the Angolan-Namibian coastal waters to be anomalously high during March of 1995, with positive temperature anomalies of up to 8°C. Using both direct observations as well as satellite derived SST, the authors were able to determine that the unusually warm water mass stretched from Cabinda (5°S) in the north, to Lüderitz (27°S) in the south, and had a seaward extension of over 300km from the coast. Surface drogue data suggested a

maximum southward extension of this warm water intrusion by 3 March 1995. This warm event was associated with high observed mortalities in sardine *Sardinops sagax*, horse mackerel *Trachurus trachurus capensis* and kob *Argyrosomus inodorus* off the coast (Gammelsrød *et al.*, 1998).

The latest Benguela Niño event occurred in 2001. Rouault *et al.* (2002) discussed this warm event, as well as other events that occurred in 1984, 1986, and 1995, in terms of their links to above average rainfall over Southern Africa. Rouault *et al.* (2002) suggest that the occurrence of these warm events in the tropical south-east Atlantic during late austral summer, at the time of maximum annual SST, can amplify local atmospheric stability, evaporation and rainfall. These events were found to be associated with above-average rainfall over the tropical south-east Atlantic and western Angola/Namibia. If the large-scale circulation was favourable, precipitation anomalies were found to extend farther into southern Africa.

Agulhas Intrusions

During late 1985 and in 1986, Shannon *et al.* (1990a) documented a major perturbation in the Agulhas retroflection. These authors used a combination of SST, climatological, and satellite SST data to describe the evolution of the abnormal intrusion of Agulhas Current water along the west coast of southern Africa. This intrusion appears to have started as a small offshoot of the Agulhas Current, flowing around the edge of the Agulhas Bank, and developed into a major change in the retroflection. The authors suggest that by 10 January 1986 this perturbation had developed into a ring centred on 36°S:19°E with a diameter of about 160km. Over the next six months there was intermittent leakage of Agulhas water around Cape Point, and by early July there were signs of a large body of Agulhas water to the south of Cape Town. September saw the displacement of the northern and eastern thermal boundaries slightly northwards and westwards respectively. During the next three months the warm water advanced farther north and west, and progressively lost its surface expression. By the second half of 1987, the Southeast Atlantic had returned to a more normal situation with the retroflection once again prevailing in a more familiar position of 20°E (Lutjeharms and Van Ballegooyen, 1988). The authors were unsure as to how far north the Agulhas intrusion moved. However, Agenbag and Shannon

(1987) suggested that it reached as far north as Lüderitz (26.5°S). Shannon *et al.* (1990a) examined possible causes of the 1986 intrusion and have provided evidence that suggests that it was linked to changes in winds in the south-west Indian and south-east Atlantic Oceans. The modulation of wind-stress within these two regions may have resulted in lower volume transport of the Agulhas Current and a southward displacement of the zero line of wind-stress curl. The observed changes in the large-scale atmospheric forcing, and in particular the sudden relaxation of the Indian Ocean trades and in westerlies south of Africa after August 1985, are consistent with model predictions of increased leakage of Agulhas water into the Atlantic (Boudra and Chassignet, 1988).

As stated above, this perturbation resulted in 1986 being the warmest year for the previous century in terms of southeast Atlantic SST. However, it also resulted in a substantial equatorward flow of Subantarctic Surface Water in the form of cold filaments. One of these filaments extended into the Benguela region as far north as 33°S during January 1987. This feature persisted for at least two months during the summer of 1986/87 and appeared to temporarily restrict the flow of surface Agulhas water into the south-east Atlantic (Shannon *et al.*, 1990b). Perturbations of this nature could be important climatically and for the transfer of water and biota between the South Atlantic and Southern oceans.

Wind-Stress

From the above, it is clear that wind-stress variability is a major driving force behind ocean variability over a wide spectrum of time and space scales. Changes in longshore wind are important for barotropic processes on the shelf, surface drift and Ekman transport, stratification and for upwelling and frontal dynamics in general. Changes in the large-scale wind field have likewise been linked to several large-scale advective processes which have been identified as important contributors to the variability in the oceans around Southern Africa, namely Benguela Niños (Shannon *et al.*, 1986), Agulhas Current fluctuations, intrusions of Agulhas water from the retroflexion into the South Atlantic (Shannon *et al.*, 1990a) and the El Niño Southern Oscillation (Reason *et al.*, 2000). However, Shannon *et al.* (1990b) cautioned against taking an over-simplified view of wind and ocean variability. In reality, one finds a complex

system of feedbacks between the ocean and the atmosphere. In addition to this, it is likely that changes in the ocean heat and salt budgets might be just as important as the wind in controlling ocean variability in both the regional and southern hemisphere contexts (Shannon *et al.*, 1990b).

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CHAPTER 3

DATA

The data set analysed here was derived from the SeaWinds scatterometer, which was launched on 19 June 1999 onboard the QuikSCAT satellite. The geophysical data record began on 15 July 1999. The 16-month data record analysed here extends from 01 August 1999 through to 29 November 2000.

SEAWINDS ON QUIKSCAT

As the largest source of momentum for the upper ocean, winds affect the full range of ocean movement, from individual surface waves to complete current systems. Winds over the ocean modulate air-sea exchanges of heat, moisture, gases, and particulates. This modulation regulates the interaction between the atmosphere and the ocean, which establishes and maintains both global and regional climates.

In the past, weather data could be acquired over land, with our knowledge of ocean surface winds coming from infrequent, and sometimes inaccurate, reports from ships and buoys. Scatterometry, which had its origin in early radar used in World War II, has grown over the last three decades from limited experiments on the Skylab missions (1973 & 1974), to modern scatterometers flown on ERS-2 and QuikSCAT, and soon on METOP and ADEOS-2. Scatterometers are now providing global wind fields with unparalleled precision and temporal resolution, with the QuikSCAT data being routinely assimilated into the NCEP and ECMWF meteorological models since early 2002, contributing considerably to their prediction skill (Flament, 2000).

However, much remains to be done to understand physical aspects of the radar signal and its relationship to the wind. Factors like sea state, fetch, sensor saturation and bias, interference with rain and other atmospheric processes, need to be further explored to exploit the potential synergy of concurrent heterogeneous satellite sensors (Flament, 2000).

NASA's Quick Scatterometer (QuikSCAT) was lofted into space on 19 June 1999. The SeaWinds on QuikSCAT mission was a "quick recovery" mission to fill the gap created by the loss of data from the NASA Scatterometer (NSCAT), when the satellite it was flying on, ADEOS-1, lost power in June 1997. The QuikSCAT satellite is in a sun-synchronous, 803km, circular orbit with a local equator crossing time at the ascending node of 6:00 A.M. $\pm$ 30 minutes and an orbital period of 101 minutes ($14.25 \text{ orbits.day}^{-1}$). The repeat time for the QuikSCAT satellite is 4 days or 57 orbits (Perry, 2001).

The SeaWinds Instrument

The SeaWinds instrument on the QuikSCAT satellite is an active microwave radar designed to measure the electromagnetic backscatter from the wind roughened ocean surface. The instrument is a conically scanning pencil-beam scatterometer, which in comparison with the NSCAT fan-beam scatterometer has the following advantages: higher signal-to-noise ratio, smaller in size, and superior coverage (Portabella and Stoffelen, 2000). SeaWinds uses a rotating 1-meter dish antenna with two spot beams, an H-pol beam and a V-pol beam at incidence angles of 46° and 54° respectively, that sweep in a circular pattern. The antenna radiates microwave pulses at a frequency of 13.4 GHz (Ku-Band) across a 1800 km-wide swath centred on the spacecraft's nadir subtrack, making approximately 1.1 million 25-km ocean surface wind vector measurements and covering 90% of the Earth's surface every day. The SeaWinds swath is divided into equidistant across-track wind vector cells (WVC) numbered from left to right when looking along the satellite's propagation direction. The nominal WVC size is 25 km x 25 km, and all backscatter measurements centred in a WVC are used to derive the WVC wind solutions. Due to the conical scanning, a WVC is generally viewed when looking forward (fore) and a second time when looking backward (aft). As such, up to four measurement classes emerge: H-pol fore, H-pol aft, V-pol fore, and V-pol aft, in each WVC. Due to the smaller swath (1400 km) viewed in H-pol at 46° degrees incidence, the outer swath WVCs have only V-pol fore and aft backscatter measurements (Portabella and Stoffelen, 2001).

The QuikSCAT system is designed to measure winds between $3 \text{ and } 30 \text{ m.s}^{-1}$ with an accuracy better than (the greater of) 2 m.s^{-1} or 10% in speed and 20° in direction with

a spatial resolution of 25 km. QuikSCAT generates an internal calibration pulse and associated load pulse every half-scan of the antenna. In ground processing, the load pulses are averaged over a 20-minute window, and the cal pulses over a 10-pulse (approximately 18-second) window, to provide the current instrument gain calibration needed to convert telemetry data numbers into power measurements for the σ_0 calculation (Perry, 2001).

DATA PROCESSING

Level 2B Processing

SeaWinds on QuikSCAT science processing proceeds through a well-defined series of level conversion stages, producing more refined products at each stage. The products are created in the following order: Level 0 (Science Telemetry Processing), Level 1A (Engineering Unit Converted Telemetry), Level 1B data (Time-Ordered Earth-Located Sigma0s), Level 2A Data (Surface Flagged Sigma-0s and Attenuations in 25km Swath Grid), Level 2B Data (Ocean Wind Vectors in 25km Swath Grid), and Level 3 (Daily, Gridded Ocean Wind Vectors) (Perry, 2000).

Sigma-0 Grouping and Surface Flags

The sigma-0 grouping algorithm prepares the SeaWinds sigma-0 data for wind retrieval processing. The data contained in the Level 1B product are grouped by geographic location into wind vector cells (WVC). The grouped sigma-0 data are saved in WVC rows in the L2A product. After the sigma-0 data have been assigned to a WVC, each cell is checked for land and sea ice. The land map used is the same CIA land-sea map used for NSCAT. The sea ice mask is generated from weekly National Ice Centre ice edge data. Both the land and ice flagging algorithms check the centre of the sigma-0 cell against the land-sea map and the ice mask (Perry, 2000).

The Wind Vector Cell Preparation algorithm operates on a row of WVC values, passed from the Grouping algorithm, one WVC at a time. It then must determine if there are sufficient data of sufficient quality to perform wind retrieval. This algorithm checks each WVC to determine the data counts (total and by beam), quality flags, and surface flags. It then computes the centroid of the sigma-0 locations to give a WVC

location (latitude/ longitude; the binning grid is essentially "thrown away" at this point), and passes the "good" data to the Wind Retrieval algorithm. Upon return from wind retrieval, the ambiguous wind vector data are placed in the Level 2B output buffer (Perry, 2000).

Wind Retrieval

An accurate model function is essential to deriving ocean wind vectors from scatterometer measurements. Varieties of Ku-band models exist, some of which are derived from scattering theory, and some which are empirically derived. The tabular form of the model function, using a table of real-space (non-dB) sigma-0 values, is used for SeaWinds, with modifications for the incidence angle range and the resolution of the table in azimuth and incidence angle. The model function used for SeaWinds is the QSCAT-1 model (Perry, 2000).

Ambiguity Removal

The SeaWinds ambiguity removal algorithm uses a modified median filter technique to select a unique wind vector out of a set of ambiguous wind vectors at each wind vector cell. The algorithm is a direct adaptation of the ambiguity removal algorithm used for NSCAT (Perry, 2000).

NWP Initialization of Ambiguity Removal

The baseline ambiguity removal algorithm for SeaWinds incorporates the Numerical Weather Product (NWP) initialisation technique used for NSCAT-1 and NSCAT-2 processing. In this "nudging" technique, the median filter algorithm is initialised with either the first or the second ranked wind vector solution, whichever is closer to the direction of the NWP analysis field. The median filter algorithm then proceeds as described above to generate the final wind vector selections (Perry, 2000).

Direction Interval Retrieval with Threshold Nudging (DIRTH) Algorithms

At far swath, ambiguity removal skill is degraded due to the absence of inner beam measurements, limited azimuth diversity, and boundary effects. Near nadir, due to non-optimal measurement geometry (fore and aft looking measurement azimuths approximately 180° apart), there is a marked decrease in directional accuracy even when ambiguity removal works correctly. Two algorithms were developed, direction

interval retrieval (DIR) to address the nadir performance issue, and threshold nudging (TN) to improve ambiguity removal at far swath. The two algorithms work independently and need not be used together. However, both were used to obtain the DIRTH solutions, `wind_speed_selection` and `wind_dir_selection`, in the Level 2B product (Perry, 2000).

Level 3 (Daily, Gridded Ocean Wind Vectors)

The Level 3 data were obtained from the Direction Interval Retrieval with Threshold Nudging (DIRTH) wind vector solutions contained in the QuikSCAT Level 2B data and are provided on an approximately $0.25^\circ \times 0.25^\circ$ global grid. Separate maps are provided for both the ascending pass (6am LST equator crossing) and descending pass (6pm LST equator crossing). By maintaining the data at nearly the original Level 2B sampling resolution and separating the ascending and descending passes, very little overlap occurs in one day. However, when overlap between subsequent swaths does occur, the values are over-written, not averaged. Therefore, a SeaWinds on QuikSCAT Level 3 file contains only the latest measurement for each day (Perry, 2000).

The SeaWinds on QuikSCAT Level 3 data set consists of gridded values of scalar wind speed, meridional and zonal components of wind velocity, wind speed squared and time given in fraction of a day. The wind retrievals are calibrated to the so-called neutral-stability wind at a height of 10m above the sea surface, i.e., the wind that would exist if the atmospheric boundary layer were neutrally stable. The vector wind-stress τ and 10-m neutral stability vector wind V_{10} are related using;

$$\underline{\tau} = \rho C_d \underline{V}_{10} |\underline{V}_{10}|$$

where ρ is the air density and C_d is the neutral stability drag coefficient (Large and Pond 1981). Rain probability, determined using the Multidimensional Histogram (MUDH) Rain Flagging technique, is also included as an indicator of wind values that may have degraded accuracy due to the presence of rain.

The u and v wind-stress components and the derivative wind fields, namely curl and divergence, that are investigated here, have been sub sampled from this QuikSCAT Level 3 data set, to a spatial domain of 10°S to 35°S and 5°E to 20°E. The data have also been smoothed temporally, to 2-day composites, and spatially, to a one-half degree resolution.

There are both advantages and disadvantages associated with such a temporal and spatial smoothing of the data. The major advantage of this technique is the significant reduction of sampling errors within the data set, thus one can obtain a more superior coverage of the area in question. An obvious disadvantage is that the smoothing will result in the loss of some detail within the wind field, when compared with the original Level 3 Daily Gridded Ocean Wind Vectors.

Calibration and Validation

A comprehensive calibration and validation study is underway by the QuikSCAT Science Working Team. Chelton *et al.* (2001), found a systematic mean difference of 0.74 m.s^{-1} between 3 day averages of QuikSCAT and TAO measurements of 10m neutral-stability winds at the locations situated in the equatorial eastern Pacific. These authors concluded that this might be an indication that the wind speeds from the QuikSCAT pre-launch algorithm are biased slightly low. This pre-launch bias has been removed in the version of the QuikSCAT data analysed in this study. Freilich *et al.* (2000; in Chelton *et al.* 2001) have compared the QuikSCAT wind directions with over 12000 buoy observations of winds collected by the National Data Buoy Centre and concluded that the standard deviation of the directional difference is 27° . The directional accuracy is better characterized in terms of the random component error, which Freilich *et al.* (2000; in Chelton *et al.* 2001) find to be about 0.7 m.s^{-1} with no significant dependence on the wind speed. The directional uncertainty of the QuikSCAT winds thus decreases rapidly with increasing wind speed (see also Freilich and Dunbar, 1999).

CHAPTER 4

ARTIFICIAL NEURAL NETWORKS AND WAVELET ANALYSIS

THE KOHONEN SELF-ORGANISING MAP

A number of techniques can be used to identify and characterise wind-stress patterns. One such technique is through the use of artificial neural networks (ANNs). ANNs attempt to mimic the computational power of the mammalian brain by massively interconnecting very simple computational “neurons”. Typically, the human brain consists of 10^{11} neurons, each one with an average of 10^3 - 10^4 connections. It is believed that the immense computing power of the brain is the result of the parallel and distributed computing performed by these neurons. The result of following the design philosophy of massively interconnecting simple units has provided models that have proved to be successful in a number of applications including text to speech recognition, protein structure analysis, hand writing recognition, and image and signal processing (Misra, 1997).

This study utilises one type of unsupervised ANN particularly adept at pattern recognition and classification, the self-organising map (SOM) (Kohonen *et al.*, 1995). In contrast to clustering algorithms, the basic SOM methodology is not primarily concerned with the grouping of data or identification of clusters, although data clustering will result from a SOM analysis. The initial step in the SOM routine is to define a random distribution of nodes within the data-space. Each node has an associated reference vector equal in dimension to the input data. As data are presented to the SOM the similarity between the data and each of the node reference vectors is calculated. The reference vector of the “winning” node is then modified, via the user defined *learning rate*, to reduce the difference between itself and the input vector. The input vector is thus simply used to adjust the location of the SOM node in the data-space (Hewitson & Crane, 2002). The major difference between most cluster algorithms and the SOM is that for a SOM, not only is the “winning” node updated but that all of the surrounding nodes are also adjusted in proportion to their distance from that particular node. This iterative process continues until the changes in the

node location are negligible. The net result is that the SOM will concentrate nodes in regions of the data-space which have a higher data density. The reference vectors are iteratively adjusted such that they span the data-space with each node representing a position analogous to the median of the nearby data samples. This procedure effectively identifies “arch-type” points in the data-space that span the continuum of the data (Hewitson & Crane, 2002).

SOMs have a number of advantages over traditional statistical methods. First, they are able to solve non-linear problems of almost infinite complexity, and secondly they are more robust in handling noisy and missing data than traditional methods (Dayhoff, 1990). For identifying groups in a data set, the primary advantage of SOMs is that the relationship among the identified patterns can be visualised in the same form as the original data. This is an advantage over other multivariate techniques such as cluster analysis, MDS and EOF analysis, which output various abstractions of the original data. As the output patterns duplicate the input format, SOMs are more easily interpreted than the outputs from conventional multivariate techniques.

A weakness of the SOM, when compared with more conventional multivariate techniques, is that there is no robust statistical theory underpinning it, as there is for parametric statistics. Thus, there is no significance testing of the identified patterns, whereas EOF analysis provides statistical testing of the underlying principal components.

As stated above, SOMs are in wide spread use across a number of disciplines, but have had little exposure to date in both climatological and oceanographic literature. SOMs were introduced to the Physical Geography community as part of a broader discussion on neural nets by Hewitson and Crane (1994), but only the ANN component has become widely adopted. Hewitson and Crane (2002) have related sea level pressure fields, over the north-eastern USA, to station precipitation in the centre of the domain. This was done with the aim of demonstrating the range of analytical approaches available to investigators when using a SOM analysis. Main (1997) examined the seasonality of circulation in southern Africa, using both observed data, obtained from the Australian Southern Hemisphere data set, and simulated sea-level pressure data derived from the GENESIS version 1.02 General Circulation Model. In

terms of oceanography, Ainsworth (1999) and Ainsworth and Jones (1999) have used SOMs to improve chlorophyll estimates from satellite by better pixel classification, while Silulwane *et al.* (2001) used SOMs to identify characteristic chlorophyll profiles in the ocean.

In order to extract characteristic wind-stress patterns over the Benguela Upwelling Regime for the period of 1 August 1999 – 29 November 2000, two SOMs were performed. These SOMs were performed using the SOM_Pak software Version 3.1 (Kohonen *et al.*, 1995), which is freely available from the Neural Network Research Centre at the Helsinki University of Technology¹.

To apply the SOM to satellite images, some pre- and post-processing of the data are necessary. Each wind vector could be described as either u and v -components or speed and direction. Either is suitable and will give very similar results, so u and v -components were chosen for simplicity. Input data consisted of the combined u and v wind-stress components (columns) by the number of days (rows).

For the first study, each scatterometer image was transformed into a single row vector by concatenating each row in the image (Figure 4.1). All u -components were placed in the first half of the rows followed by all v -components. The land mask was removed so that input data consisted only of sea pixels. The final input matrix consisted of 1970 columns (985 grid points x 2 components) x 244 rows (number of 2 day periods). After the SOM was performed, scatterometer images were recreated from the node weights by recombining the u and v component of each grid point and reconstructing the two-dimensional shape of the image. The land mask was then reinserted.

A perceived difficulty in using the SOM is that there are no quantitative methods by which a researcher can choose the number of dimensions represented by the SOM array. If the data have a continuum of underlying patterns, the number of patterns chosen is dependent upon the level of detail required in the analysis. A large SOM array identifies a large number of patterns and in so doing reveals more detailed

¹http://www.cis.hut.fi/research/som_lvq_pak.shtml

structure within the data, whereas a small SOM array identifies fewer, more generalised patterns. Thus, the decision on the number of patterns in a SOM analysis is similar to that of the level of similarity required to identify groups in a cluster analysis. As indicated above, the Benguela Upwelling Regime is an extremely dynamic and variable environment. SOM arrays with dimensions of 4x3, 5x3, 6x3, 7x3, and 5x4 were investigated (not shown here). However, a 6x4 SOM array (Figure 5.4), with a rectangular topology, due to the nature of the data set as well as its relatively low mapping error, was found to be most representative of the data. These 24 synoptic patterns represent most of the underlying wind-stress patterns, with any anomalies being clearly identifiable through a relatively higher mapping error. The SOM was randomly initialised. The initial training consisted of 1000 iterations, a learning rate (α) of 0.2, and update radius of 5, with a bubble neighbourhood function (Kohonen *et al.*, 1995). The second training consisted of 50000 iterations a learning rate (α) of 0.02, an update radius of 0, and again a bubble neighbourhood function.

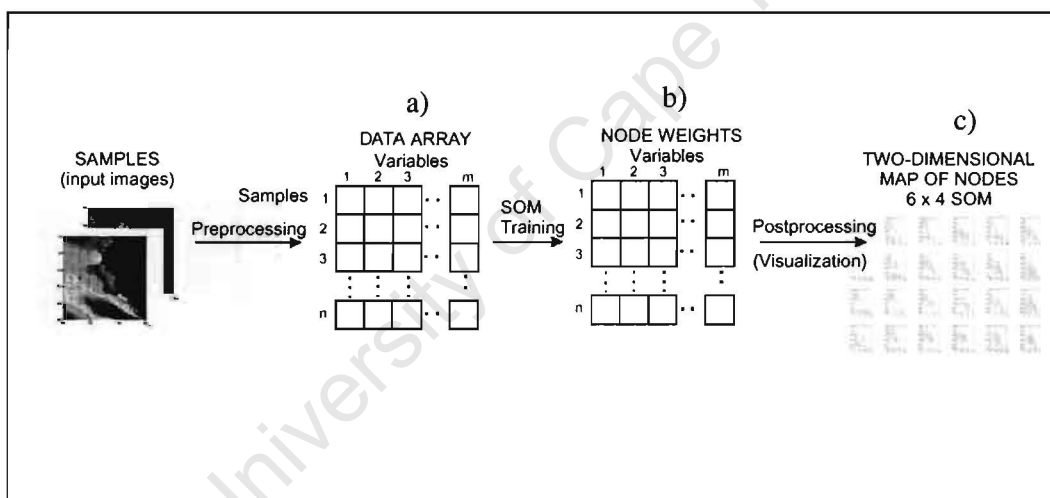


Figure 4.1 A schematic showing the implementation of the self-organizing map: (a) pre-processing (b) training, and (c) post-processing.

The data set for the second SOM analysis consisted of the u and v -components along a transect 200km offshore from 10-35°S, for every 0.5° of latitude (See also Figures 5.16 and 5.17). As above, all u -components were placed in the first half of the rows followed by all v -components. The final input matrix consisted of 488 columns (244 two day composites x 2 components) x 51 rows (number of sampled latitudes). The

output of the second SOM analysis grouped the data into six time series, representing six distinct wind regions (Figure 5.8), with each time series consisting of both a u and a v -component. The v -component time series was extracted for each of the six regions and the periodicity of each of these time series was investigated using wavelet analysis.

For this second analysis, a 3x2 SOM array, with a rectangular topology, was selected. After the data were randomly initialised, an initial training of 1000 iterations (learning rate (α) = 0.2, update radius = 5, bubble neighbourhood function) was conducted. The second training consisted of 50000 iterations (learning rate (α) = 0.02, update radius = 0, bubble neighbourhood function).

For a more detailed account of the principal of the SOM with regards to topology, initialisation, training, learning rate, and convergence, as well as the basic SOM algorithm, the reader is referred to Kohonen *et al.* (1995).

WAVELET ANALYSIS

Wavelet analysis is an analysis tool well suited to the study of multi-scale, non-stationary processes occurring over finite spatial and temporal domains (Lau and Weng, 1995). By decomposing a time series into time-frequency space, one is able to determine both the dominant modes of variability and how these modes vary in time (Torrence and Compo, 1998). The wavelet transform has been used for various studies in meteorology and oceanography, including turbulence, surface gravity waves, low-level cold fronts, and the variations in ENSO indicators within the Southern Oscillation Index and Niño 3.4 SST.

The continuous wavelet transform (CWT) allows for the decomposition of a signal $x(t)$ into fundamental contributions localised in time and frequency. This is done through the use of wavelets. These wavelets are obtained from a single function ψ by translations and dilations

$$\psi_{b,a}(t) = \frac{1}{a} \psi\left(\frac{t-b}{a}\right), \quad (1)$$

where $a > 0$ is the dilatation parameter and b is the time translation parameter. The CWT expands the time series $x(t)$ into a two-dimensional parameter space (b, a) and yields a measure of relative amplitude of local activity (over an interval proportional to a) at scale a and time b (Mélise *et al.*, 2001). The choice of wavelet ψ depends on the signal to be analysed. For this study, the Morlet wavelet (Figure 4.2) was chosen. This wavelet is one of the most widely used wavelets in geophysics (Lau and Weng, 1995). The Morlet wavelet is a complex cosine wave modulated by a Gaussian function

$$\psi(t) = \pi^{1/4} e^{-t^2/2} e^{i\omega_0 t}, \quad (2)$$

with $i = (-1)^{1/2}$ and where $\omega_0 = \pi (2/\ln 2)^{1/2}$ (Daubechies, 1992) is chosen to be large enough to ensure that $\psi(t)$ satisfies the admissibility condition which practically is equivalent to

$$\int_{-\infty}^{\infty} \psi(t) dt = 0. \quad (3)$$

As the Morlet wavelet is complex, the wavelet transform coefficient $W(b, a)$ is also complex and may be expressed in terms of real and imaginary parts, modulus and phase.

By inverting the scale of the Morlet wavelet, the CWT becomes a time-frequency analysis where the dilatation parameter a corresponds to the period and the translation parameter b corresponds to the time. In (1), the normalisation $1/a$ is used instead of the usual $1/(a)^{1/2}$. With this normalisation, the components of the CWT may be directly compared to each other and the Morlet wavelet can be interpreted as a bandpass linear filter of weight $1/a$ centred around $\omega = \omega_0/a$. This filter allows the extraction of the different local components of the signal such as its local value, amplitude and phase, for each point of the (b, a) time-frequency space (Mélise *et al.*, 2001).

Wavelet analysis has several attractive advantages over the more traditional power spectral analysis, especially when dealing with time series with time-varying amplitudes. In contrast to the power spectrum, which is a plot of amplitude squared with frequency averaged over the entire time series for each frequency component, the wavelet transform provides a localised, “instantaneous” estimate of the amplitude and phase for each frequency component of the series. This gives wavelet analysis an advantage in analysis of nonstationary data in which the amplitude and phase of the harmonic components may change rapidly in time. While the more traditional power spectrum of the non-stationary time series would smear out detailed information on the changing features, the wavelet transform keeps track of the evolution of the signal characteristics throughout the time series (Emery and Thomson, 1998).

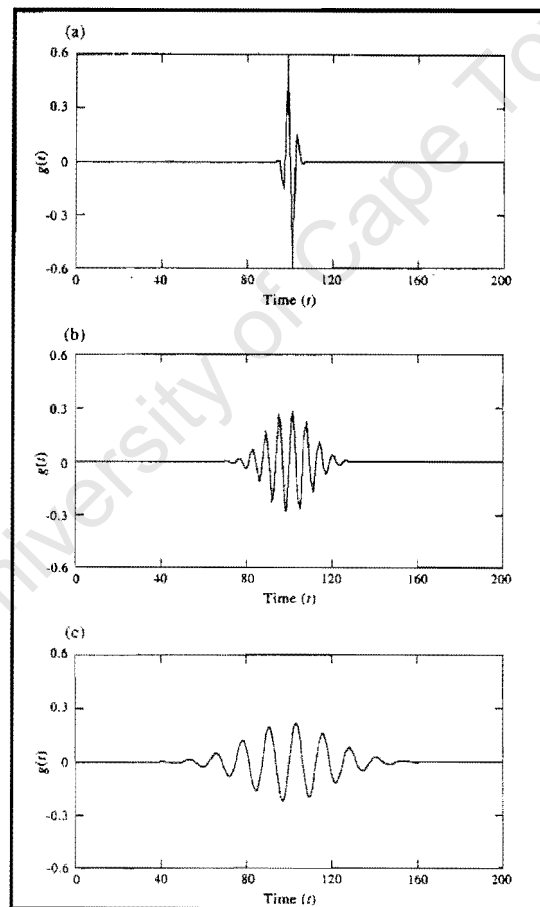


Figure 4.2. The Morlet wavelet, given by, $\psi(t) = \pi^{1/4} e^{-t^2/2} e^{i\omega_0 t}$ (after Emery and Thomson, 1998).

CHAPTER 5

RESULTS AND DISCUSSION

MEAN MONTHLY COMPOSITES

Introduction

Wind-stress is the force exerted by the wind on the sea surface and is responsible for wave and current generation and vertical mixing. Wind-stress curl is a fundamental forcing agent for dynamic ocean processes. It is the external input term in ocean model formulations expressed in terms of vorticity balance. The same process when considered from vertical structure in the flow field appears as the divergence of surface Ekman transport. Wind-stress curl thus controls “Ekman pumping” and associated vertical advection velocity in regions seaward of the immediate effects of coastal boundaries (Bakun and Nelson, 1991).

The global climatic context of 1999–2000 was characterised by a pronounced La Niña event that was in a mature stage during the austral summer of 1999–2000. The results presented below should thus be regarded with caution when comparing them to the climatological average. In order to place the data presented (01 August 1999 – 29 November 2000) into perspective, they are compared to the climatologies presented by Kamstra (1985) and Bakun and Nelson (1991), as well as the long-term climatology (1968–1996) of the surface vector wind field off the coast of southern Africa, as derived from the 2.5° resolution NCEP/NCAR reanalysis dataset².

² <http://www.cdc.noaa.gov/cgi-bin/Composites/printpage.pl>

01 August 1999 – 01 August 2000

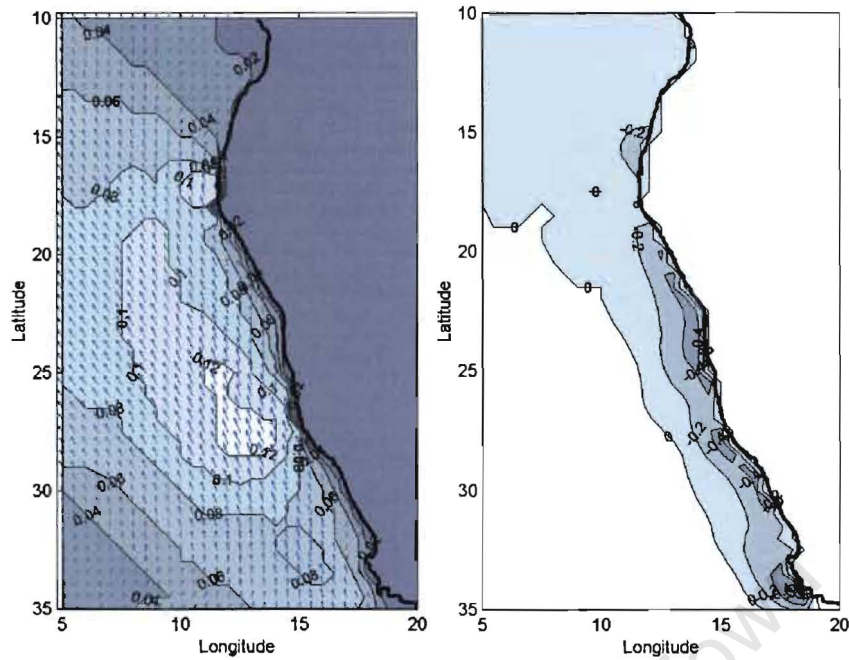


Figure 5.1 The annual distribution (01 August 1999 – 01 August 2000) of wind-stress (N.m^{-2}) and wind-stress curl (N.m^{-3}) over the Benguela Upwelling Regime, as derived from the QuikSCAT satellite. The contour intervals are 0.02 N.m^{-2} and 0.2 N.m^{-3} respectively.

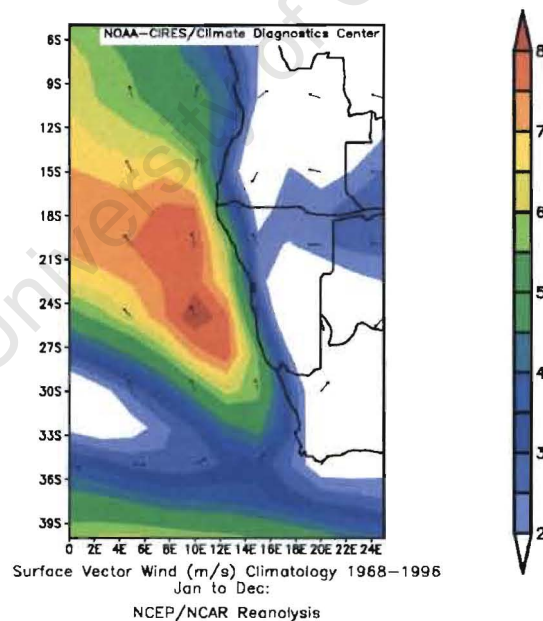


Figure 5.2 The long-term climatology (1968 –1996) of the surface vector wind speed field off the coast of southern Africa, as derived from the 2.5° resolution NCEP/NCAR reanalysis dataset².

The annual distribution (01 August 1999 – 01 August 2000) of QuikSCAT wind-stress and wind-stress curl data is presented in Figure 5.1. Figure 5.2 depicts the long-term climatology (1968 –1996) of the surface vector wind speed field off the coast of southern Africa. The NCEP/NCAR reanalysis product provides a consistently assimilated dataset of the state of the atmosphere at 28 levels in the vertical, with a temporal resolution of four times daily and a spatial resolution of 2.5°. For more detailed description of the NCEP/NCAR reanalysis product, the reader is referred to Kalnay *et al.* (1996). Mean monthly distributions of wind-stress are presented in Figure 5.3(a-l).

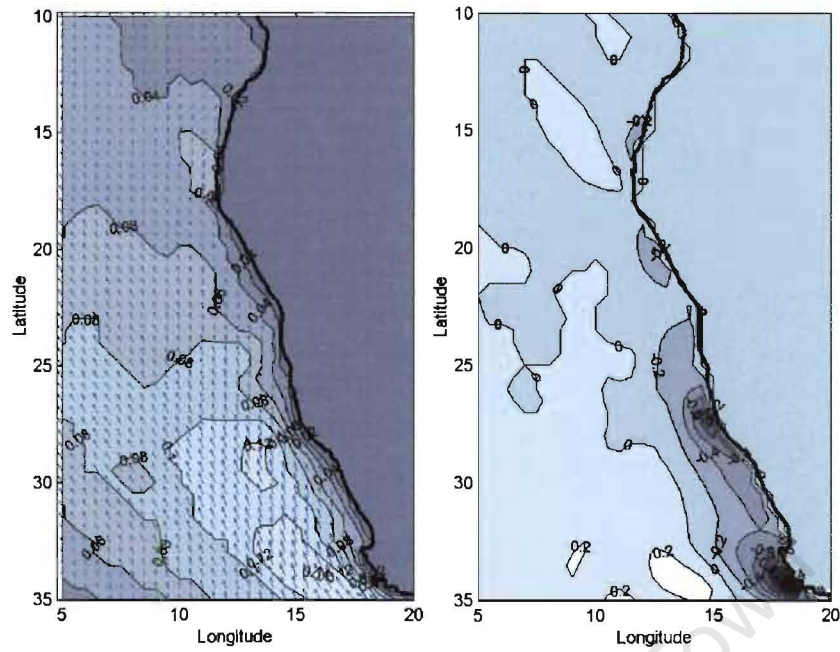
01 August 1999 – 29 November 2000 in Context

The global climatic context of the summer of 1999–2000 was characterized by a pronounced La Niña event. This event started in the austral winter 1998 and was in a mature stage during the austral summer of 1999–2000. Using NCEP/NCAR meteorological parameters at different levels of the atmosphere, Roy *et al.* (2001) compared each month of the 1999–2000 summer with a composite of summer months since 1949 during which La Niña was present. These authors found the meteorological setting observed in the South Atlantic Ocean, during the summer 1999–2000, to be similar to that of the atmospheric patterns observed during other La Niña episodes. The predominant characteristics of such events include a strengthened high-pressure system in the Atlantic and Indian Oceans, a southward displacement of the Atlantic high-pressure condition associated with a low-pressure anomaly to the east, extending along the South African and Namibian coastlines, and strengthened south-easterly winds along the coast of South Africa (Roy *et al.*, 2001) (See Appendix 2).

Given the above findings it is quite remarkable, then, that there are so many similarities between the annual wind-stress field (Figure 5.1), as derived from the QuikSCAT dataset, and the long-term climatology derived from the NCEP/NCAR reanalysis dataset (Figure 5.2).

a)

January



b)

February

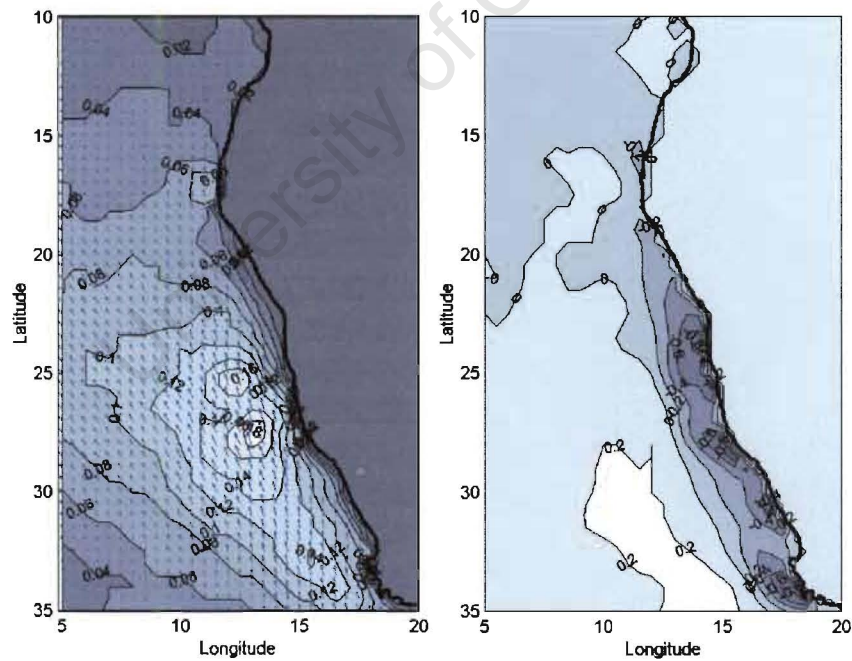


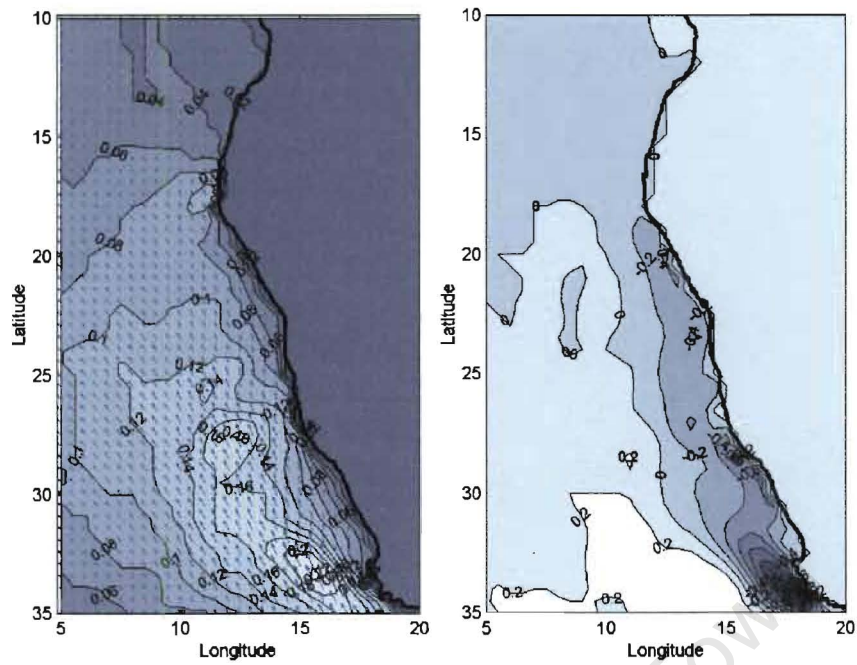
Figure 5.3 The Benguela Upwelling Regime: surface wind-stress (N.m^{-2}) and wind-stress curl (N.m^{-3}) distribution. The contour intervals are 0.02 N.m^{-2} and 0.2 N.m^{-3} respectively. (a) January, (b) February.

Both the QuikSCAT data as well as the reanalysis data show south-easterly winds throughout the Benguela region, with a wind-stress (0.12 N.m^{-2}) situated between 25°S and 28°S and a wind speed (8 m.s^{-1}) maximum situated between 23°S and 26°S respectively (See Appendix 3 for an approximate wind-stress to wind speed conversion graph). These maxima are situated approximately 200 – 250 km offshore. The QuikSCAT data show a persistent secondary maximum just off the coast at Cape Frio (17°S). North of 15°S , and close to the coast, weak south-westerly wind-stress is observed throughout the study period, ultimately becoming south-easterly further offshore. The secondary maximum off Cape Frio, is not resolved in the reanalysis data, possibly due to the coarse spatial resolution of the reanalysis dataset. However, weak south-westerly winds north of 15°S , becoming south-easterly offshore, are clearly resolved. Also resolved in Figures 5.1 & 5.2 is the weakening in both wind-stress and wind speed, respectively, towards the south-west quadrant of the study area (See Appendix 3). The NCEP/NCAR reanalysis diagram covers a larger spatial domain than that of the QuikSCAT wind field. Both the westerly wind belt, as well as the centre of the SAA depicted in the Figure 5.2, are thus not resolved within Figure 5.1.

Roy *et al.* (2001) documented two unusual oceanographic events that occurred off the west coast of South Africa during the 1999–2000 summer. The first event was a strong and sustained SST warming that occurred in mid-December and lasted for two weeks. This warming was attributed to the breakdown of upwelling favourable winds due to a large southerly shift of the SAA, as well as an increased influence of the attendant low-pressure anomaly experienced during La Niña (See Appendix 2). The second event was a sustained south-easterly wind which blew from mid to late summer. During February and March 2000, the centre of the SAA moved eastwards, affecting coastal areas with strong south-easterly winds for an extended period of time. These high pressure conditions were still abnormally developed during April and May 2000, possibly resulting in the prevention of cold fronts and cut-off lows, with associated westerly winds, from disrupting the southerly circulation. These two events were separated by a period of moderate upwelling intensity extending from January to February.

c)

March



d)

April

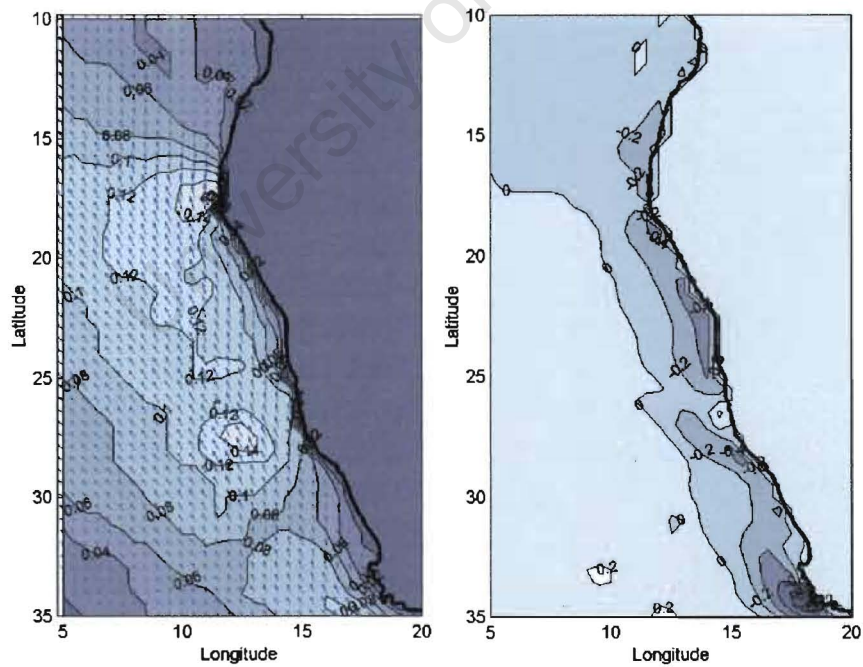


Figure 5.3 (Continued) (c) March, (d) April.

Monthly Wind-Stress Composites for 1999-2000

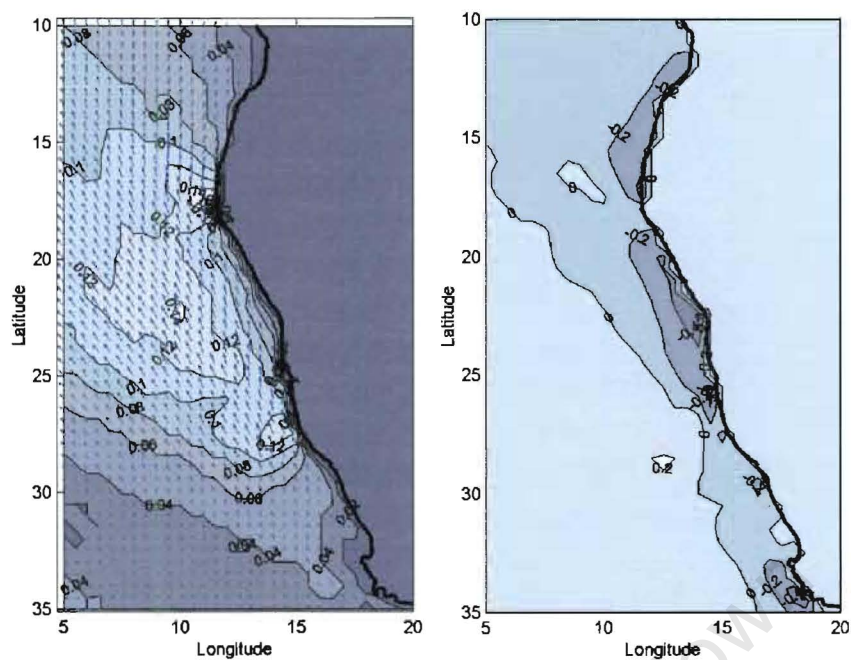
The present QuikSCAT finding of wind-stress maxima at $\sim 25^{\circ}\text{S}$ and 17°S relate well to the results of Parrish *et al.* (1983). According to those authors, the locations of maximum offshore Ekman drift are the Cunene and Lüderitz upwelling cells, situated at 17 and 25°S respectively. Their results, however, differ with those of Lutjeharms and Meeuwis (1987), who showed low levels of wind-stress at the Cunene cell.

Using 24000 ships' records spanning fifty years obtained from the South African Data Centre for Oceanography (SADCO) and from the then South African Weather Bureau, Kamstra (1985) described the general environmental features relevant to wind-speed over the southern Benguela. These data were grouped together in half-degree squares for analysis. For the mean annual wind field, he observed that the strongest wind speeds in the coastal regions were generally to the south of Cape Point. Further north, the wind speeds decreased to a minimum at Elands Bay before strengthening once again at the northern boundary of the study area, situated at 28°S . Two small areas of relatively lower wind speeds were recorded, one just north of Hondeklip Bay and another slightly larger region at Alexander Bay. The Alexander Bay minimum was found to expand northward and intensify during the summer. Wind speeds were found to generally increase at a relatively constant rate with distance offshore along the entire coast from Cape Point to Lüderitz reaching approximately 6 m.s^{-1} 300 km offshore. This corresponds well to the findings of Chao (1985) who investigated topographically forced coastal jets in the lower atmosphere along the west coast of the United States. Lastly, with regards to the mean annual wind field, Kamstra (1985) commented that the wind-speed pattern for summer was very similar to that of the mean annual winds, except that in the summer offshore winds were slightly stronger, thus reflecting the semi-permanent nature of the SAA.

The 16 months of data analysed here show little support for the above findings, with Figure 5.1a containing no maximum south of the 34°S and no minima in the vicinity of Elands Bay, Hondeklip Bay and Alexandra Bay. There is support, however, for the finding that the wind-stress tends to increase to the north. The winds presented here also tend to be stronger than those presented by Kamstra (1985), with wind speed maxima of $\sim 8.5 \text{ m.s}^{-1}$ ($\sim 0.11 \text{ N.m}^{-2}$ See Appendix 3). In addition, there is no

e)

May



f)

June

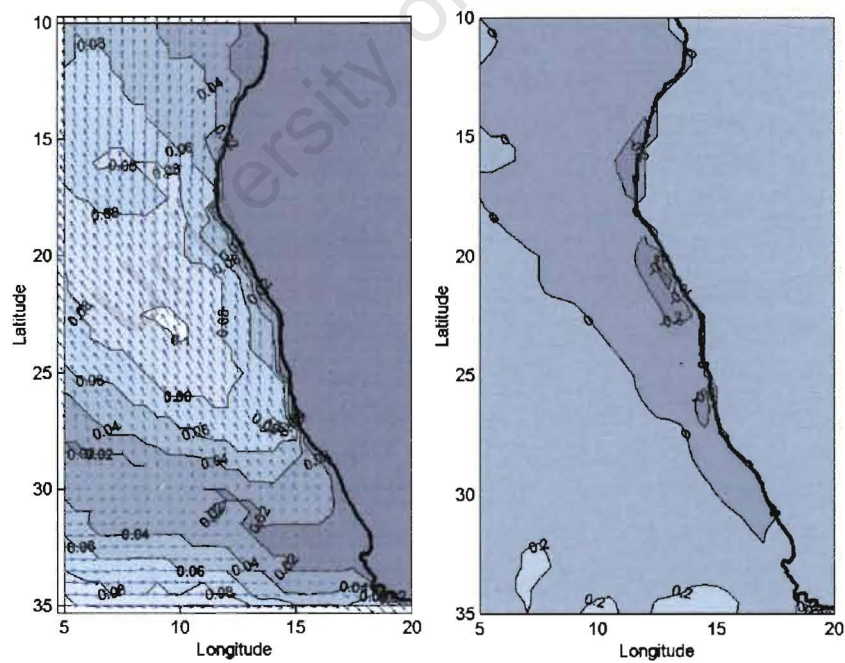


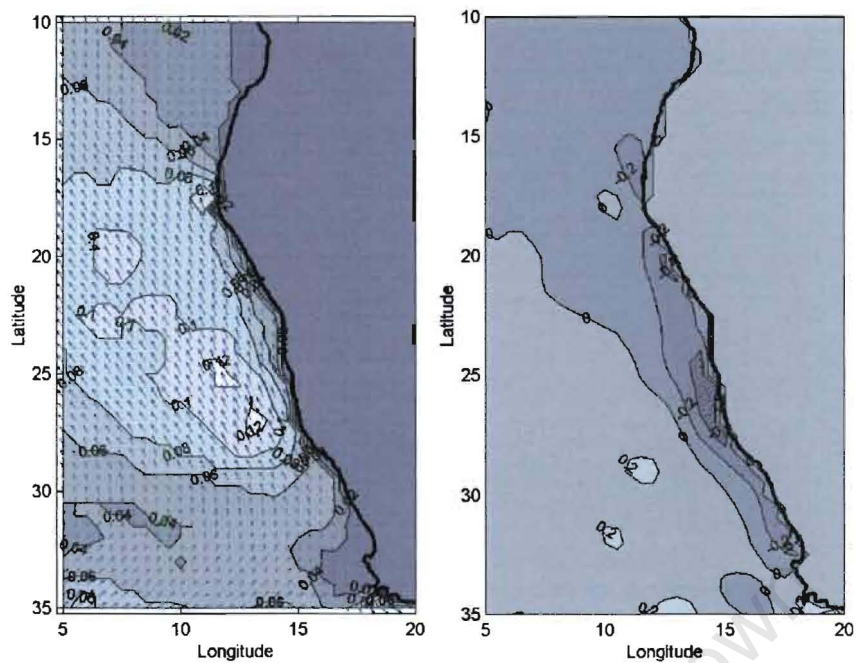
Figure 5.3 (Continued) (e) May, (f) June.

general gradual increase in wind-stress with offshore distance. Wind-stress increases rapidly off the coast at 28°S, while this increase occurs more gradually further south. For December and January, the strongest wind-stress tends to occur to the south of 24°S with stress maxima occurring at 28-30°S and to the west of the Cape Peninsula at 34°S. A weaker relative maximum is found near Cape Frio. Winds intensify during the months of February and March, where wind-stress of 0.14, 0.18, and 0.22 N.m⁻² are recorded approximately 300 km offshore at Lüderitz, Alexandra Bay and Cape Point respectively. The Cape Point maximum was situated in roughly the same position for both January and March, but was further north in February, possibly reflecting the semiannual oscillation (SAO) (Meehl *et al.*, 1998). Associated with this maximum are high (low) levels of temporal (spatial) variability for both the u and v wind-stress components (See Appendix 1). These high levels of temporal variability may infer the presence of a low level coastal “jet”. This “jet” appears to be associated with the intermittent eastward tracking migratory anticyclones that ridge south of South Africa and is consequently associated with the synoptic cycle. Jury (1984) found a shallow equatorward wind “jet”, driven by the SAA, to occur relatively frequently to the west to north-west of the Cape Peninsula. This “jet” was found to be both channelled and deflected by the Cape Peninsula. Using vertical cross sections, Jury (1984) illustrated the 17-19 November 1979 evolution of such a wind “jet” which extended across the Cape Peninsula. This “jet” reached speeds of 22m.s⁻¹ (>0.65 N.m⁻², see Appendix 3) at the 150 m level just off the Peninsula (some 40 km seawards from the mean coastal alignment). The core of the jet was constrained below the 500 m level by a sharp air temperature inversion, which tilted downwards towards the coast. Many characteristics of this wind jet are comparable to those found by Zemba and Friehe (1987). The secondary maximum identified earlier off the coast at 17°S is maintained throughout the months of February and March, strengthening to 0.1 N.m⁻² in March.

By April the seasonal movement of the SAA becomes apparent, with the most intense stress occurring to the north of 28°S. The maximum off Cape Frio has further intensified to ~0.14 N.m⁻². This is possibly related, in part, to a late summer intensification of the tropical low over southern Angola. May brings with it a significant change to the wind-stress of the southern Benguela, as weak southerly to south-westerly wind-stress is experienced south of 28°S. The most intense wind-stress

g)

July



h)

August

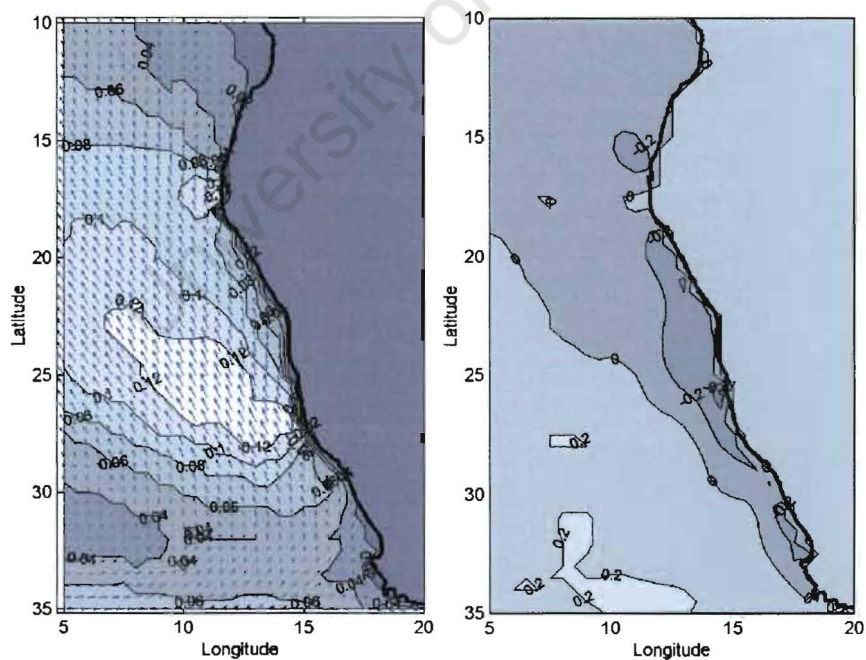


Figure 5.3 (Continued) (g) July, (h) August.

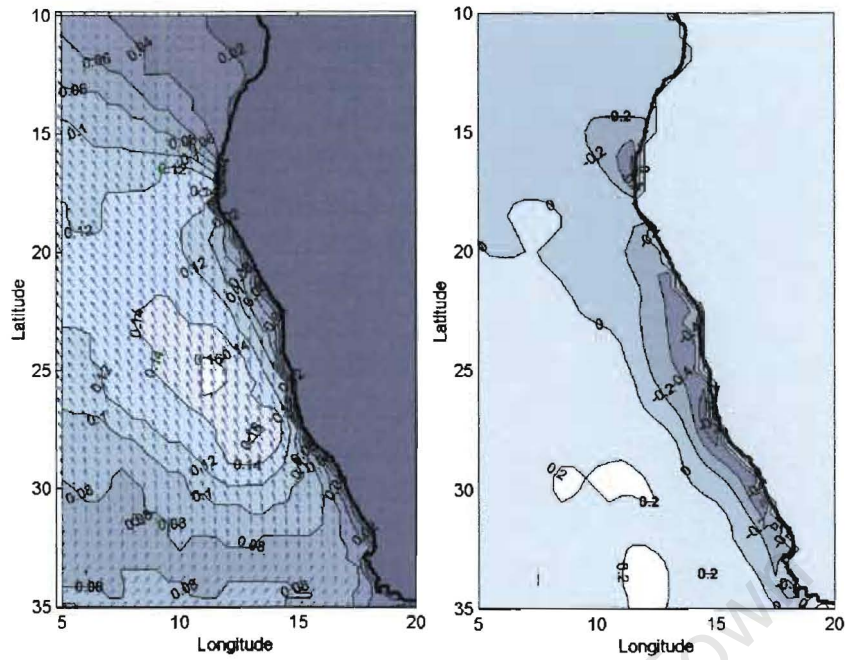
occurs ~400 km off the coast at 22.5°S, with the strong wind-stress situated just off the coast at 17°S and 28°S providing secondary maxima. The wind field for June clearly depicts the dominance of the westerly wind belt in the southern Benguela. A significant stress minimum, that began to develop in May, is shown to expand to the north of Cape Columbine. This u-shaped minimum stretches both offshore for several hundred kilometres, as well as reaching as far north as 28°S, finally wedging in behind the local maximum situated immediately to the west-north-west. The most significant maximum occurs at 22.5°S:10°E. The local stress maximum off the coast at 17°S is still present, if somewhat weaker ($\sim 0.6 \text{ N.m}^{-3}$). July, August and September all show very similar wind-stress fields, with the maximum occurring north of 30°S. There is relatively weak southerly to south-westerly wind-stress characterising the southern Benguela. Off the coast of Cape Frio, the wind-stress maximum strengthens ($\sim 0.1 \text{ N.m}^{-3}$), appearing as a lobe that stretches away to the north-north-west. A reduction in the stress minimum to the north of Cape Columbine is also evident. For October, strong south-easterly wind-stress is found throughout most of the study region, with the exception occurring along the coast north of 15°S where light south-westerly wind-stress occurs. This onshore flow is possibly associated with the tropical low that develops over southern Angola in spring. The area of maximum stress occurs in a 200 km wide band between 100 and 200 km offshore, stretching from the Cape Peninsula in the south to Alexandra Bay in the north. As spring progresses, the wind-stress tends to lessen somewhat. However stress maxima are still to be found close to the coast at 17°S ($\sim 0.14 \text{ N.m}^{-2}$) and 27.5°S ($\sim 0.12 \text{ N.m}^{-2}$), and further offshore at 35°S ($\sim 0.1 \text{ N.m}^{-2}$).

Monthly Wind-Stress Curl Composites

Using historical ship data ($1^\circ \times 1^\circ$ resolution), extracted from the NOAA National Climatic Data Centre's (NCDC) global file of surface marine winds, as monthly means, Bakun and Nelson (1991) reported that the coastal area of cyclonic wind-stress curl appeared as a wedge extending southwards from $\sim 20^\circ\text{S}$. This wedge narrowed in offshore extent towards the south, having its most limited poleward extent during the winter months when it reached to just south of Lüderitz. The wedge had extended to Cape Columbine by early spring and beyond the Cape Peninsula by late spring. The single year average derived from QuikSCAT data presented here is partially at odds

i)

September



j)

October

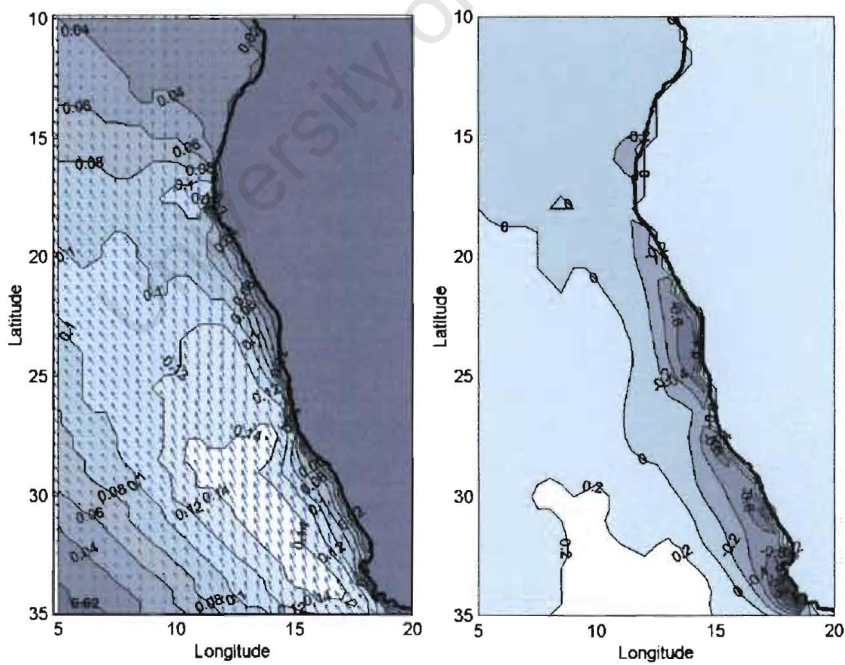


Figure 5.3 (Continued) (i) September, (j) October.

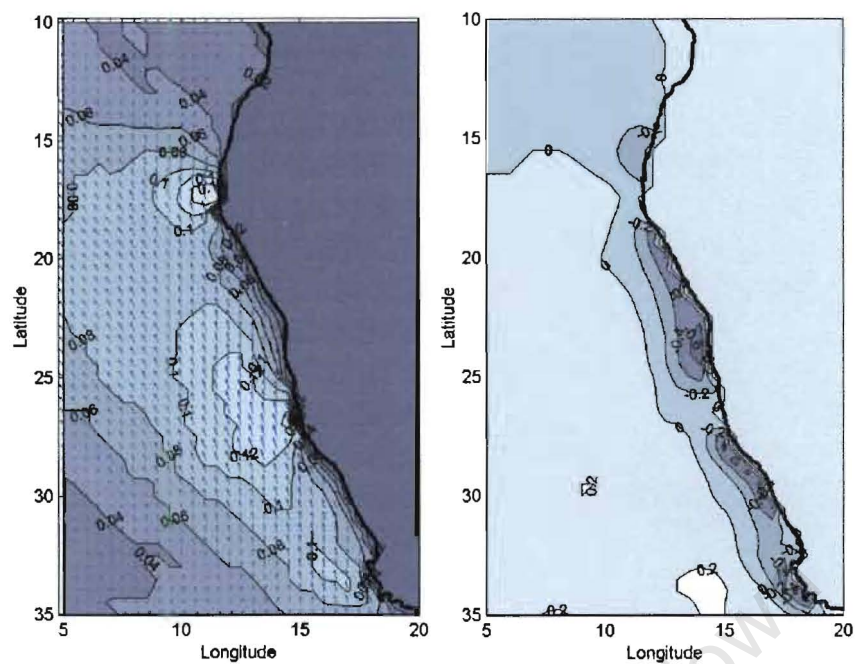
with the Bakun and Nelson (1991) climatology. In the former, the coastal area of wind-stress curl occurs more as a 200km wide band along the entire coastline, with local spring/summer maxima occurring at approximately 26, 28 and 34°S. A wedge of cyclonic curl derived from QuikSCAT winds does, however, appear during winter. This wedge extends southward from approximately 18°S to 32°S in June, to 33.5°S in July and south of the Cape Peninsula in August.

For the month of March, an intense band of cyclonic wind-stress curl is found to the west of Cape Point. This band, which contains values in excess of -1.4 N.m^{-3} , is aligned inshore of the axis of maximum wind-stress. A weaker local maximum is located to the north at 30°S, with curl values reaching -0.6 N.m^{-3} . The Cape Peninsula maximum weakens greatly during the months of April and May, disappearing completely in June, and developing once again in early spring. A northward migration of cyclonic curl occurs in April and May, with 100-200 km wide bands of cyclonic curl occurring between 18 and 25°S and 14 and 17°S, the latter stretching as far north as 11°S in May.

Bakun and Nelson (1991) recorded a similar seasonal migration of wind-stress curl. They were also able to identify an area of persistent anticyclonic curl, some hundreds of kilometres in diameter lying 200km offshore, and extending somewhat northward of Cape Frio. This area of anticyclonic curl is also found in the results presented here. However this region displays a greater amount of spatial as well as temporal variability than reported by Bakun and Nelson (1991), who found evidence for the presence of this anticyclonic curl throughout the year. The monthly averages presented here for 01 August 1999 – 29 November 2000 show it to only occur from late spring through to early autumn. The area of anticyclonic curl to the north of 15°S during late winter and spring, as found by Bakun and Nelson (1991), does not appear in the QuikSCAT data presented here. The reason for this disparity may be due in part to the fact that a La Niña event resulted in the south-east Atlantic experiencing anomalous wind-stress conditions for the 1999-2000 period (Roy *et al.*, 2001).

k)

November



THE SOM ANALYSIS

Synoptic Wind-Stress States

A 6x4 SOM output array of the QuikSCAT wind-stress vectors for the entire dataset (01 August 1999 – 29 November 2000) is shown in Figure 5.4. Each map represents a synoptic pattern within the data, constructed from the weights on that particular node. Patterns are spread in a continuum across the two-dimensional node space. On the left side of the SOM array, wind-stress is predominantly westerly to south-westerly (south of 25°S), whereas on the right side, wind-stress is southerly to south easterly throughout the region. The number of input images that are similar to each pattern, given as a relative frequency, are shown in the top right of each pattern. All output patterns were represented in the data, with the largest percentage (7.8%), of the scatterometer images mapping to node 9. The stress vectors are plotted at every fourth point, to aid interpretation.

Frequency Analysis of Synoptic States

The relative frequency of each synoptic state for each season is shown in Figure 5.5. This map can be visualized as being superimposed on the SOM array, so that the coordinates of the relative frequency map correspond to those of the SOM array. The top left corner of the frequency map thus corresponds to the top left corner of the SOM array. Frequency maps highlight the intra-seasonal to semi-annual variability in synoptic states. The frequency of node mappings for spring and autumn are spread relatively evenly across the entire SOM array, with relative maxima of approximately 12% at nodes 9 and 14 respectively. Node 9 depicts relatively strong south-easterly wind-stress close to the coast. Moving north from the Cape Point (34.5°S) towards Lüderitz (26.5°S), these south-easterlies increase in both stress and offshore extent, with the area of maximum stress occurring at 25°S:12°E. By contrast, node 14 depicts a band of relatively strong south-easterly wind-stress between 15 and 28°S, with a relative maximum at 18°S:11°E. Light south-westerly wind-stress is found off the Cape Peninsula. This stress weakens and becomes more southerly in direction towards 28°S. The most frequent pattern in summer (~18%) is node 10, which has a more uniform south to south-easterly wind field in comparison to that of spring and autumn.

Wind-stress tends to be marginally stronger in the southern Benguela, south of 28°S , with the wind-stress off the Cape Peninsula having a strong south-easterly component. Further north, the wind-stress becomes more southerly in direction.

The winter frequency map differs markedly from those of the other seasons, with the most frequently mapped states occurring to the left side of the SOM array. These patterns tend to contain westerly, north-westerly, and south-westerly wind-stress south of 28°S , reflecting the seasonal shift in the midlatitude westerly wind-belt and the northwards retreat of the SAA. Relatively strong south-easterly wind-stress occurs further north between 10 and 28°S . The winter frequency plot is bimodal, with nodes 2 and 22 having a frequency of approximately 12% each. Node 2 shows strong north-westerly wind-stress south of 28°S . Wind-stress closer to the coast tends to be lighter than the offshore region, in addition to having a more easterly direction. Further north, there is once again an increase in wind-stress, with wind-stress north of 26°S containing a strong southerly component. Node 22 contains a substantially different pattern in comparison to node 2. There is a strong anticyclonic component to the circulation, with high wind-stress confined to a band between 18 and 28°S . Within this band, the strongest wind-stress is situated off the coast at approximately 26°S .

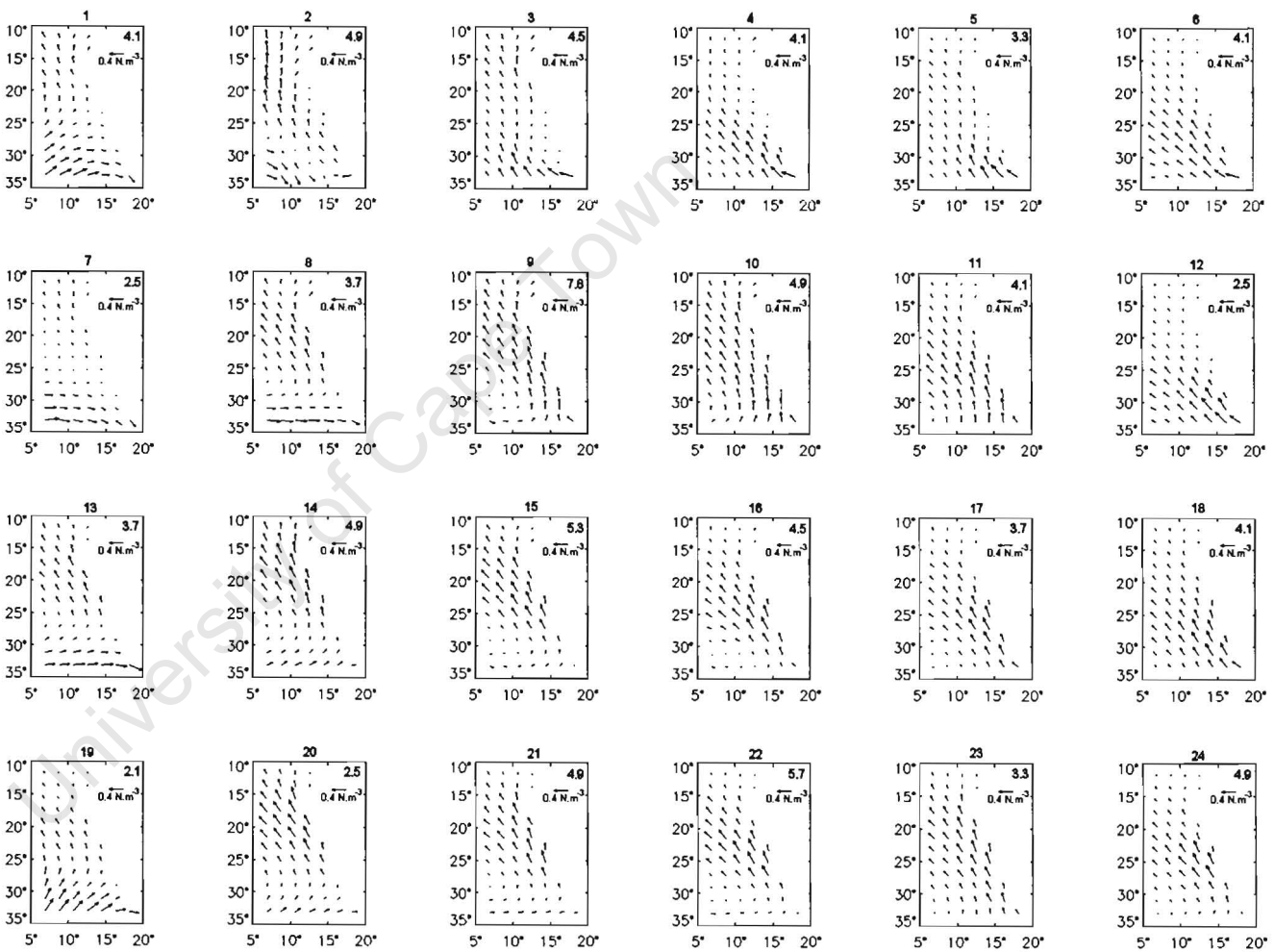


Figure 5.4 A 6x4 SOM of 244 QuikSCAT images (two-day composites) depicting wind-stress

vectors over the south-east Atlantic. Wind-stress vectors are plotted at every fourth point, to aid interpretation. Nodes are numbered 1–24. The number of input images that are similar to each pattern, given as a relative frequency, are shown in the top right of each pattern. A two-step training schedule was used: the initial training consisted of 1000 iterations, a learning rate of 0.2, and an update radius of 5; the second training consisted of 50000 iterations, a learning rate of 0.02, and an update radius of 0.

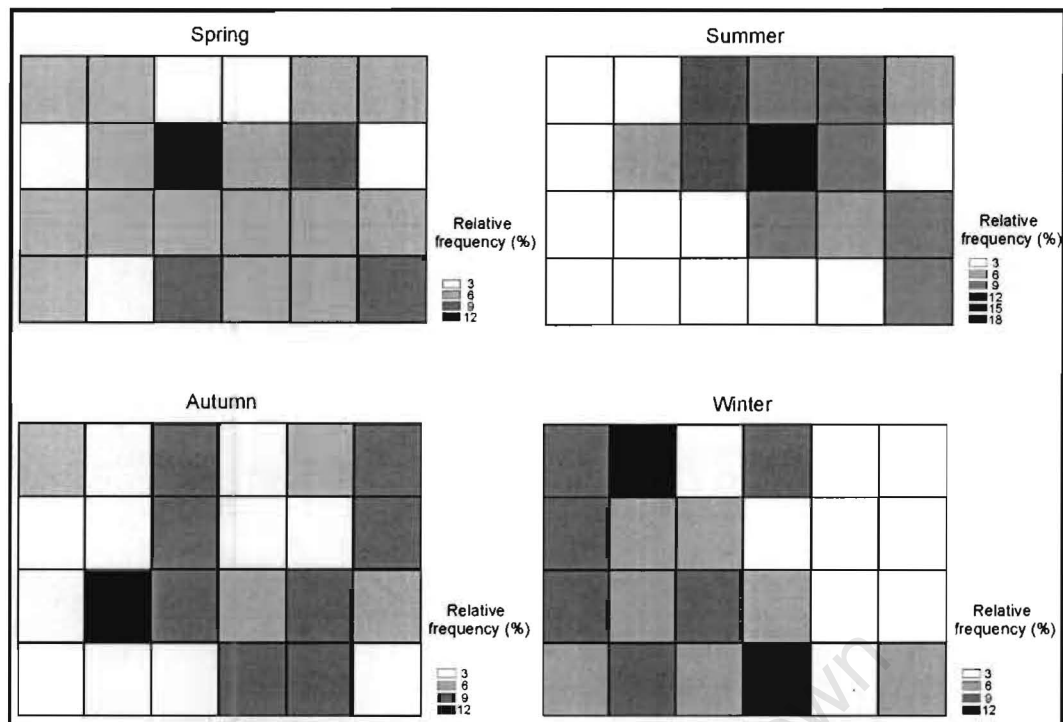


Figure 5.5 Seasonal SOM frequency maps showing the relative frequency of each pattern on the SOM array.

Identifying trajectories

As similar synoptic wind patterns map to adjacent nodes, the trajectory through time of the SOM patterns can be used to examine the temporal evolution of synoptic events. To illustrate this technique, two intense southerly wind events were selected, one in October 1999 and the other in February 2000 (See Figure 5.16 & 5.17). Their trajectories show rapid changes in synoptic states during these events (Figure 5.6). A more detailed discussion of the atmospheric state surrounding each of these events is found below.

14-22 October 1999

For 14 October 1999, the wind-stress maximum lies between 20-26°S, with the south Benguela experiencing relatively weaker southerly wind-stress (Figure 5.6). The daily synoptic charts (Figure 5.7a & b) for this period show a relatively weak cold front to the south-west of Cape Town. On 15 October 1999 the trailing edge of this front was situated over the south-western Cape, resulting in relatively weak westerly winds over the south Benguela. The stronger wind-stress to the north of 26°S (Figure 5.6) is

possibly associated with the increased pressure gradient along the Namibian coast as a result of the surface trough situated over Namibia (Figure 5.7a). On 16 October 1999, a relatively uniform south-easterly wind field prevailed over most of the study area. The wind-stress maximum is now situated at approximately 28°S. This southward shift in the maximum is due to the SAA beginning to ridge to the south-east of South Africa on 16 October 1999 (Figure 5.7c). Figure 5.6 shows this atmospheric state to persist until 18 October 1999. The daily synoptic charts support this (Figure 5.7e), showing a marked intensification in wind speed along the coast of the south-western Cape where speeds of approximately 15m.s^{-1} were measured (Cape Town). Figure 5.7f (19 October 1999) shows the ridging of the migratory anticyclone that budded off from the SAA south of South Africa. Absorption of this migrating anticyclone by the South Indian High Pressure is shown in Figure 5.7g (20 October 1999). The winds associated with the west coast trough, detailed in Figure 5.7g, are not depicted for 20 October 1999 in Figure 5.6. However, the southerly shift in the wind-stress maximum, associated with the budding off of the migrating anticyclone, is clearly shown in Figure 5.6. On 22 October 1999 (Figure 5.6) there was a rapid movement back to the initial atmospheric state of 14 October 1999, i.e. a relatively uniform wind field over the entire Benguela region with relatively stronger south-easterly wind-stress to the north of 26°S. This is consistent with interpretation from the daily synoptic charts for 21, 22 and 23 October (Figures 5.7h, i and j) which show relatively weak pressure gradients over the west coast of southern Africa. The relatively weak wind-stress to the south appears to be as a result of a mid-latitude cyclone and its associated front passing to the south of South Africa and in so doing modulating the wind-stress in the southern Benguela.

11-21 February 2000

For 11 February 2000, there is a reasonably uniform southerly to south-easterly wind field over most of the study area (Figure 5.6), which is consistent with the daily synoptic charts presented in Figure 5.19a & b. The daily synoptic charts (Figure 5.19c & d) for 13 and 14 February 2000 show a slight north-easterly shift in the SAA. In addition to this a westward shift in the interior heat low, which includes the possible absorption of the trapped coastal low, as well as the ridge of high pressure extending along the south coast of South Africa, all result in an increase in the pressure gradient between 26 and 32°S (Figure 5.19d). This is reflected as a belt of relatively strong

south-easterly wind-stress between 23 and 28°S (Figure 5.6) on both 13 and 15 February 2000. By 17 and 19 February, this band of strong south-easterly wind-stress had shifted south (Figure 5.6) in sympathy to the SSA which begins to ridge south-east of South Africa over the two-day period of 15 and 16 February 2000 (Figure 5.19e & f). The relatively strong pressure gradient seen on 14 February (Figure 5.19d) is maintained through 15 February, but has all but vanished by 16 February. 16 February also sees the interior heat low extending southwards in the Northern Cape region. Offshore airflow from the escarpment, as indicated by the wind direction at Windhoek, has assisted in formation of another trapped coastal low at 25°S (Reason and Jury, 1990). Figures 5.19g & h (17 and 18 February 2000) show the ridging south of South Africa and absorption by the South Indian High Pressure of the migratory anticyclone that budded off from the SAA. Also visible is the marked decrease in the pressure gradient off the west coast of southern Africa. This atmospheric state is maintained for 19 and 20 February 2000 (Figures 5.19i & j), with the region returning to an atmospheric state similar to that of 11 February 2000 by 21 February 2000 (Figure 5.6).

Examining the temporal evolution of synoptic events using the SOM can provide a useful method to compare and contrast wind events. For instance, both events returned to an atmospheric state close to that of the initial state. However, this was achieved via two markedly different trajectories. For each trajectory, the same pattern on the SOM array, i.e. pattern 18 for the October event and pattern 24 for the February events, was repeated over consecutive days, indicating that the system was stationary for 3-4 days.

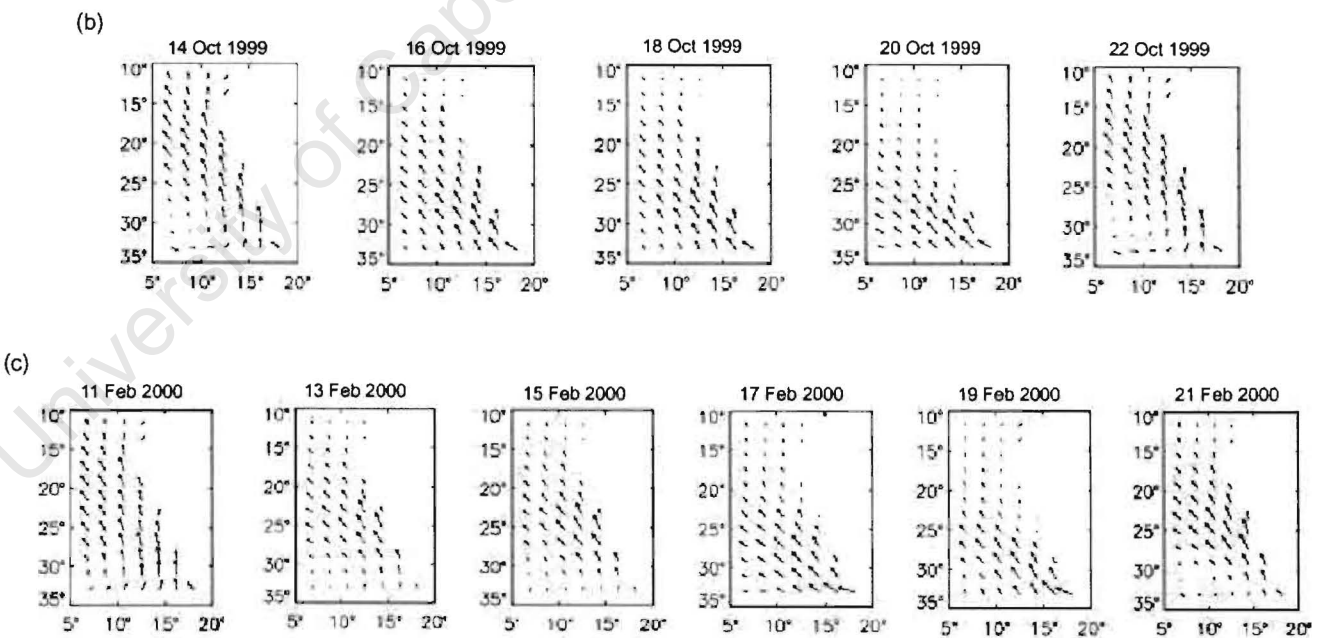
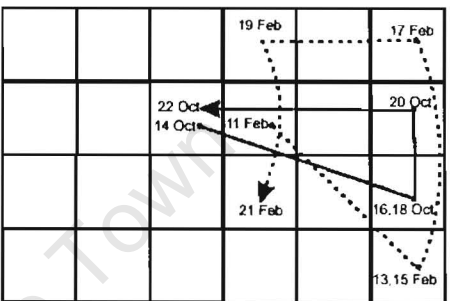


Figure 5.6 SOM trajectories of two intense wind events. The first event led to a wind-stress maximum off the coast at 34°S and the second event resulted in a wind-stress maximum off the coast at 26°S. (a) Trajectories overlaying SOM array, (b) the nodes visited during the October 1999 wind event and (c) the nodes visited during the February 2000 wind event.

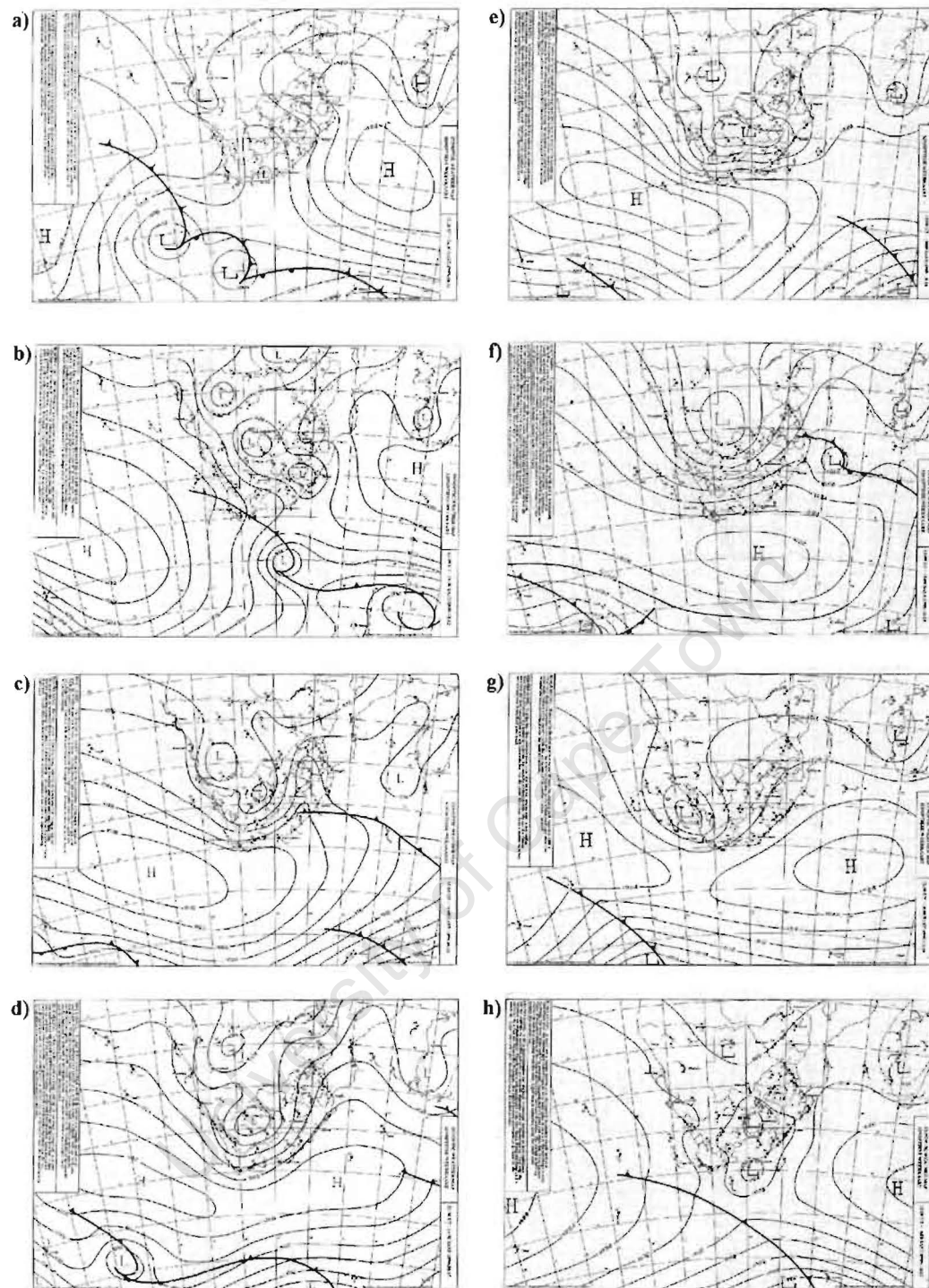


Figure 5.7 Daily synoptic weather maps, produced by the South African Weather Service, for the period 14 – 22 October 1999.

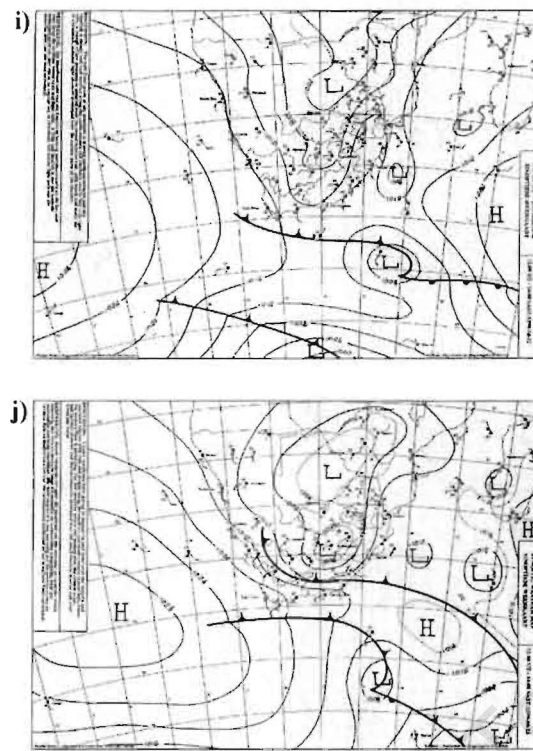


Figure 5.7 Daily synoptic weather maps, produced by the South African Weather Service, for the period 14 – 22 October 1999.

Regional Wind Cells

The data set for the second SOM analysis consisted of the u and v -components along a transect 200km offshore from 10-35°S, for every 0.5° of latitude (See also Figures 5.16 and 5.17). The output of the second SOM analysis, grouped the data into six time series, representing six distinct wind regions or cells (Figure 5.8). Each time series represents the wind-stress magnitude and direction, for the period 01 August 1999 – 29 November 2000, for that particular region. These six time series were smoothed with a 15-point (30 day) running mean (Figure 5.9) so as to aid interpretation.

Latitude (°S)	Wind-Stress Cell Number
10 – 15	Cell 1
15.5 – 18.5	Cell 2
19 – 23.5	Cell 3
24 – 28.5	Cell 4
29 – 32.5	Cell 5
33 – 35	Cell 6

Table 5.1 The above table gives details of the 6 regional cells, as derived from the SOM analysis of the extracted data set, and their respective latitudinal positions

The most northern cell, Cell 1, lies between 10° and 15°S. This region is dominated by light south-westerly wind-stress throughout the study period, with an average wind-stress of 0.3 N.m^{-2} . This onshore flow is possibly associated with the tropical low that occurs over Angola. A period of stronger wind-stress occurs from April 2000 through to the end of June 2000, where values reach in excess of 0.8 N.m^{-2} .

Cell 2 (15.5-18°S) differs substantially from Cell 1 and is more representative of the wind-stress experienced over the Benguela Upwelling regime. Here, the wind-stress has a strong and consistent south-south-easterly component to it, with an average wind-stress approximately 0.1 N.m^{-2} . The period of stronger wind-stress occurring to the north is also evident within this region. Again, the most substantial peak in wind-stress occurs at the beginning of May 2000, where wind-stress of 0.2 N.m^{-2} is

recorded. Periodic wind events, where wind-stress exceeds 0.2 N.m^{-2} , are evident from Figure 5.8. These events tend to be confined to the winter months. Secondary maxima in wind-stress occur mid-way through July 2000 and again towards the end of August 2000.

Cell 3 ($19\text{-}23.5^\circ\text{S}$) contains a similar wind-stress pattern to that of Cell 2, with the wind-stress containing a relatively strong and consistent south-south-easterly component. Once again, wind-stress values of approximately 0.1 N.m^{-2} are recorded (Figure 5.9). The three periods of stronger wind-stress, evident in Cell 2, namely from April 2000 through to the end of June 2000, mid-way through July 2000, and towards the end of August 2000, are also evident in Cell 3. However, the April - June maximum is marginally weaker and less pronounced. The beginning of a period of weaker wind-stress is evident towards the end of November 1999, with this minimum becoming more marked to the south.

Within Cell 4 ($24\text{-}28.5^\circ\text{S}$) the wind-stress maintains a strong and consistent south-south-easterly component, similar to that of the northern cells. Three periods of relatively strong wind-stress ($\sim 0.1 \text{ N.m}^{-2}$) are evident. The first period extends from the beginning of the time series, 01 August 1999, to roughly the beginning of December 1999. The second period extends from approximately the beginning of January 2000 to the beginning of May 2000, while the last period again lasts approximately five months from mid-June 2000 through to mid-November 2000. Separating these maxima are two shorter periods of relatively low wind-stress ($\sim 0.05 \text{ N.m}^{-2}$) lasting for approximately one month each. Also noticeable in Cell 4 is the increase in wind-stress variability relative to Cell 3. This increase in variability may be due to the fact that Cell 4 is affected to a greater degree than Cell 3 by both migrating anticyclones and westerly disturbances passing south of South Africa.

To the south lies a region that stretches from 29°S in the north to 32.5°S in the south, Cell 5. It is here that wind reversals, i.e. north-westerly winds, become more prominent with the intrusion of the westerly wind belt during the winter months. However these are infrequent and wind-stress with a south-south-easterly component still dominates. This region contains reduced wind-stress for the months of May, June

and July 2000. A maximum in stress occurs in mid-October 1999. No such maximum occurs during October 2000.

Cell 6 (33-35°S) is the most southern cell. From the time series contained within Figure 5.9, it appears as though this particular cell contains five distinct periods. The first period extends from approximately the beginning of August through to mid-October 2000. This period is dominated by relatively weak south-westerly wind-stress. This period is followed by a lengthy period, mid-October 1999 through to mid-May 2000, when relatively strong south-easterly wind-stress ($\sim 0.1 \text{ N.m}^{-2}$) is dominant. For approximately the last six weeks of this particular period there is a steady decrease in wind-stress magnitude, however the south-easterly component is maintained. The period of mid-May through to the end of July 2000, contains weak north-westerly winds, associated with the northerly migration of the westerly wind belt. The beginning of August 2000 brings with it, once again, light southerly to south-westerly wind-stress. This period lasts until approximately the end of September. The final period extends from the beginning of October to the end of November 2000. As in 1999, this period is dominated by relatively strong south-easterly wind-stress.

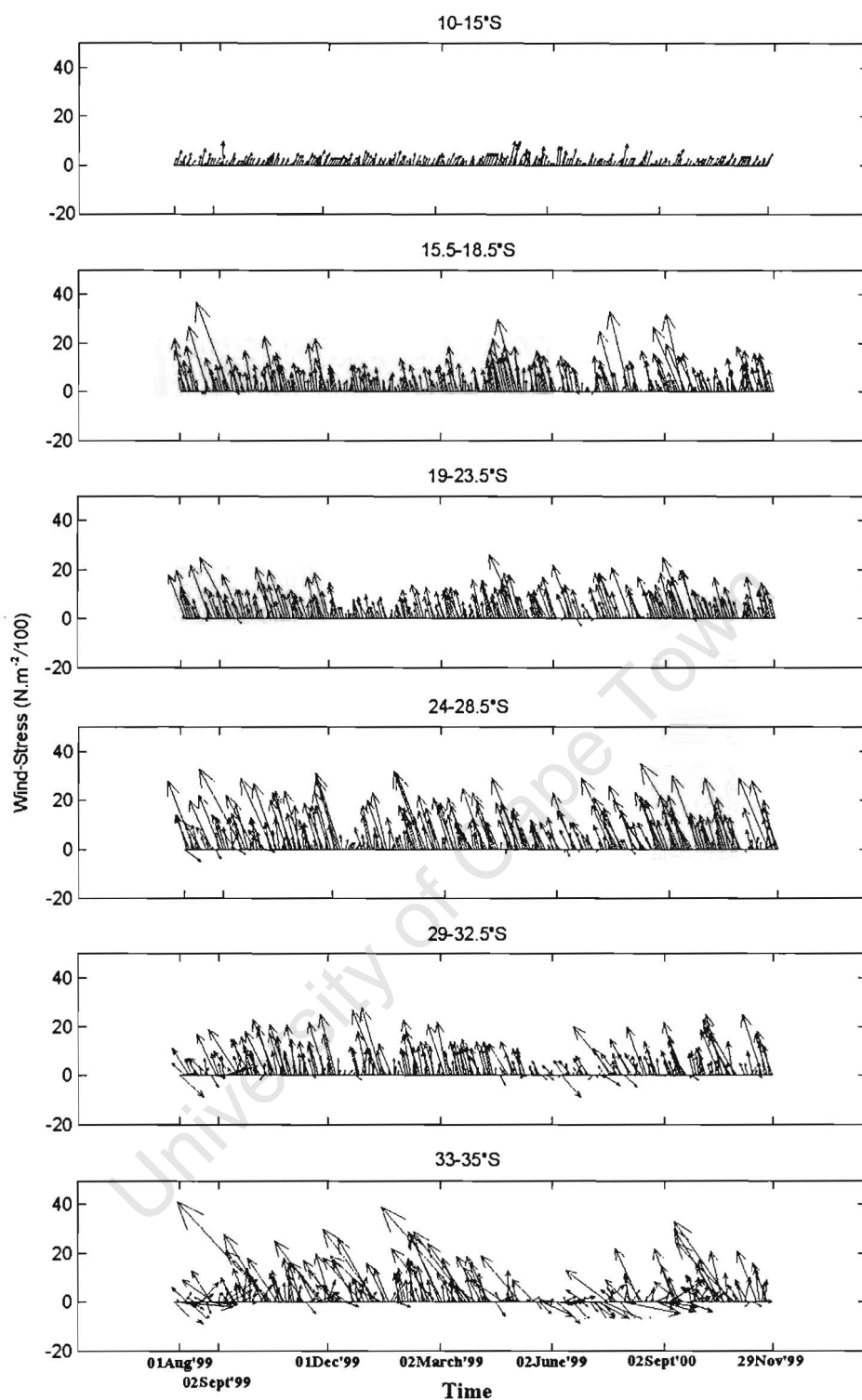


Figure 5.8 Feather plot of the six regional wind cells (10–15°S, 15.5–18.5°S, 19–23.5°S, 24–28.5°S, 29–32.5°S, and 33–35°S) as derived from the SOM. Each time series represents the typical wind-stress magnitude ($\text{N.m}^{-2}/100$) and direction, for the period 01 August 1999 – 29 November 2000.

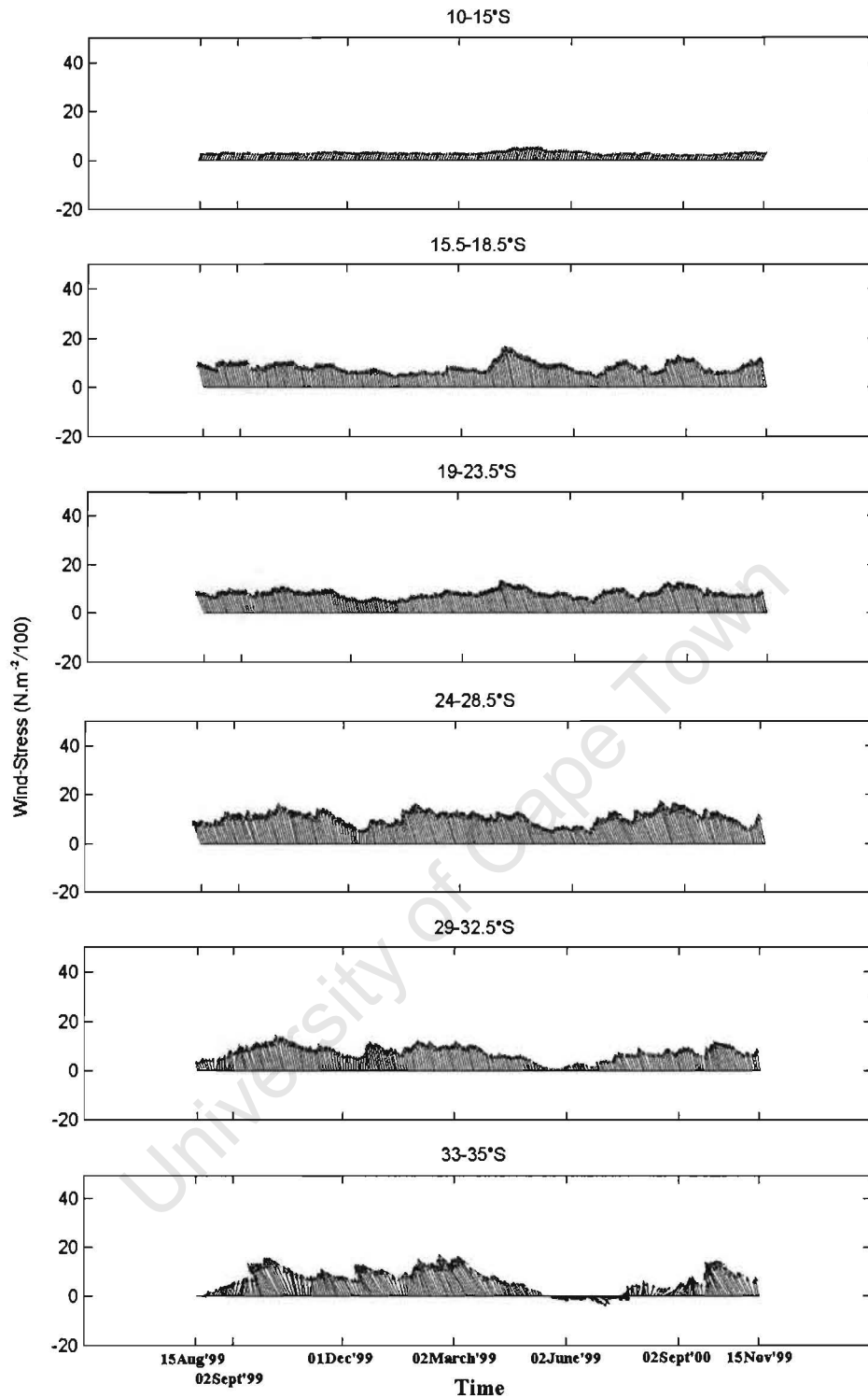


Figure 5.9 Feather plot of the six regional wind cells (10–15°S, 15.5–18.5°S, 19–23.5°S, 24–28.5°S, 29–32.5°S, and 33–35°S) as derived from the SOM. Each time series represents the typical wind-stress magnitude ($\text{N.m}^{-2}/100$) and direction, for the period 15August 1999 – 15 November 2000. All six time series were smoothed with a 15-point (30 day) running mean.

WAVELET ANALYSIS OF REGIONAL WIND CELLS

A wavelet transform is used to conduct a spectral analysis for each of the six cells identified above. In this study, only the v -component of the wind-stress was used. As mentioned above the band of primary interest is the 4–64 day period range.

The standardised time series and the real part of its Morlet wavelet transform, as well as an integration over the total time series, 01 August 1999–29 November 2000, for each of the identified cells is presented in Figures 5.10–5.15 (a, b, & c).

The wavelet power spectra, presented below, span a range of frequencies with periods from approximately 4–64 days. The exact form of these spectral peaks is partly governed by the wavelet “uncertainty principal” (Chui, 1992) which precludes the possibility of achieving high localisation in time and frequency simultaneously, and thus depends somewhat on the choice of wavelet basis function. That said, the event-like, intermittent, and pulsating nature of the variability within the Benguela upwelling system is one attribute of these data that stands out. Regions of high variability are well resolved and tend to appear as isolated peaks. The occurrence and duration of these spectral peaks is, however, somewhat irregular and strongly dependent on latitude.

Most of the extra-tropical intra-seasonal variability in circulation over the Southern Hemisphere occurs in the westerly wind belt between 40 and 60°S (Jones and Simmonds, 1993), with the latitude of maximum variability shifting from the equatorward side of this latitude range at 200hPa poleward into the circumpolar trough at the surface (Kiladis and Mo, 1998). At lower latitudes, low frequency intra-seasonal variability on the 30–60 day time scale is dominated by the Madden – Julian oscillation (MJO). At higher frequencies, a host of equatorially trapped disturbances are important at low latitudes. These include equatorially trapped Rossby waves (Kiladis and Wheeler, 1995) and mixed Rossby-gravity waves (Liebmann and Hendon, 1990), which in the past have been identified as “easterly waves”. Of these, the MJO in particular displays important relationships to extra-tropical intra-seasonal variability over the Southern Hemisphere (Jones and Simmonds, 1993).

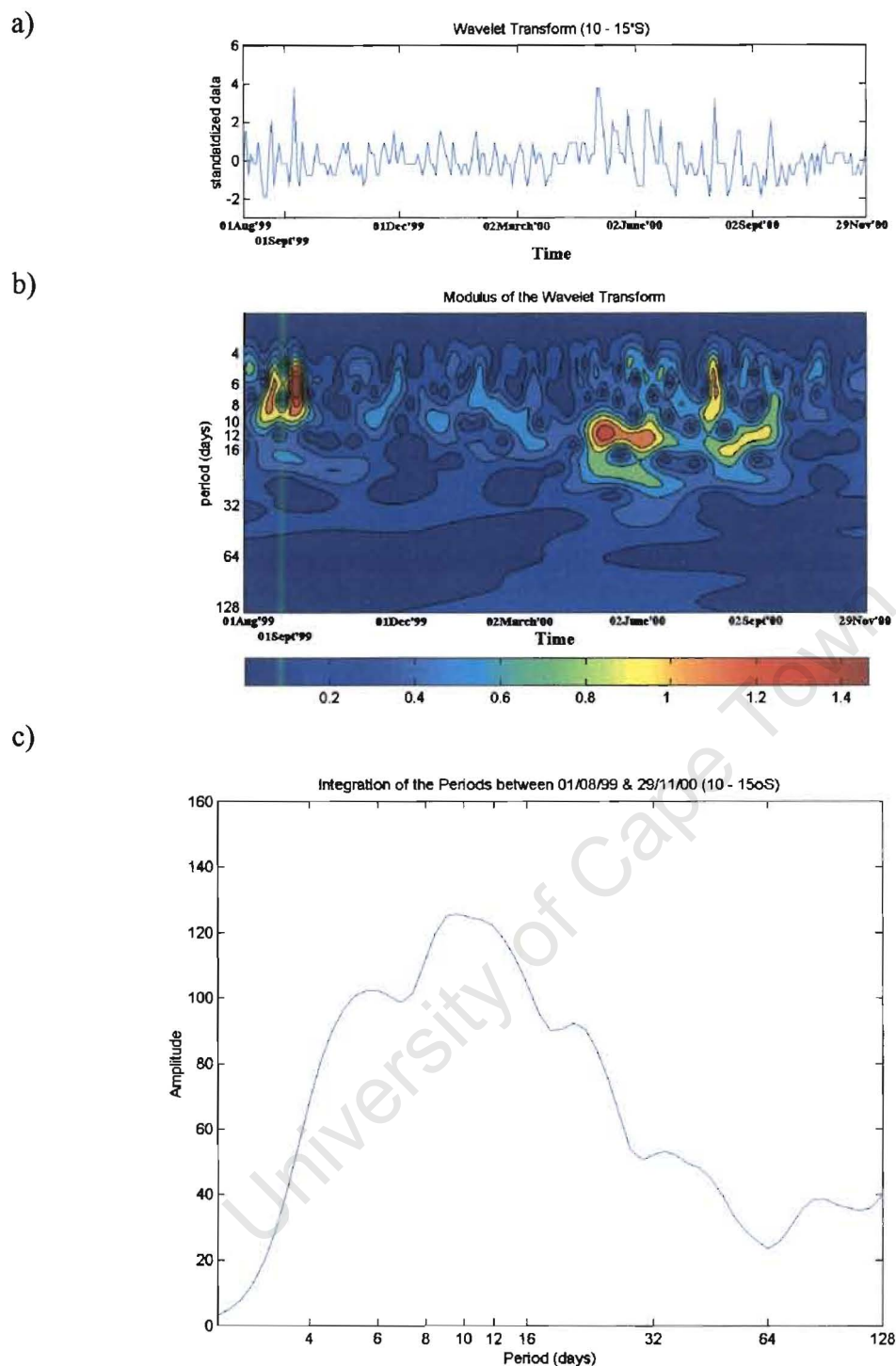
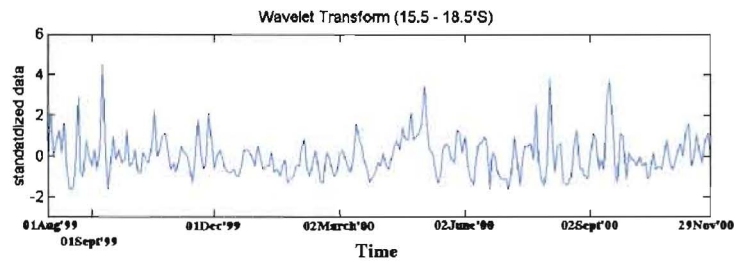


Figure 5.10 Cell 1 (10–15°S). (a) Standardized time series of the longshore (v) component for 01 August 1999 – 29 November 2000, (b) the modulus of the wavelet transform, using the Morlet wavelet, of the longshore (v) component for 01 August 1999 – 29 November 2000, and (c) the integrated wavelet transform for 01 August 1999 – 29 November 2000. The contour interval in (b) is 0.2.

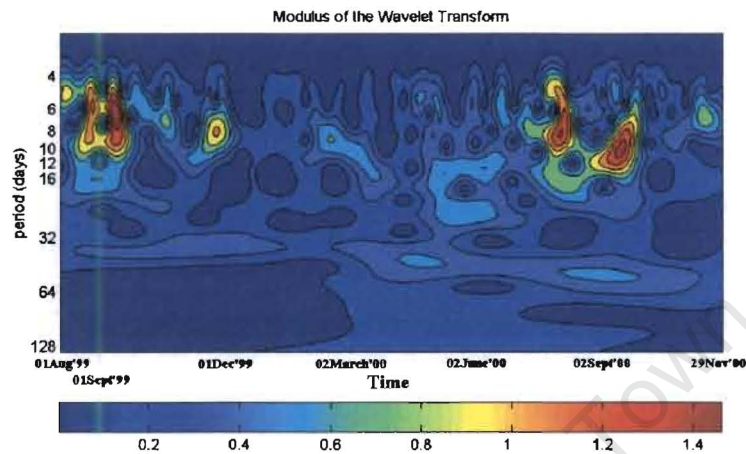
The MJO is an eastward-propagating disturbance that has its maximum amplitude within the tropics of the eastern Indian to western Pacific Oceans. The “oscillation” was discovered by Madden and Julian (1971, 1972) through the spectral analysis of wind and pressure data collected from a network of radiosonde stations. Anderson and Rosen (1983) showed that this tropical oscillation is related to the 50 day oscillation in relative angular momentum of the atmosphere. Anderson *et al.* (1984) showed that these quasi-periodic oscillations have amplitudes and frequencies that vary widely, with periods ranging from 31–63 days. Observations by Parker (1973) and Weickmann (1983) have shown that the range of periods associated with these oscillations can extend from as long as 90 days to as short as 25 days. Hartman *et al.* (1991) investigated the seasonal dependence of the frequency and intensity of tropical oscillations with intra-seasonal time scales. They showed that the MJO in the Indian Ocean region varies in its preferred period of 50 days in the austral summer to approximately 35 days during the austral winter, although evidence is weak. However, during the period September – December, there is a pronounced spectral peak of between 20 and 25 days in the western tropical Pacific. This spectral peak reveals that this intra-seasonal oscillation is distinct from the MJO, but may also result from coupling of large-scale motion and latent heat release on far smaller scales (Hartman *et al.*, 1991).

Generally speaking, the occurrence of spectral peaks in wind-stress for the two northern most cells differ greatly from those situated to the south, with Cell 3 appearing as an area of transition between the so-called tropical cells to the north and the extra-tropical cells to the south. To the north, the maximum in wind-stress occurs between the beginning of May and the end of September for periods ranging from 4 to 24 days. The integration diagrams (Figures 5.10c, 5.11c, & 5.12c) show that there is a maximum at the 10 day period for Cell 1. Further south, this peak shifts to higher frequencies, with Cells 2 and 3 containing an 8 and a 7 day period respectively. Liebmann and Hendon (1990), observed similar results, though at slightly higher frequencies (~4 day period), in the 850 mb meridional wind along the equator. Cell 2 also contains a relatively weaker ~46 day peak. This peak extends throughout the latter part of 1999 before waning and reappearing in mid-February 2000. These findings are supported by Madden and Julian (1994) who concluded that the 40–50

a)



b)



c)

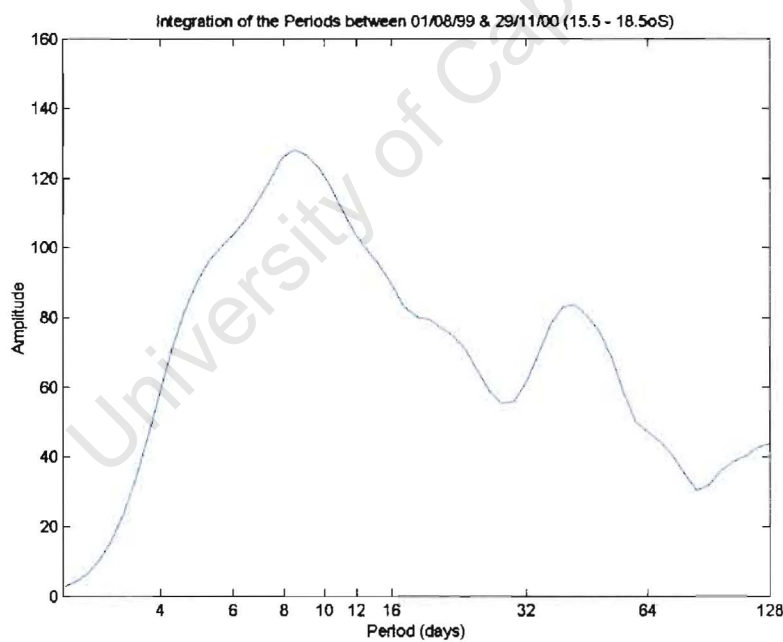
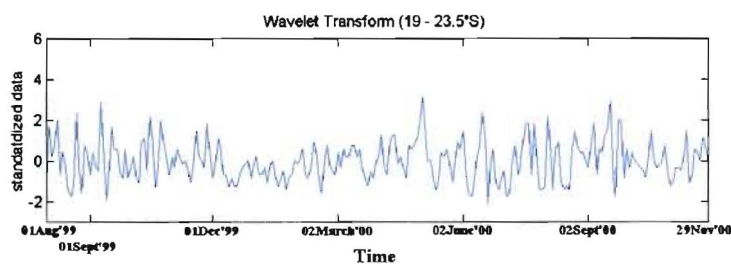


Figure 5.11 Cell 2 (15.5–18.5°S). (a) Standardized time series of the longshore (v) component for 01 August 1999 – 29 November 2000, (b) the modulus of the wavelet transform, using the Morlet wavelet, of the longshore (v) component for 01 August 1999 – 29 November 2000, and (c) the integrated wavelet transform for 01 August 1999 – 29 November 2000. The contour interval in (b) is 0.2.

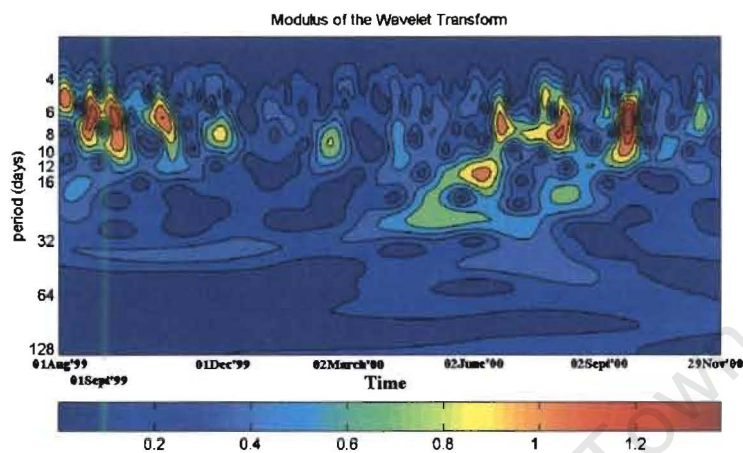
day oscillation, discovered by Madden and Julian (1971, 1972), is a relatively broadband phenomenon with a central frequency that is not fixed but varies in time. This oscillation is also found to be seasonally dependant. Madden (1986) found the oscillation to be associated with the seasonal migration of the Inter Tropical Convergence Zone (ITCZ). He concluded that the oscillation is most significant during the months of December, January and February, and has its smallest amplitude between the months of June and August. During the austral winter 2000, the ~46 day peak appears to strengthen and shift in frequency to approximately a 60 day period. This spectral peak is not visible to the north of 15°S.

The ~46 day peak identified at Cell 2 is not apparent in the Cell 3. However, a ~40 day peak, with a similar intensity as the 46 day peak in Cell 2, is visible for much of the study period between 24 and 35°S. This peak has two relative maxima. These maxima occur in all three of the southern cells, and last for approximately 2 months each. The first peak occurs between November and December 1999, while the second peak occurs between September and October of the following year. These peaks are, however, slightly offset within Cells 4 and 5, with Cell 6 preceding them by 10–15 days. Also noted is the fact that there is a slight decrease in wavelet power within Cell 4 at the ~40 day peak. This intra-seasonal perturbation thus appears to originate south of South Africa prior to influencing the wind-stress at Cell 6, and then, 10–15 days later, the wind-stress at Cells 5 and 4 respectively. What is unclear, however, is whether this ~40 peak has a separate source or is linked to the tropical oscillation, perhaps through atmospheric teleconnections. According to Madden and Julian (1994), it is not unreasonable to expect some impact at higher latitudes because of the magnitude of the tropical disturbance but, in their experience, “robust mid-latitude responses are hard to find”. Alternatively, Dickey et al. (1991) and Marcus et al. (1994) indicate the possibility of a separate non-tropical source for these oscillations, possibly through the interaction of zonal westerly flow with mountain topography (e.g. the Andes Mountains). Park and Schubert (1992) have suggested that a major component of the 20–70 day variability over the Atlantic region is remotely forced. This forcing occurs via the MJO and Rossby wave trains that appear to have their origins in the mid-latitudes of the Pacific. These wave trains are said to make up the dominant contribution to the zonal-wind fluctuation pattern in the Southern Hemisphere.

a)



c)



c)

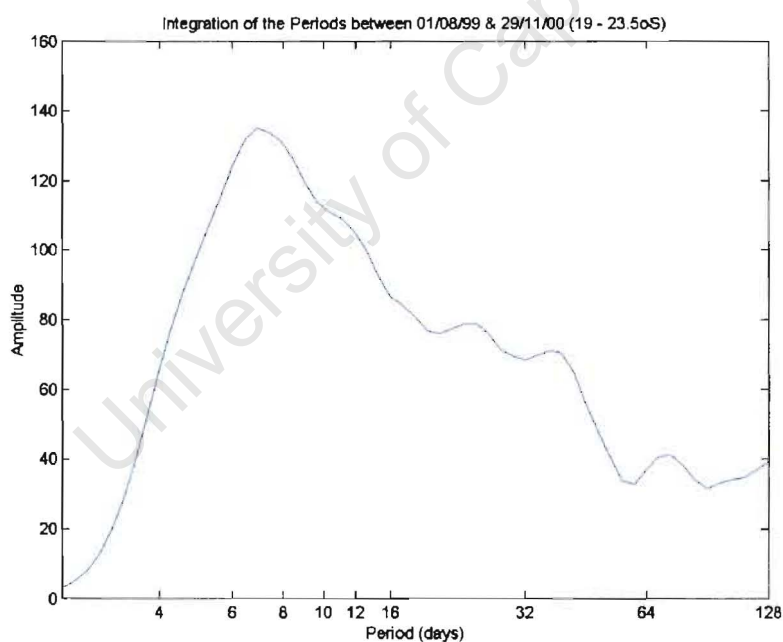
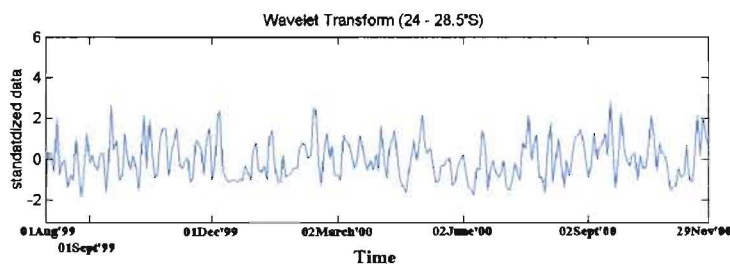


Figure 5.12 Cell 3 (19–23.5°S). (a) Standardized time series of the longshore (v) component for 01 August 1999 – 29 November 2000, (b) the modulus of the wavelet transform, using the Morlet wavelet, of the longshore (v) component for 01 August 1999 – 29 November 2000, and (c) the integrated wavelet transform for 01 August 1999 – 29 November 2000. The contour interval in (b) is 0.2.

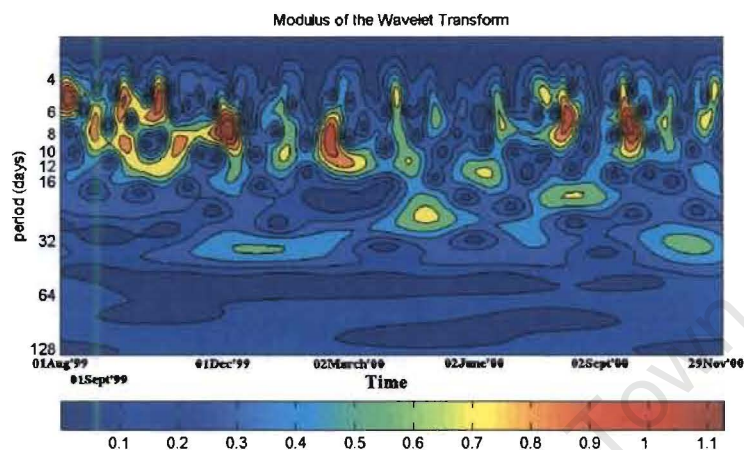
Cells 3, 4 and 5 all contain a band of high frequencies ranging from 4–12 days. This broad peak, which includes the so-called synoptic cycle of 2–8 days (Preston-Whyte and Tyson, 2000), is most pronounced in Cell 3, becoming less so to the south. Authors such as Preston-Whyte and Tyson (1973), Kamstra (1987), Ko and Vincent (1995), and Jury *et al.* (1990a) have all found similar results to those presented here.

Using spectral analysis, Kamstra (1987) investigated the inter-annual variability in the spectra of the daily surface pressure at four stations in the Southern Hemisphere, namely Alexander Bay (1970–1984), Cape Town (1970–1984), Gough Island (1970–1982), and Marion Island (1970–1980). Utilising data that consisted of the 0600 and 1800 GMT surface pressures, corrected to barometer height and reduced to sea level, he concluded that none of the spectra showed any consistent peak through the successive years. The 6 day peak in Cape Town pressures discussed by Preston-Whyte and Tyson (1973) was most clearly defined in 1970, 1975 and 1983, with 1974 containing a 5 day peak. A very distinct 3 day peak occurred in 1973 with lesser peaks in 1971 and 1981. The presence of such a peak in some years thus indicates that the 3 day peak in 1966 need not necessarily have been spurious as was suggested by Preston-Whyte and Tyson (1973). The same conclusion was drawn by Hunter (1984), who presented spectra for pressures at Durban and also found a 3 day peak in 1978, suggesting this to be quite a common period between the passage of coastal lows along the southeast coast. Due to the two day composite nature of the QuikSCAT data analysed here, it is not possible to resolve such a 3 day spectral peak. Jury *et al.* (1990a) investigated the relationship between large-scale Rossby wave structure and the local shelf response of the Benguela Upwelling region. Using daily averaged sea temperatures, shelf currents, sea levels, and hemispheric maps at the 500 hPa level for the period 1983–1985, as well as autocorrelation statistics, the authors were able to isolate a number of cases in which pulse frequencies were related to Rossby wave number and structure in the African sector of the Southern Hemisphere. The dominance of 8000 km (mode 4) waves led to slow oscillations that averaged 23 days in period. Shorter, fast-moving Rossby waves were found to evoke a rapid pulsing of the shelf environment with periods of 9–16 days. These rapid pulsing events were found to be more common in summer, the reason for which is discussed below.

a)



b)



c)

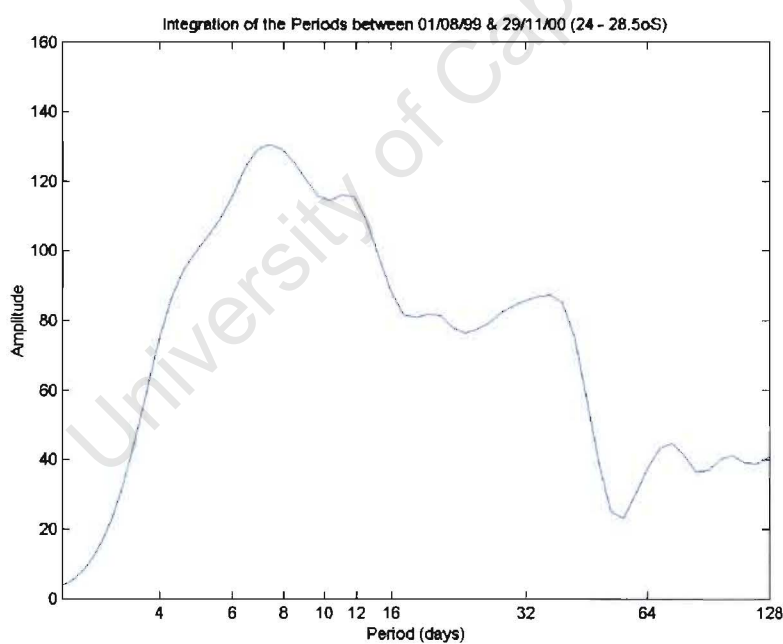
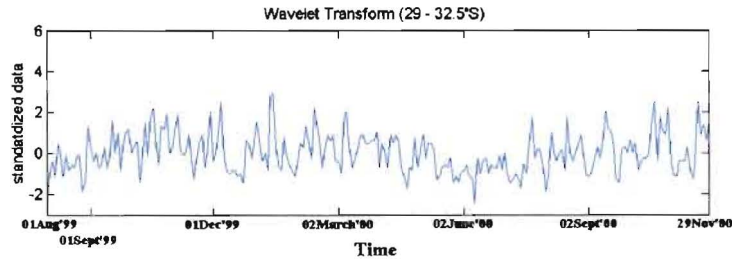


Figure 5.13 Cell 4 (24–28.5°S). (a) Standardized time series of the longshore (v) component for 01 August 1999 – 29 November 2000, (b) the modulus of the wavelet transform, using the Morlet wavelet, of the longshore (v) component for 01 August 1999 – 29 November 2000, and (c) the integrated wavelet transform for 01 August 1999 – 29 November 2000. The contour interval in (b) is 0.1.

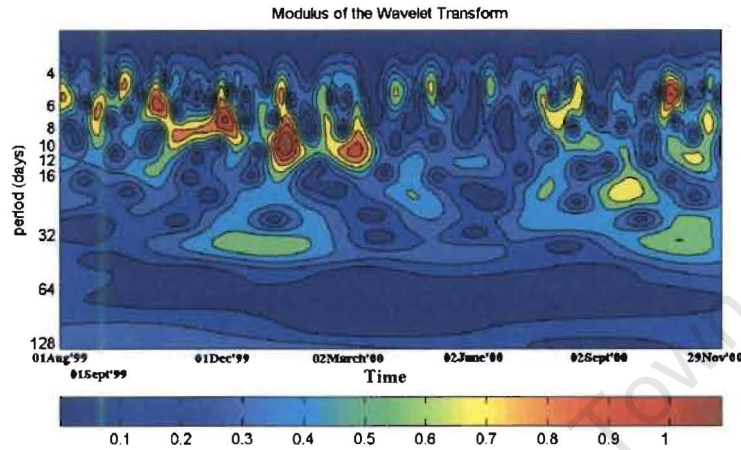
Ko and Vincent (1995) observed the variability in the strength of the zonal flow in the south Pacific jet. During four Southern Hemisphere summers, these flow patterns were found to be quasi-periodic with a period of between 7 and 14 days. While some of these jet strengthenings can be related to tropical convection, a large number appear to be generated through local baroclinic instability, which is considered, along with barotropic instability, by authors such as Frederikson and Webster (1988) and Young and Houben (1989), to be the source for much of the internal intra-seasonal variability over the extra-tropics. The timescale identified by Ko and Vincent (1995) is similar to a dominant timescale of Northern Hemisphere Asian jet fluctuations (Kiladis and Mo, 1998) and happens to be close to the period of the organisation of synoptic scale eddies into baroclinic “wave packets” observed by Lee and Held (1993) during the Southern Hemisphere summer. The wave packets are comprised of a group of baroclinic waves whose amplitudes are strongly modulated in the zonal direction by an envelope of wave activity. The envelopes propagate quite clearly around the globe at 45°S for many cycles, with a mean period of approximately 12 days.

Cells 4–6 show little in the way of a seasonal component when compared to the more northern cells where the majority of the spectral peaks occur between late autumn and early spring. However, a brief reduction in the intensity of events or pulses in wind-stress does appear to occur within the three southern most cells for the months of April through to mid-July, but for the most part a major event occurs approximately once every 30 days. Previous research (Coastal Low Workshop, 1984) has indicated that the coastal low may be subsumed by frontal perturbations near Cape Town in mid-winter and that the SAA may not be able to ridge south of South Africa. Without this anticyclone, enhancement of the equatorward wind-stress preceding the coastal low is less likely. In summer, the coastal lows tend to become more dominant. The retreat of the westerlies ensures a more regular anticyclone bud-off process, while the simultaneous reduction in atmospheric density over the African continent caused by heating contributes to coastal low development and the equatorward wind-stress Jury *et al.* (1990a).

a)



b)



c)

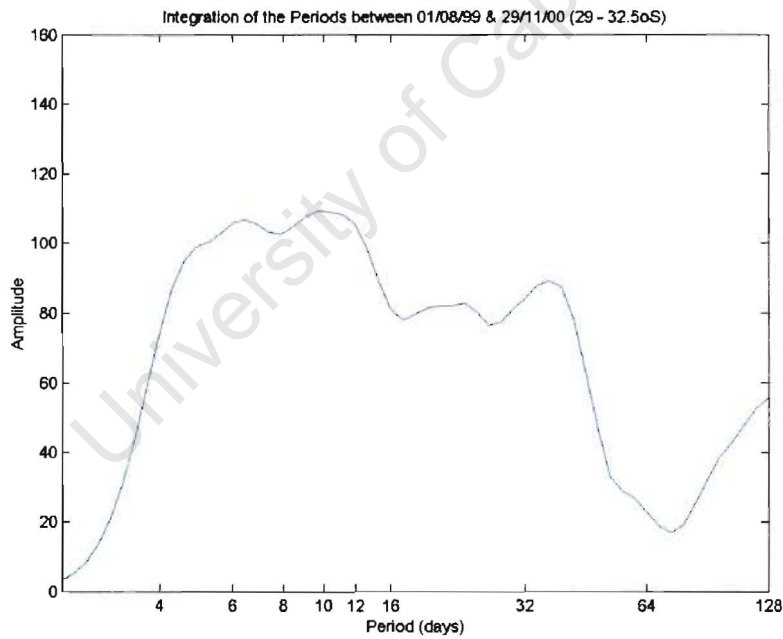
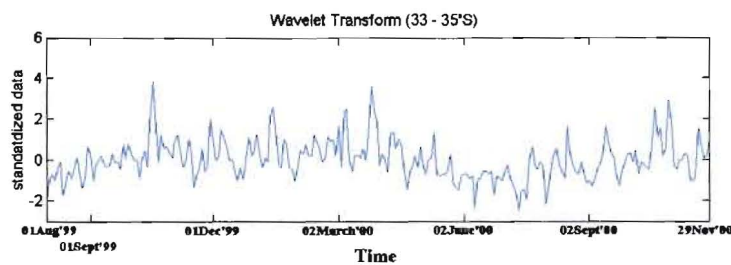


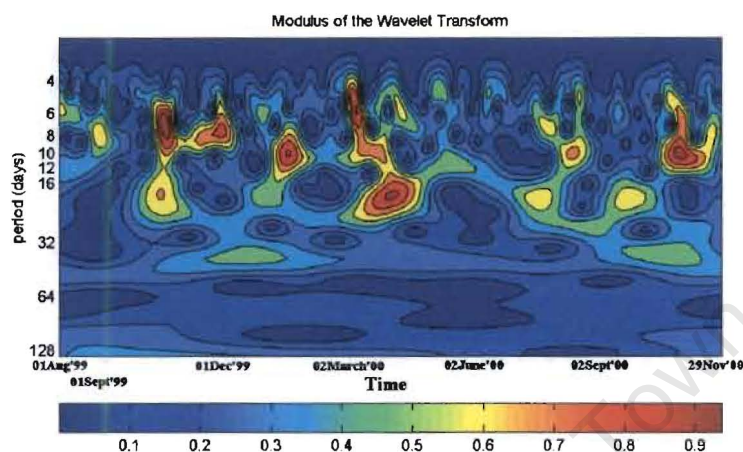
Figure 5.14 Cell 5 (29–32.5°S). (a) Standardized time series of the longshore (v) component for 01 August 1999 – 29 November 2000, (b) the modulus of the wavelet transform, using the Morlet wavelet, of the longshore (v) component for 01 August 1999 – 29 November 2000, and (c) the integrated wavelet transform for 01 August 1999 – 29 November 2000. The contour interval in (b) is 0.1.

At present one of the shortcomings of the wavelet analysis is the lack of accurate statistical significance tests for non-stationary processes, in general, and for wavelet transforms, in particular. The majority of traditional significance tests are derived from the assumption of identical repeated cycles for non-stationary processes and are therefore inappropriate for wavelet transforms (Lau and Weng, 1995). An alternative is to use Monte Carlo methods. However, Monte Carlo methods are also inadequate in the context of non-linear systems, whereby one is attempting to localise frequencies that may be part of a hierarchy of frequency structures related to the fundamental driving frequencies. Some of the higher harmonics have small amplitudes and may be difficult to distinguish from noise. Yet their presence, in conjunction with the fundamental frequencies, is physically important but not necessarily statistically significant in the traditional sense (Lau and Weng, 1995). In this regard, the results of the wavelet analysis presented above should be regarded with caution, pending the development of an accurate statistical significance test. The formulation of such a test is an important research topic and, as such, is beyond the scope of this dissertation. That said, Torrence and Compo (1998) have argued in defence of using the chi-square test on non-stationary data. These authors listed several reasons for this argument, including the fact that many wavelet transforms of real data appear similar to transforms of red-noise processes. They thus felt that it is difficult to argue that large variations in wavelet power imply non-stationarity.

a)



b)



c)

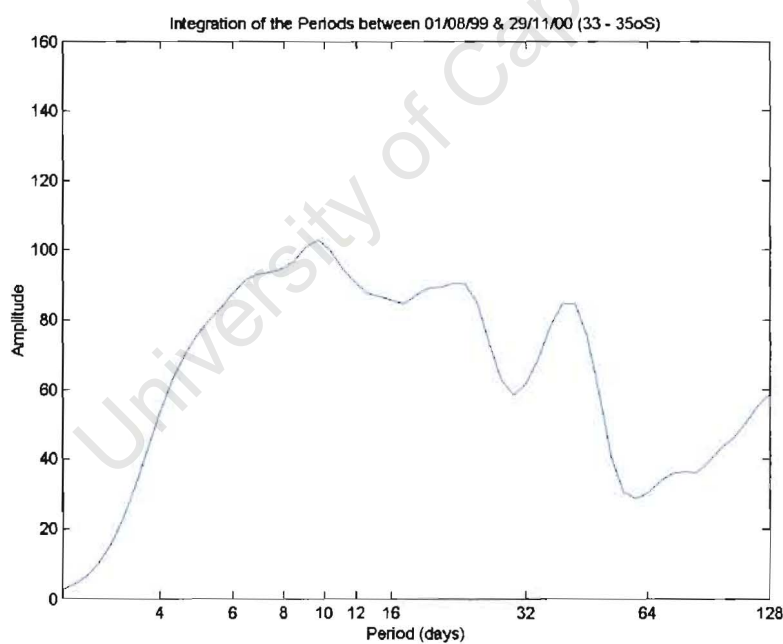


Figure 5.15 Cell 6 (33–35°S). (a) Standardized time series of the longshore (v) component for 01 August 1999 – 29 November 2000, (b) the modulus of the wavelet transform, using the Morlet wavelet, of the longshore (v) component for 01 August 1999 – 29 November 2000, and (c) the integrated wavelet transform for 01 August 1999 – 29 November 2000. The contour interval in (b) is 0.1.

CASE STUDY OF THE 11-20 FEBRUARY 2000 WIND EVENT

There is a reasonably good understanding of the winds associated with the ridging of the SAA south of South Africa. However, little is known of the pulsing wind events that occur between 24 and 28°S. Generally speaking, these central Benguela wind events are associated with localized (24 – 30°S) and intense wind-stress. The 11 – 20 February 2000 wind event was identified as being an intense ($\sim 0.4 \text{ N.m}^{-2}$) wind event from the two latitude (10 – 35°S) vs. time (01 August 1999 – 29 November 2000) Hövmüller (Figures 5.16 & 5.17) plots of u and v wind-stress components along a transect 200km offshore along the west coast of southern Africa. A description of the meso-scale atmospheric state as well as a qualitative look at oceanic response to this forcing is given below. The distance of 200km offshore was chosen due to the topographical influence on both the magnitude and direction of wind-stress close to the coast (Johnson, 2001).

Geographic Setting

The curved anticyclonic airflow over the Benguela region is associated with the SAA. Strong heating of the continent in summer leads to relatively lower pressures in the interior, viz. a heat low. During the summer, when this heat low contrasts sharply with the SAA, the pronounced zonal pressure gradients produce strong, upwelling favourable, southerly and south-easterly winds along the south-west coast of Africa. These winds are laterally confined by both the topography of the continental escarpment and the thermal barrier of the heat low (Kamstra, 1985). The SAA is maintained throughout the year, retreating between 4 and 6° to the north during winter. On the scale of days, however, the SAA may ridge eastward and to the south of the continent. Extended ridging of the SAA may lead to the budding off of eastward tracking migratory anticyclones. These anticyclones drift eastwards, south of South Africa and into the Indian Ocean prior to being subsumed into the South Indian High. In so doing, they may strongly affect the southern Benguela by producing, for example, strong south-easterly winds over the Western Cape (Preston-Whyte and Tyson, 2000).

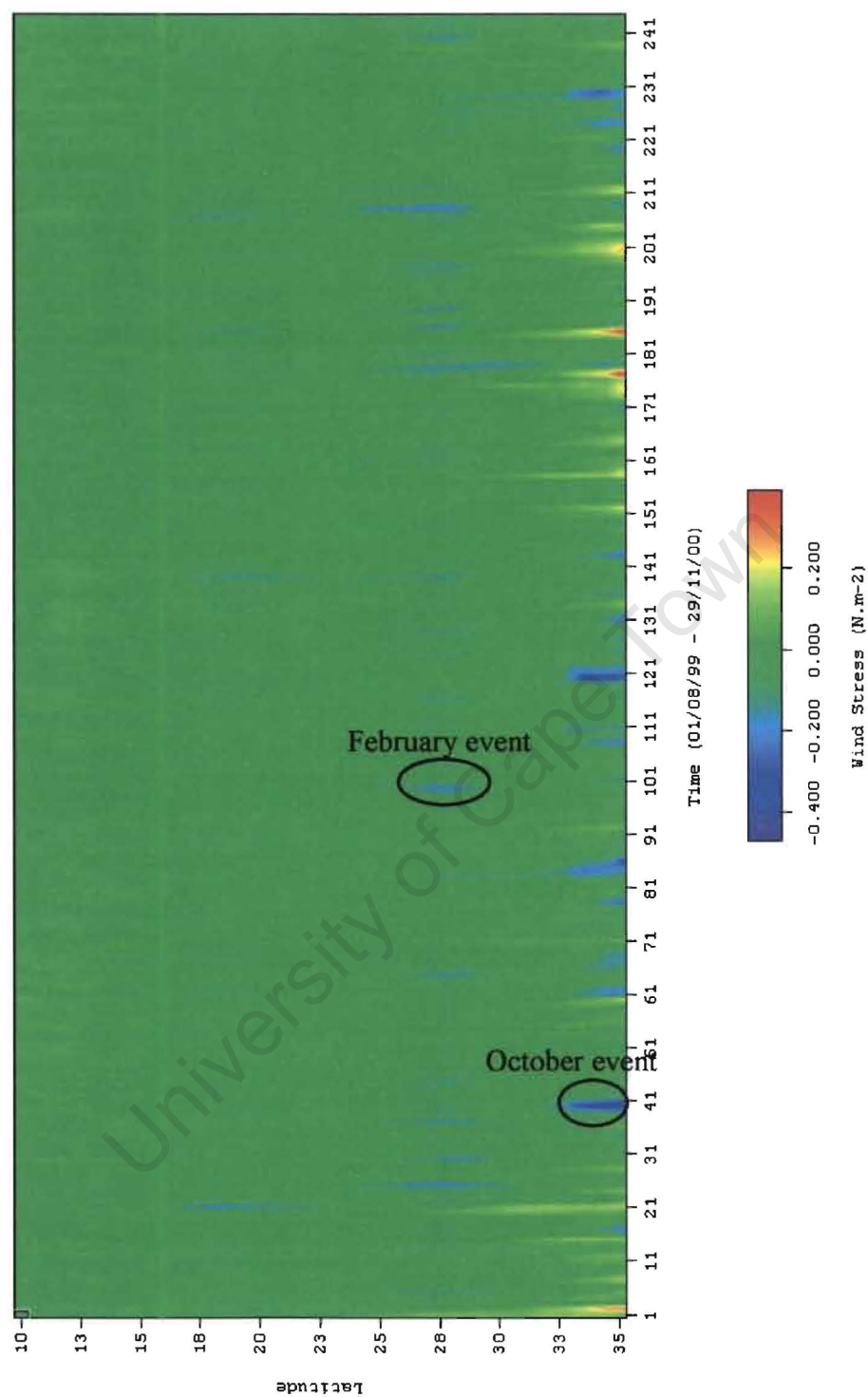
u-component

Figure 5.16 Hövmuller Plot of Latitude (10–35°S) vs. Time (01 August 1999 – 29 November 2000) for the, 200km offshore, *u* wind-stress (N.m⁻²) component. The October 1999 as well as the February 2000 wind events are highlighted.

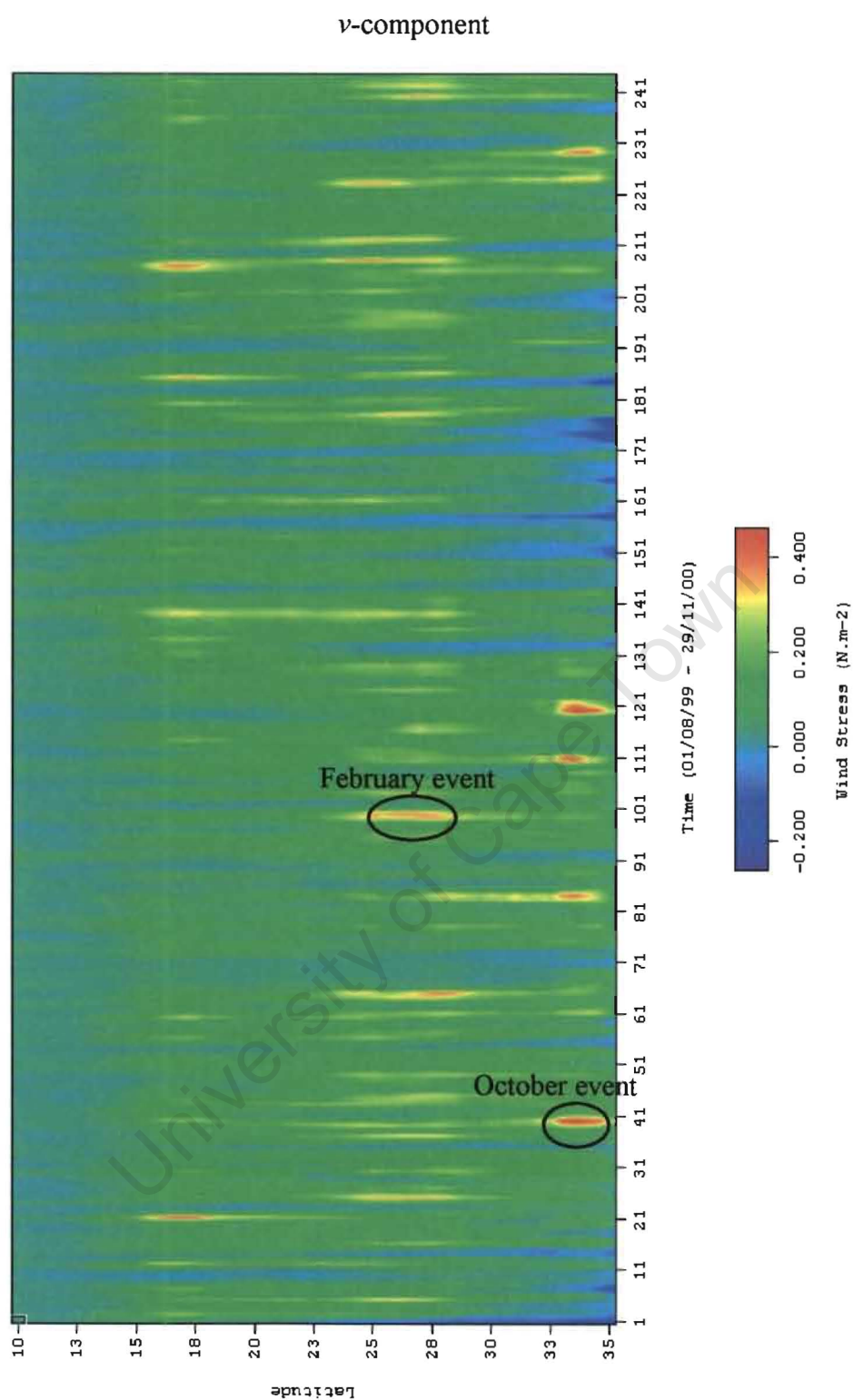


Figure 5.17 Hövmuller Plot of Latitude (10–35°S) vs. Time (01 August 1999 – 29 November 2000) for the, 200km offshore, v wind-stress (N.m⁻²) component. The October 1999 as well as the February 2000 wind events are highlighted.

The area under investigation is located off the west coast of southern Africa (25–30°S and 12–17°E). This low, relatively featureless semi-desert is characterised by a relatively straight coastline with few major capes (e.g. Lüderitz, ~26.5°S) and embayments (e.g. Walvis Bay, ~23°S). The region contains only two river systems of any significance, namely the Orange and Cunene Rivers. Because of the arid nature of the region, the fresh water input into the coastal ocean is relatively small, with exceptions occurring during times of summer flooding in the interior (Shillington, 1998).

The most marked feature of the bathymetry of the South Atlantic is the Mid-Atlantic ridge. Secondary ridges segment the eastern trough into the Angola, Cape, and Agulhas Basins. The most significant of these secondary ridges is the Walvis ridge (Figure 2.1). This ridge runs along a northeast-southwest axis and connects the Mid-Atlantic ridge to the African continent. The Walvis ridge serves to separate the Angola Basin from the Cape Basin.

Off of Cape Town and Cape Columbine the continental shelf is relatively narrow with a width of roughly 40 km wide. West of the Orange River the shelf widens to approximately 180 km, before once again narrowing to approximately 30 km north of 17°S (Shannon and Nelson, 1996). Double shelf breaks are common to the southwest coast of Africa. An example of one of these double shelf breaks is found off the coast at Walvis Bay. Here there are inner and outer breaks beginning at depths of about 140 and 400m respectively (Shannon and Nelson, 1996).

The features mentioned above are important in the local forcing of any one particular upwelling regime. For example, capes and promontories act to trigger instabilities within upwelling jets, which may in turn result in filament formation. Localised or intensified coastal upwelling can also often be attributed to particular promontories thus making them important with respect to spatial variability within upwelling regimes (Hill *et al.*, 1998).

The 11 – 20 February 2000 Wind Event

Charts of sea level pressure and 500hPa geopotential heights, as derived from the NCEP/NCAR reanalysis dataset (Kalnay, 1996), are presented in Figure 5.20a – j. Figure 5.21a – c shows the oceanic SST response to this particular wind event, within the northern Benguela (17 – 29°S). Each image is a two-day composite such that 14 February 2000 is a composite of 13 and 14 February. The images, which have a spatial resolution of 1 km, are derived from the AVHRR instrument flown on the NOAA series of satellite.

The QuikSCAT data, presented in Figure 5.18a-e, reveal that upwelling favourable winds began to strengthen on 11 February and lasted for 10 days, reaching peak intensity 300km off the coast, at ~27°S, between the 14 and 16 February 2000 with the average wind-stress exceeding 0.35 N.m^{-2} . Towards the end of the event, a gradual reduction in wind-stress propagated down the coast in a poleward direction. The winds were forced by the interaction between trapped coastal lows situated off the west coast of southern Africa, eastward propagating mid-latitude cyclones, the interior heat low, and the ridging of the SAA south of South Africa (Figure 5.19a – j). The horizontal structure of the wind field (Figure 5.18a–e) derived from the QuikSCAT scatterometer data changed considerably during this period.

The QuikSCAT two-day composite for 11 February 2000 (Figure 5.18a), which is a composite of the wind-stress field for 11 and 12 February 2000, shows a relatively uniform wind field over the entire Benguela Upwelling System. To the south, the wind-stress has a strong southerly component, becoming more south-easterly towards the north. A stress maximum of 0.15 N.m^{-2} is situated approximately 200 km off the coast at 27.5°S. Secondary wind-stress maxima are situated to the north at 23.5°S and 17°S, where wind-stress of 0.125 and 0.1 N.m^{-2} , respectively, are found. Even though there is a tendency for the QuikSCAT two-day composites to smooth out meso-scale atmospheric features, the onshore flow associated with the trapped coastal low situated to the north of 25°S (Figure 5.19a & b) is evident in Figure 5.18a.

The meso-scale structure of SST for 13 and 14 February 2000 (Figure 5.21a) shows a contiguous alongshore upwelling band between 25.5 and 29°S with no sign of active upwelling. This band of 14-15°C water has a greater offshore expression, of approximately 100 km, between 27 and 25.5°S. South of this, the 14-15°C water extends approximately 50 km offshore.

For the period of 13 and 14 February 2000, there appears to be a slight north-easterly shift in the SAA (Figure 5.19c & d). In addition to this, a westward shift in the interior heat low, which includes the possible absorption of the trapped coastal low, as well as the ridge of high pressure extending along the south coast of South Africa all result in an increase in the pressure gradient between 26 and 32°S (Figure 5.19d). The enhanced wind-stress associated with the strengthening of this pressure gradient can be seen in Figure 5.18b where a stress maximum of 0.375 N.m^{-2} is situated approximately 250 km off the coast at 27.5°S. Associated with this wind-stress maximum are three areas of strong cyclonic (negative) wind-stress curl. The first area stretches in a north-westerly direction along the coast from 27°S in the north to 31°S in the south. The wind-stress curl reaches a maximum of -1.8 N.m^{-3} in this particular region. The second area is situated between 22 and 25°S. This maximum moves slightly offshore towards the south, and contains wind-stress curl values of up to -1.4 N.m^{-3} . The third area shows far less spatial extent than the two previously mentioned areas and is associated with the Lüderitz (26.5°S) upwelling cell (Lutjeharms and Meeuwis, 1987) where the wind-stress curl reaches a maximum of -0.6 N.m^{-3} . Secondary maxima are also found off the Cape Peninsula (34.5°S) and to the north of Cape Frio at approximately 16°S.

The SAA begins to ridge south of South Africa over the two-day period of 15 and 16 February 2000 (Figure 5.19e & f). The relatively strong pressure gradient seen on 14 February (Figure 5.19d) is maintained through 15 February, but has all but vanished by 16 February. 16 February also sees the interior heat low extending southwards in the Northern Cape region. Offshore airflow from the escarpment, as indicated by the wind direction at Windhoek, has assisted in formation of another trapped coastal low at 25°S (Reason and Jury, 1990). Figure 5.18c shows that the wind-stress maximum that is visible in Figure 5.18b and is associated with the relatively strong pressure gradient between 25 and 30°S has moved slightly north to be centred at 26°S. In

addition to this, the stress maximum (0.375 N.m^{-2}) is situated slightly closer inshore in comparison to 13 February 2000 (Figure 5.18b). A secondary wind-stress maximum (0.1 N.m^{-2}) is situated just off the coast at 17°S . The wind-stress curl maximum, situated at 26°S , has intensified to -1.4 N.m^{-3} . To the north, the wind-stress curl maximum situated between 22 and 25°S has moved approximately 100 km offshore, while the area between 27 and 31°S has divided into two distinct cells. The first cell is located at 29°S , while the second cell is located further south at 30.5°S . Both cells contain wind-stress curl of the order of -0.8 N.m^{-3} .

Figure 5.21b shows clear evidence of active upwelling, in response to the significant increase in wind-stress over the period of 14 – 16 February 2000, with the alongshore upwelling band more than doubling its offshore extent to almost 250 km offshore north of Lüderitz (26.5°S). North of Mercury Island ($\sim 25.5^\circ\text{S}$), this band begins to move offshore in a north-north-westerly direction. An SST maximum (17°C) is found inshore of this band at Meob Bay (24.5°S).

Figures 5.19g & h (17 and 18 February 2000) show the ridging south of South Africa and absorption by the South Indian High Pressure of the migratory anticyclone that budded off from the SAA. Also visible is the marked decrease in the pressure gradient off the west coast of southern Africa. This decrease is reflected in the wind-stress field depicted in Figure 5.18d. The wind-stress maximum (0.2 N.m^{-2}) has shifted south in sympathy to the ridging SAA and is now situated approximately 250 km offshore at 30°S . A second maximum (0.2 N.m^{-2}) lies approximately 400 km off of Cape Columbine (33°S). Associated with this decrease in wind-stress is a significant decrease in wind-stress curl. The strongest wind-stress curl (0.8 N.m^{-3}) occurs to the south-west of the Cape Peninsula (34.5°S).

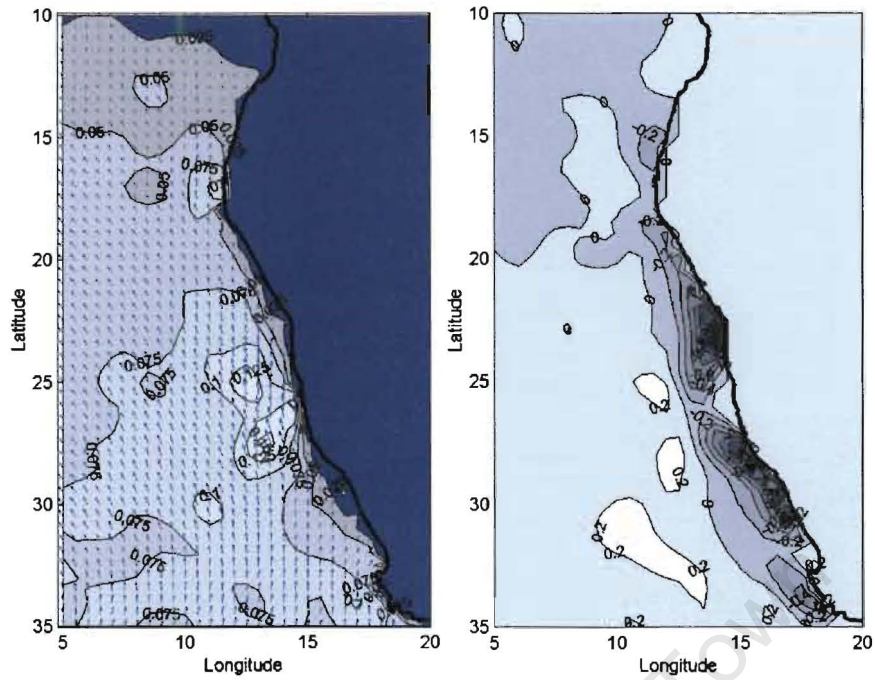
19 February 2000 (Figure 5.18e) shows weak ($<0.025 \text{ N.m}^{-2}$) southerly to south-easterly wind-stress over much of the study area. A wind-stress maximum (0.1 N.m^{-2}) is centred approximately 500 km offshore at $32.5^\circ\text{S} : 13^\circ\text{E}$. Once again, the cyclonic rotation associated with the trapped coastal low (Figure 5.19i) is evident in Figure 5.18e at approximately 29°S . By 21 February 2000 (not shown), the equatorward wind-stress had all but collapsed. This is reflected in Figure 5.21c (21 February 2000) where the previously coherent pattern of active upwelling (Figure 5.21b) has broken

down into a coalescence of warm eddies (17°C), cold patches (14°C) and relatively smaller filaments.

The development and eventual breakdown of offshore SST frontal structures in response to upwelling favourable wind events has a significant impact on marine biological processes and, in turn, on the fishing industry. A number of investigations in regions other than the southeast Atlantic Ocean have revealed that localised maxima in productivity biomass are found at the borders of, or just outside, upwelling fronts (e.g. Dengler, 1985). Simpson *et al.* (1979) concluded that, not only is the standing crop of phytoplankton several times greater at fronts, but in many cases it is considerably healthier. In the upwelling cells off southern Africa, it has furthermore been shown by Shannon *et al.* (1983) that phytoplankton blooms occur offshore and not in the centre of active upwelling cells. Lutjeharms and Stockton (1987) have suggested that an extensive field of upwelling filaments, with an abundance of long frontal lines and eddies, could be an area of considerable primary production in its own right. They further suggested that a substantial or even a major portion of the total primary production attributable to upwelling could take place in the filamentous region rather than in the coastal upwelling regime.

a)

11 February 2000



b)

13 February 2000

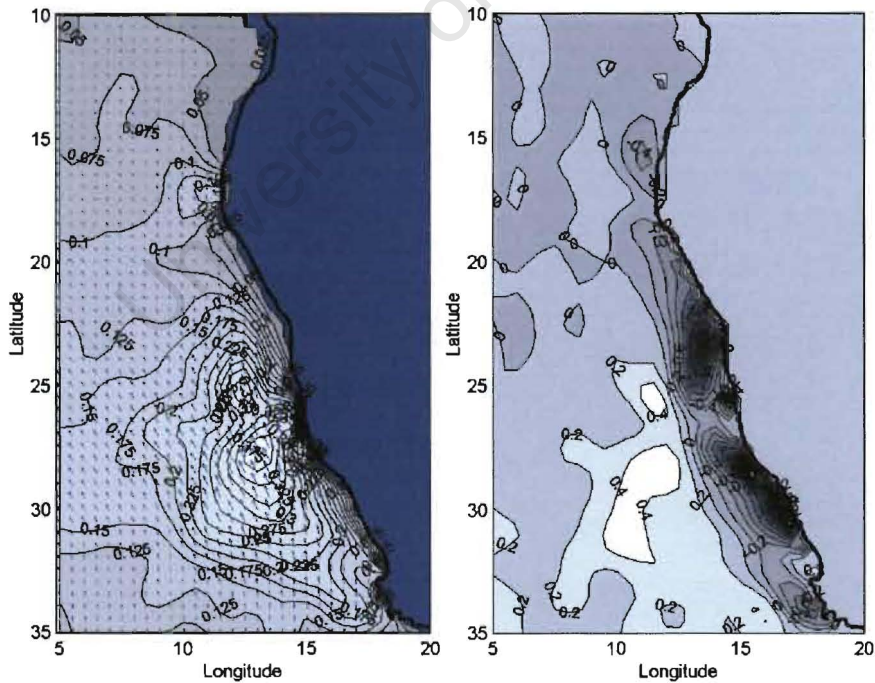
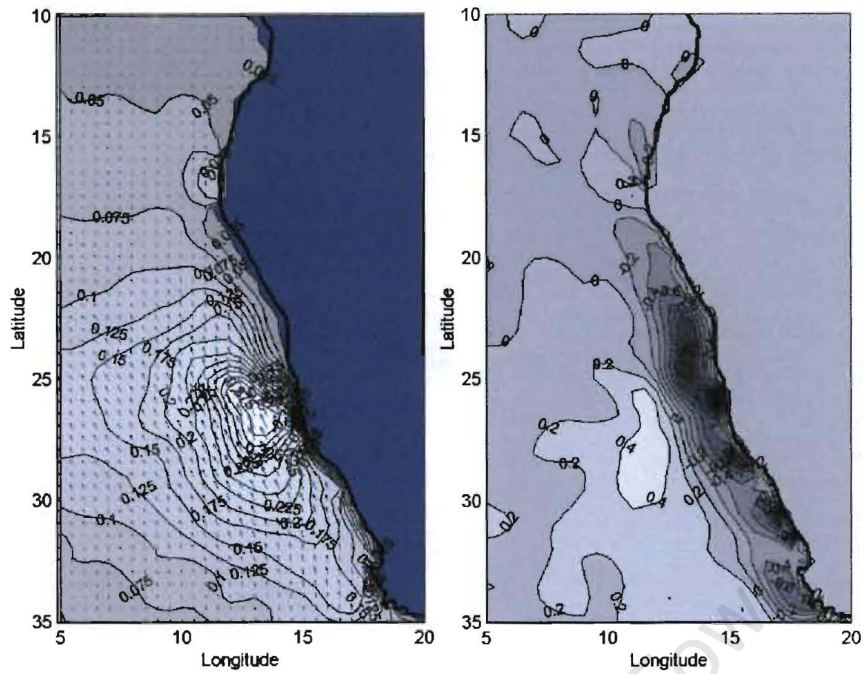


Figure 5.18 Two-day composites of the Benguela Upwelling Regime: surface wind-stress ($N.m^{-2}$) and wind-stress curl ($N.m^{-3}$) distribution. The contour intervals are $0.025 N.m^{-2}$ and $0.2 N.m^{-3}$ respectively. (a) 11 February 2000 (a composite of 11 & 12 February 2000), (b) 13 February 2000 (a composite of 13 & 14 February 2000).

c)

15 February 2000



d)

17 February 2000

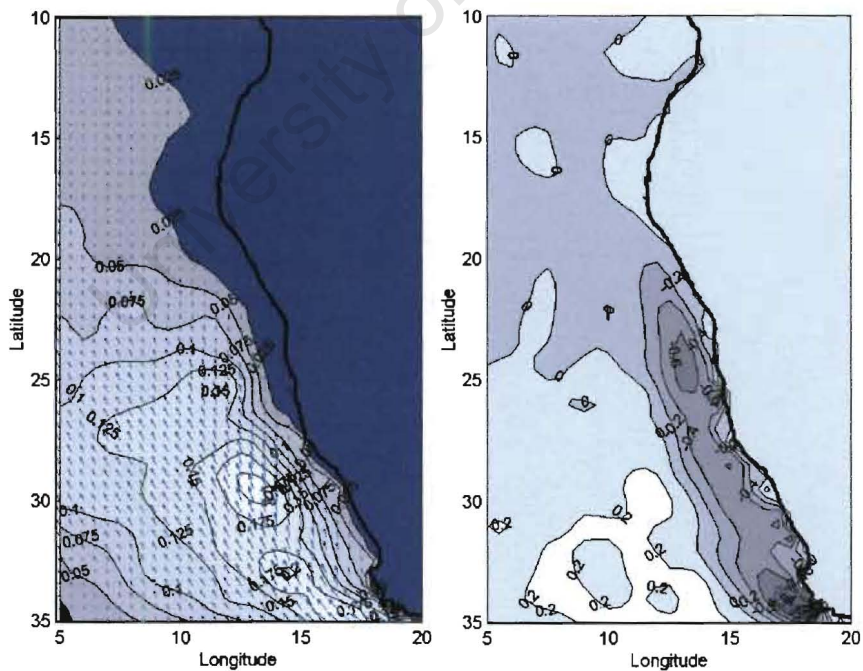


Figure 5.18 (Continued) (c) 15 February 2000 (a composite of 15 & 16 February 2000), (d) 17 February 2000 (a composite of 17 & 18 February 2000).

e)

19 February 2000

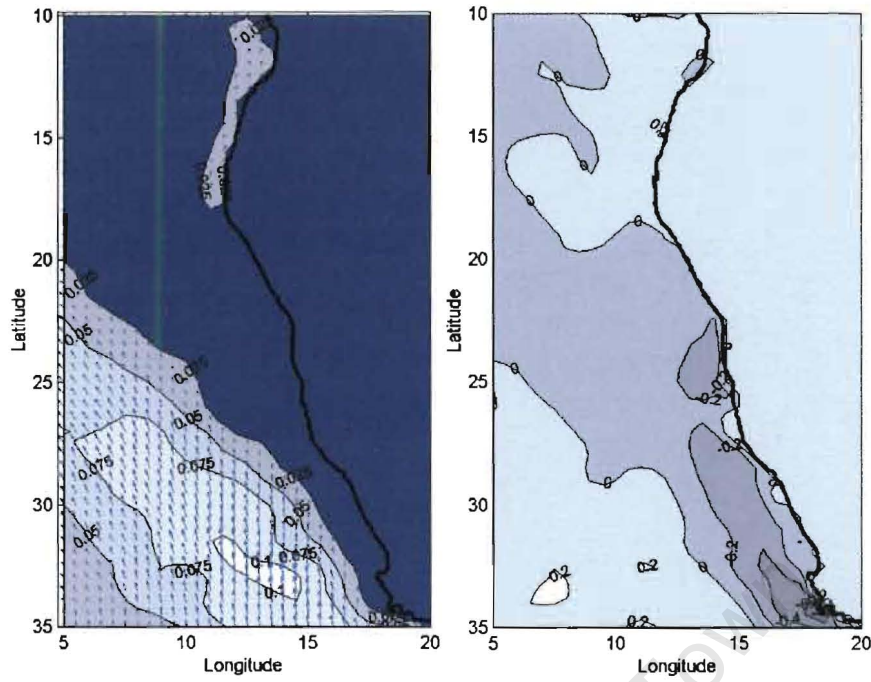


Figure 5.18 (Continued) (e) 19 February 2000 (a composite of 19 & 20 February 2000).

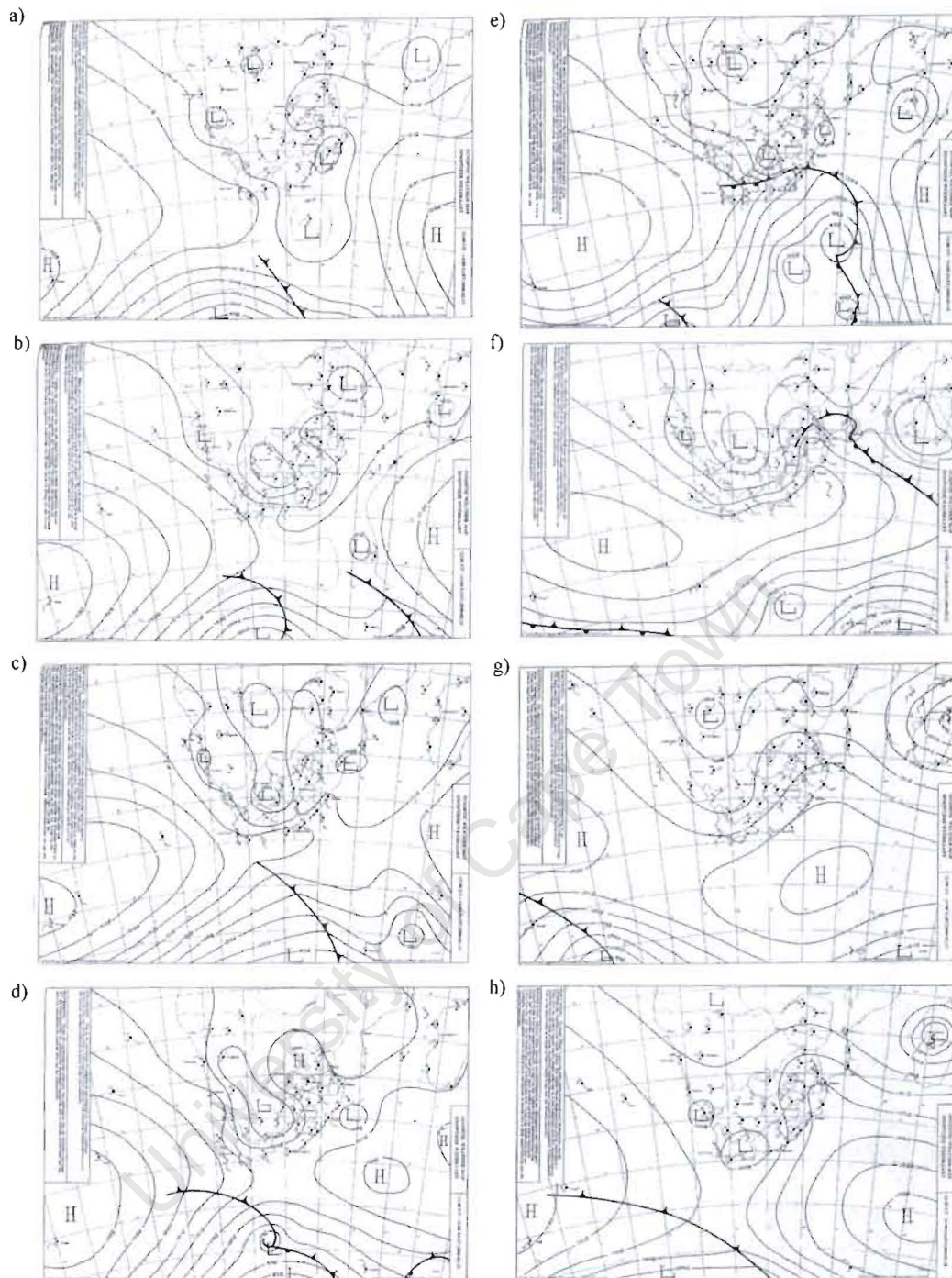


Figure 5.19 Daily synoptic weather maps, produced by the South African Weather Service, for the period 11 – 20 February 2000.

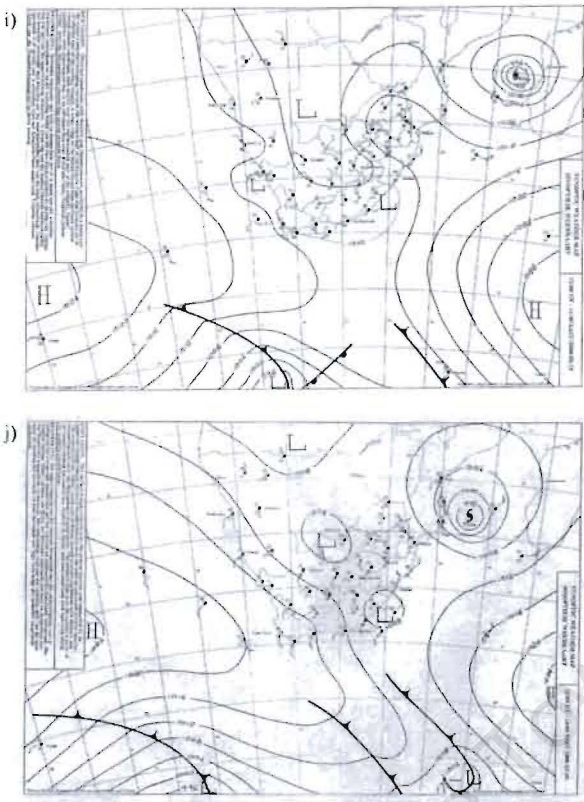


Figure 5.19 Daily synoptic weather maps, produced by the South African Weather Service, for the period 11 – 20 February 2000.

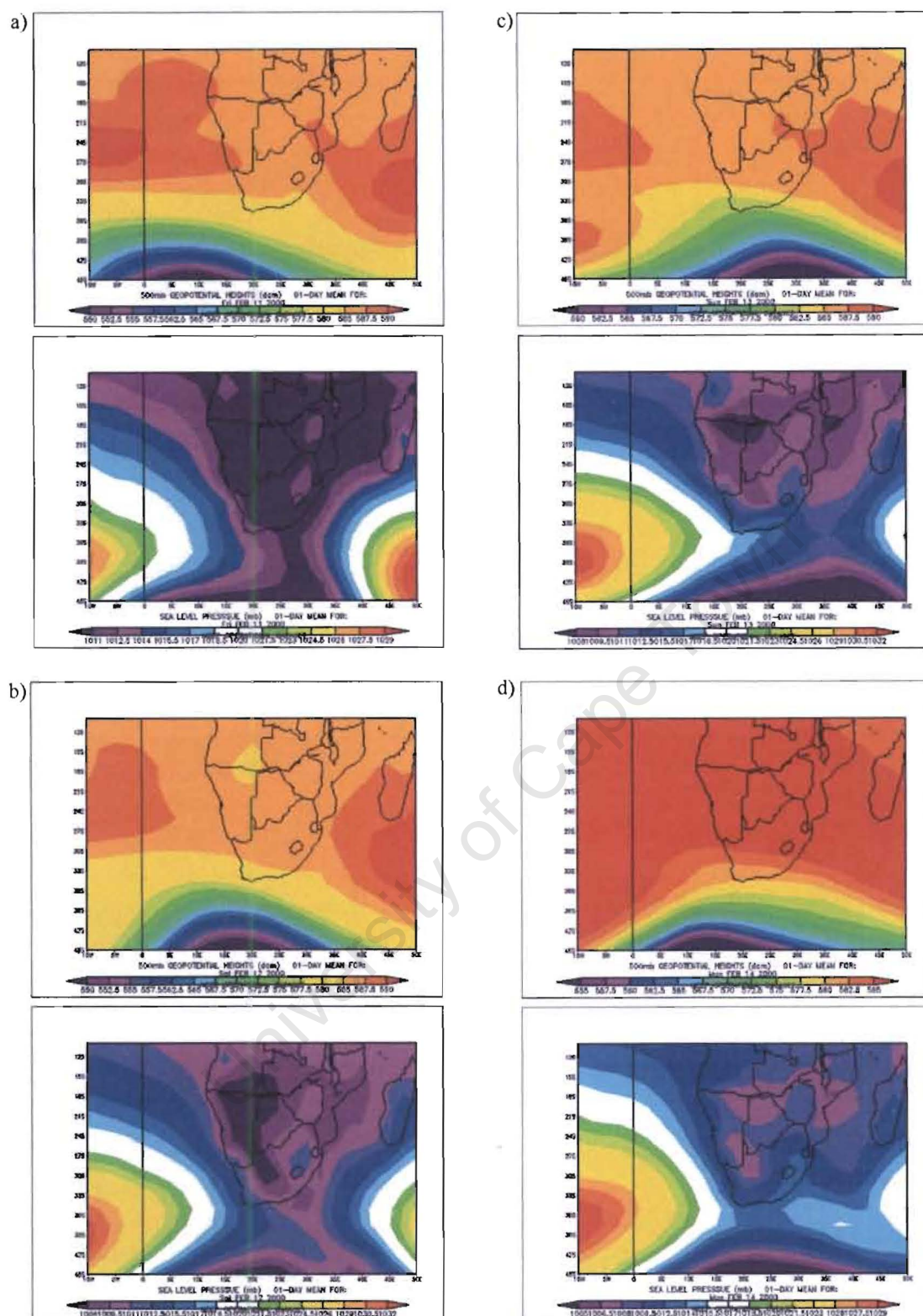


Figure 5.20 Daily synoptic weather charts of sea level pressure (SLP) and 500hPa geopotential heights, as derived from the NCEP/NCAR reanalysis dataset², for the period 11 – 14 February 2000.

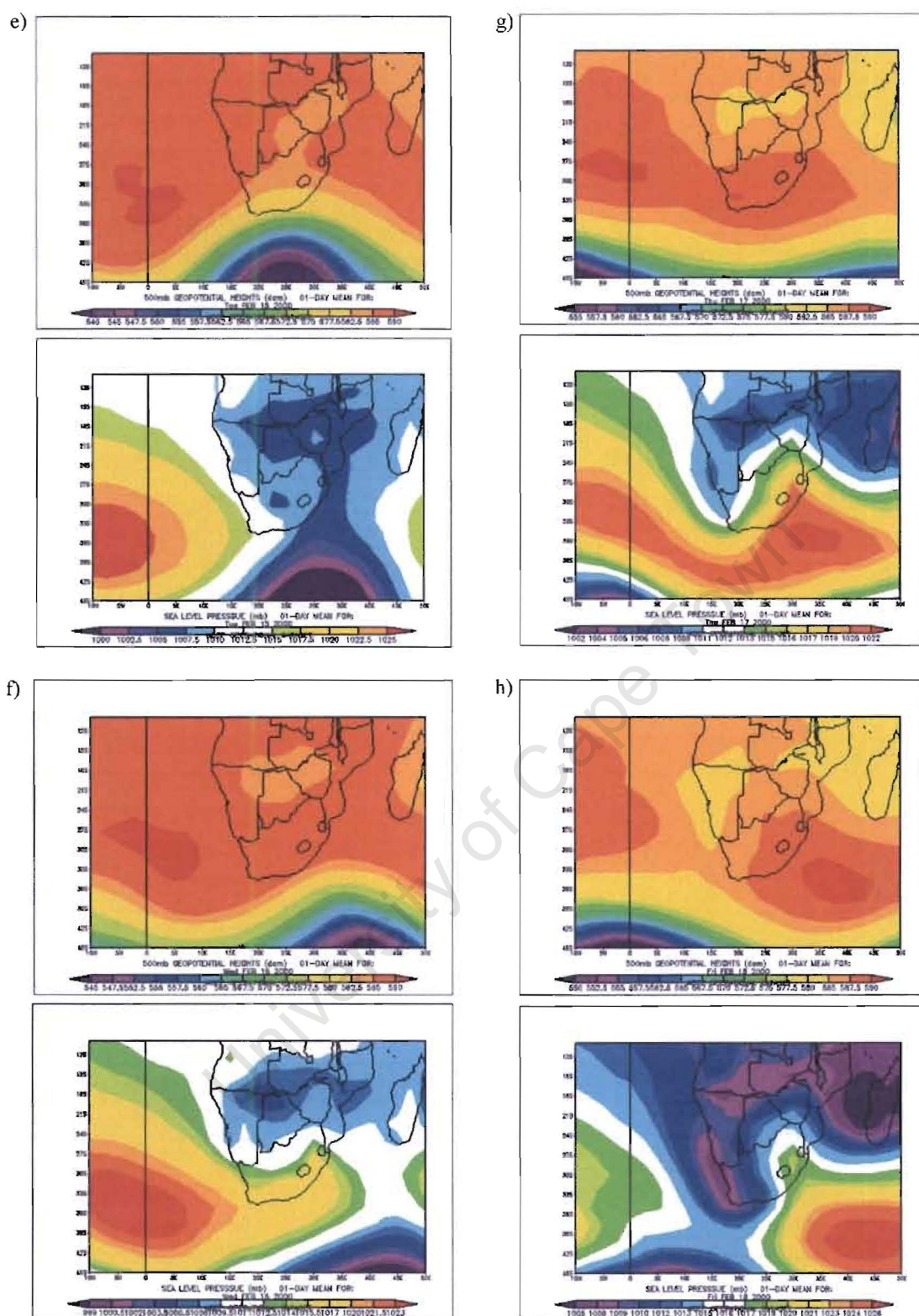


Figure 5.20 (Continued) for the period 15 – 18 February 2000.

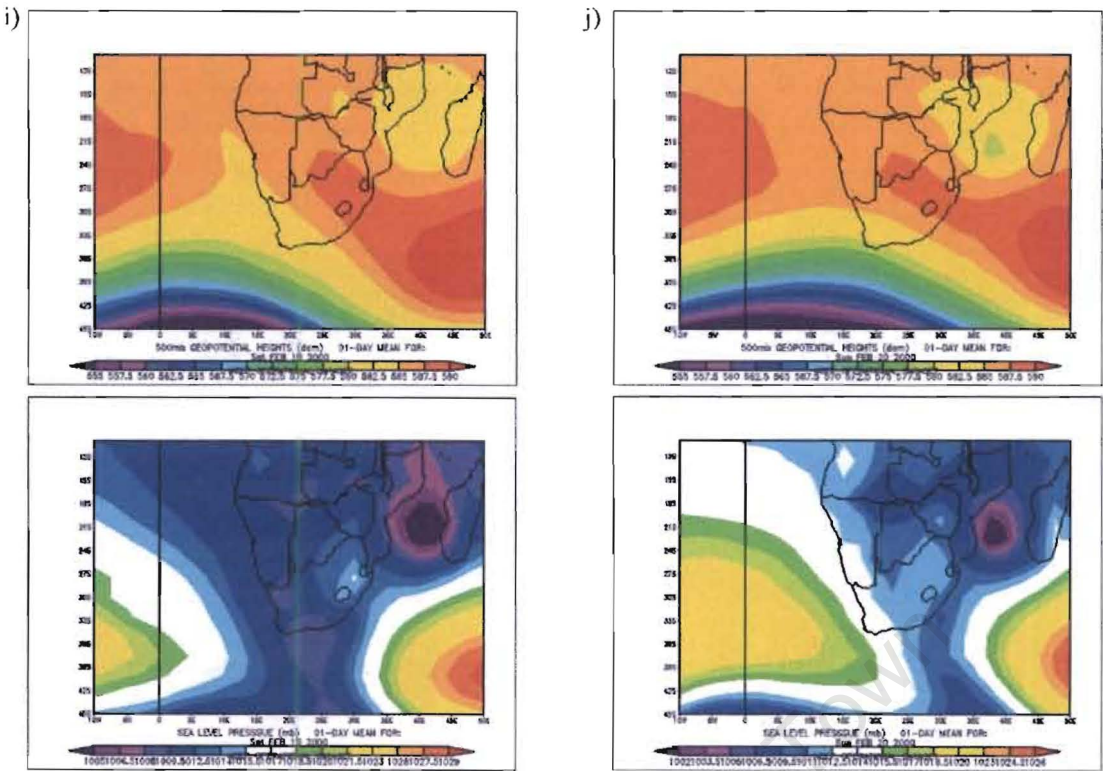


Figure 5.20 (Continued) for the period 19 – 20 February 2000.

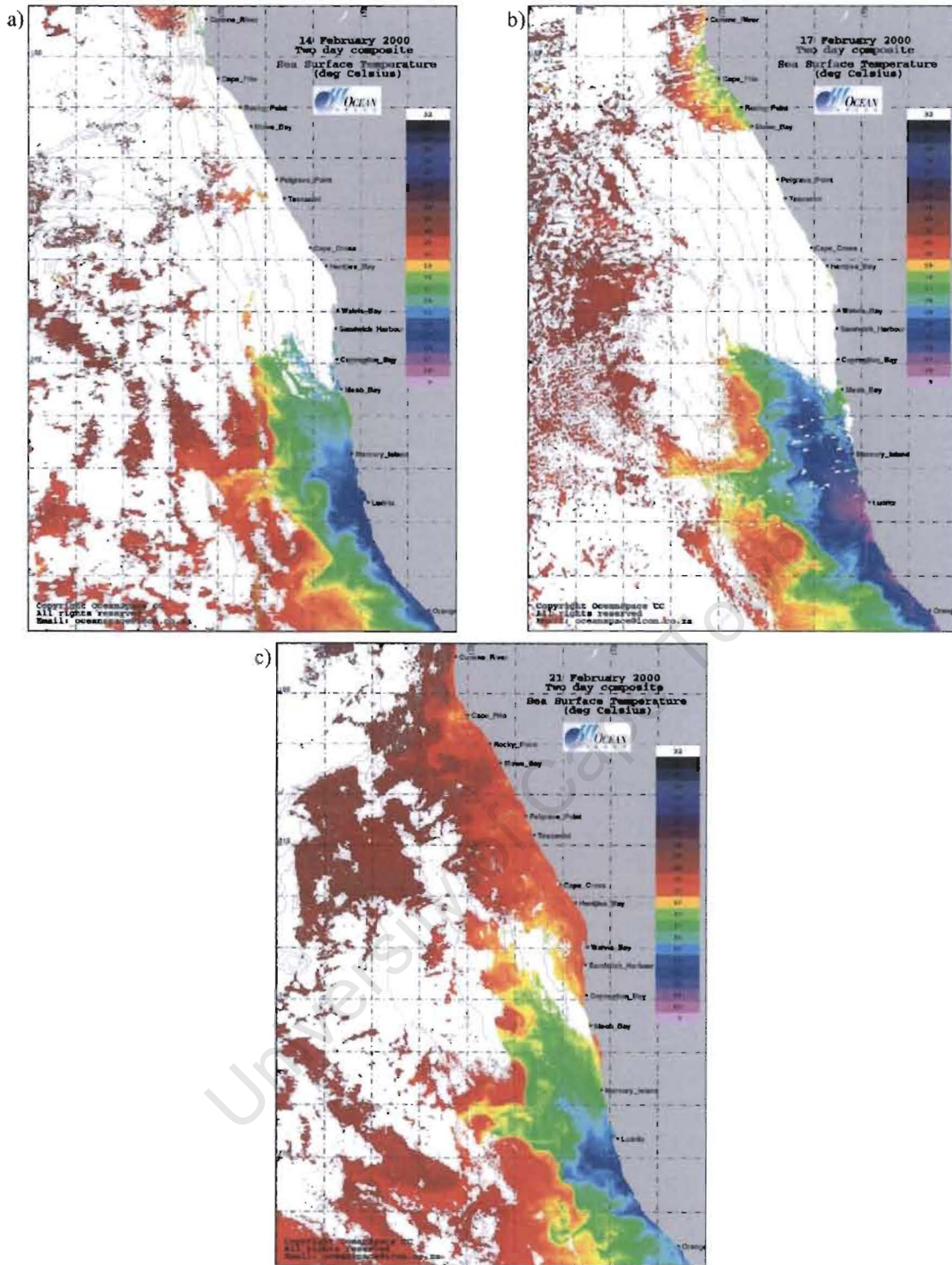


Figure 5.21 Two-day composites of SST within the Northern Benguela Upwelling Regime (17 – 29°S). The contour interval is 1°C. (a) 14 February 2000 (a composite of 13 & 14 February 2000), (b) 17 February 2000 (a composite of 16 & 17 February 2000), and (c) 21 February 2000 (a composite of 20 & 21 February 2000).

CHAPTER 6

CONCLUSIONS

The Benguela Current, which is the eastern boundary current of the South Atlantic subtropical gyre, is one of four major eastern boundary current regimes of the World Ocean. Dengler (1985) estimated that, although the combined surface for all four upwelling systems accounts for only 0.1% of the total surface area of the world oceans, these regions account for almost 30% of the world's total fish catches. The Benguela Ecosystem is thus extremely important, economically speaking, to the countries of southern Africa, and, as a result, a large amount of marine research has been directed towards the analysis and understanding of the ecosystem and its variability.

Wind-stress variability is a major driving force behind ocean variability over a wide spectrum of time and space scales. Changes in longshore wind are important for barotropic processes on the shelf, surface drift and Ekman transport, stratification and for upwelling and frontal dynamics in general. Changes in the large-scale wind-stress field have likewise been linked to several large scale advective processes which have been identified as important contributors to the variability in the oceans around Southern Africa, namely Benguela Niños (Shannon *et al.*, 1986), intrusions of Agulhas water from the retroflexion into the South Atlantic (Shannon *et al.*, 1990a) and the El Niño Southern Oscillation (Reason *et al.*, 2000).

The intention of this dissertation has been to utilise a newly available high-resolution satellite-derived data set to contribute to a better understanding of the spatial and temporal variability of wind-stress and wind-stress curl over the Benguela Upwelling System at synoptic to intra-seasonal time scales. The objectives for this dissertation were two fold:

- 1) To investigate the variability of wind-stress and wind-stress curl over the Benguela upwelling system during the period (01 August 1999 to the 29 November 2000) that these QuikSCAT scatterometer data were available. This QuikSCAT derived data set consisted of the u and v wind-stress components, as well as the

derivative wind fields, namely curl and divergence. These data were sub sampled to a spatial domain of 10°S to 35°S and 5°E to 20°E, and were smoothed temporally to 2-day composites and spatially to a one-half degree resolution.

2) To investigate both the meteorological characteristics and the oceanic response to an intense, upwelling-favourable wind event that occurred from 11-20 February 2000.

In order to place the QuikSCAT data (01 August 1999 - 29 November 2000) into a regional and global perspective, they were compared to the climatologies presented by Kamstra (1985) and Bakun and Nelson (1991), as well as to the long-term climatology (1968 -1996) of the surface vector wind speed field off the coast of southern Africa, as derived from the 2.5° resolution NCEP/NCAR reanalysis dataset.

The annual averaged QuikSCAT data (01 August 1999 - 01 August 2000) showed south-easterly wind-stress throughout the Benguela region. Wind-stress maxima were found to occur at approximately 25 and 17°S. These findings relate well to those of Parrish *et al.* (1983), who found the locations of maximum offshore drift to be located at similar latitudes. Despite the global climatic context, which was characterised by a pronounced La Niña event during the time period considered here, a number of broad scale similarities were found between the QuikSCAT and the long-term NCEP/NCAR climatology (1968 -1996) data sets. The general agreement between the QuikSCAT mean wind-stress and other data allows one to have confidence in using this scatterometer data to investigate details of spatial and temporal variability over the Benguela System.

During summer, wind-stress maxima were found at approximately 17, 29 and 34°S. These maxima were found to strengthen in late summer. The intensification of the coastal alongshore winds could be a result of thermal forcing by the land-ocean temperature contrast, as observed by Chao (1985) off the Californian coast. The seasonal northward migration of the South Atlantic Anticyclone (SAA) became apparent in late autumn when the strongest wind-stress occurred north of 28°S and a significant wind-stress minimum was observed to develop slightly north of Cape Columbine (33°S). This minimum was maintained throughout winter. The northern Benguela was characterised during winter by relatively strong south-easterly wind-

stress. A wind-stress maximum was present between 22 and 28°S for most of the winter season. In contrast, the southern Benguela was characterised by relatively weak westerly to south-westerly wind-stress. A southward migration of south-easterly wind-stress was observed during early spring. Associated with this migration was the re-establishment of the wind-stress maxima at approximately 17, 27 and 34°S. By November the Benguela Upwelling System was once again characterised by southerly to south-easterly wind stress. North of 15°S, light south-westerly wind-stress was observed to occur throughout the year close to the coast. Further offshore, this wind-stress strengthened and became more south-easterly in direction. In terms of wind-stress curl, the wedge of cyclonic curl observed by Bakun and Nelson (1991) was most visible during winter when it extended southwards from approximately 18°S to between 32 and 34.5°S. The rest of the year was broadly characterised by a band of cyclonic wind-stress curl extending between 300 and 400km offshore. This band of cyclonic wind-stress curl was observed to stretch from 19 to 34.5°S. Further north, a secondary maximum was found to occur throughout the year at approximately 16°S.

Wind-stress variability was investigated using a type of artificial neural network, known as the Kohonen Self Organising Map (SOM), as well as a wavelet analysis. Two independent SOM studies were conducted. The first study produced a 6x4 SOM output array, with each of the 24 output patterns representing a synoptic pattern within the data. The seasonal variability of these synoptic states was highlighted using seasonal frequency maps, while the movement of SOM patterns over the SOM output array was used to study the temporal evolution of two synoptic-scale wind events (14-22 October 1999 and 11-21 February 2000). The second study used the SOM to identify six distinct wind regimes within the Benguela. The northern most cell (10-15°S) was characterised by weak south-westerly winds throughout the study period, while the cells of the central Benguela were characterised by relatively strong south-easterly wind-stress throughout the study period. Between 29 and 35°S, south-easterly wind-stress was observed to occur from late spring through to the end of autumn. During the months of winter and early spring, the region was shown to experience relatively weaker westerly to south-westerly wind-stress.

The v wind-stress component, which was found to closely approximate the alongshore wind-stress off the west coast of Southern Africa, was extracted from each of the six

wind cells. A spectral analysis was conducted using a wavelet transform on each of the time series (01 August 1999 - 29 November 2000). The wavelet power spectra spanned a range of frequencies from 4 to 64 days. The event-like, intermittent, and pulsating nature of the variability within the Benguela Upwelling System was clear in the wavelet analysis. Regions of high variability were well resolved and tended to appear as isolated peaks. The occurrence and duration of these spectral peaks was, however, somewhat irregular, and strongly dependent upon latitude. All the cells south of 15°S, apart from Cell 3, had temporally variable 40-60 day peaks. These 40-60 day oscillations may be linked to the Madden-Julian oscillation (MJO) that was discovered by Madden and Julian (1971, 1972). Anderson *et al.* (1984) showed that such quasi-periodic oscillations have amplitudes and frequencies that vary widely, with periods ranging from 31-63 days. Madden and Julian (1994) concluded that this 40-50 day oscillation was a relatively broadband phenomenon with a central frequency that was not fixed but varied with time. Dickey *et al.* (1991) and Marcus *et al.* (1994) indicated the possibility of a separate non-tropical source for these oscillations, possibly through the interaction of zonal westerly flow with mountain topography (e.g. the Andes Mountains). Cells 3, 4 and 5 were all found to contain a band of high frequencies ranging from 4-12 days. This broad peak, which includes the so-called synoptic cycle of 2-8 days (Preston-Whyte and Tyson, 2000), was most pronounced in Cell 3, becoming less so to the south. The results presented here are consistent with authors such as Preston-Whyte and Tyson (1973), Jury *et al.* (1990a) and Ko and Vincent (1995).

Case study sequences of the spatial distribution of wind-stress, wind-stress curl and SST in a region off the west coast of southern Africa (25-30°S and 12-17°E) were discussed in relation to an intense, upwelling-favourable wind event that occurred from 11-20 February 2000. The winds were forced by the interaction between trapped coastal lows situated off the west coast of southern Africa, eastward propagating mid-latitude cyclones south of Africa, the interior heat low, and the ridging of the South Atlantic Anticyclone south of South Africa. The QuikSCAT data revealed that these upwelling-favourable winds lasted for 10 days, reaching peak intensity 300km off the coast at approximately 27°S between the 14 and 16 February 2000. Towards the end of the event, a gradual reduction in wind-stress propagated down the coast in a poleward direction. This reduction in wind-stress resulted in the previously coherent

pattern of active upwelling breaking down into a coalescence of warm eddies, cold patches and relatively smaller filaments. The temporal succession and magnitude of such events has been shown in previous work (e.g. Roy *et al.*, 2001) to have direct links to phytoplankton abundance, primary production and fish recruitment in the southern Benguela Ecosystem.

Finally, one unanswered question that arises from this work relates to the origin of the oscillations in surface wind-stress off the south-west coast of Africa. Breaker and Lewis (1988) speculated that similar intra-seasonal oscillations off the central Californian coast might be traceable to the tropics and linked to mid-latitudes along the Californian coast through atmospheric teleconnections. More recent research suggests that an independent source for these oscillations may arise at mid-latitudes through the interaction of non-zonal westerly flow with mountain topography (Marcus *et al.*, 1994). To answer this complex, question one approach would be to combine scatterometer data with, for example, sea level, current and SST data into a model similar to that of Hormazabal *et al.* (2002). Using a simple oceanic equatorial Kelvin/coastal trapped wave model forced by ERS scatterometer wind data from the tropical and south Pacific, Hormazabal *et al.* (2002) were able to show how intra-seasonal winds along the equatorial Pacific strongly influence atmospheric and oceanic conditions off the Chilean coast by way of long waves in the atmosphere and ocean.

CHAPTER 7

REFERENCES

- Agenbag J. J. and L. V. Shannon**, 1987. A preliminary note on a recent perturbation in the Agulhas Current retroflexion area. *Tropical Ocean-Atmosphere Newsletter*, **37**, 10-11.
- Ainsworth, E. J.**, 1999. Visualization of ocean colour and temperature from multispectral imagery captured by the Japanese ADEOS satellite. *Journal of Visualization*, **2** (2), 195-204.
- Ainsworth, E. J. and S. F. Jones**, 1999. Radiance spectra classification from the ocean color and temperature scanner on ADEOS. *IEEE Transactions on Geoscience and Remote Sensing*, **37** (3), 1645-1656.
- Anderson, J. R. and R. D. Rosen**, 1983. The latitude-height structure of 40–50 day variations in atmospheric angular momentum. *J. Atmos. Sci.*, **40**, 1584-1591.
- Anderson, J. R., D. E. Stevens and P. R. Julian**, 1984. Temporal variations of the tropical 40-50 day oscillation. *Mon. Wea. Rev.*, **112**, 2431-2438.
- Bang, N. D.**, 1973. Characteristics of an intense ocean frontal system in the upwelling regime west of Cape Town. *Tellus*, **25**, 256-265.
- Bang, N. D. and W. R. H. Andrews**, 1974. Direct measurements of a shelf-edge frontal jet in the southern Benguela system. *J. Mar. Res.*, **32** (3), 405-417.
- Bakun, A. and C. S. Nelson**, 1991. The seasonal cycle of wind-stress curl in subtropical eastern boundary current regions. *J. Phys. Oceanogr.*, **21**, 1815-1834.

- Barton, E. D.**, 1998. Eastern boundary of the North Atlantic: Northwest Africa and Iberia. In *The Sea*, Vol. 11, *The Global Coastal Ocean; Regional Studies and Syntheses*, A. R. Robinson and K. H. Brink, eds. Wiley, New York, Chap. 22.
- Bernstein, R. L., L. Breaker and R. Whritner**, 1977. California Current eddy formation: shipboard, air and satellite results. *Science*, **195**, 353-359.
- Boudra, D. B. and E. P. Chassignet**, 1988. Dynamics of the Agulhas retroflection and ring formation in a numerical model. Part I: The vorticity balance. *J. Phys. Oceanogr.*, **18**, 280-303.
- Breaker, L. C. and P. A. W. Lewis**, 1988. A 40 to 50 oscillation in sea surface temperature along the central Californian coast. *Estuarine, Coastal, and Shelf Science*, **26**, 395-408.
- Breaker, L. C., P. C. Liu and C. Torrence**, 2001. Intraseasonal oscillations in sea surface temperature, wind-stress and sea level off the central California coast. *Cont. Shelf Res.*, **21**, 727-750.
- Brink, K. H.**, 1982. A comparison of long coastal trapped wave theory with observations off Peru. *J. Phys. Oceanogr.*, **12**, 897-913.
- Brundrit, G. B., B. de Cuevas and A. M. Shipley**, 1984. Significant sea-level variations along the west coast of southern Africa 1979-83. *S. Afr. J. Sci.*, **80** (2), 80-82.
- Carton, J. A., and B. Huang**, 1994. Warm events in the tropical Atlantic. *J. Phys. Oceanogr.*, **24**, 888-903.
- Chelton, D. B., S. K. Esbensen, M. G. Schlax, N. Thum, M. H. Freilich, F. J. Wentz, C. L. Gentemann, M. J. McPhaden and P. S. Schopf**, 2001. Observations of coupling between surface wind-stress and sea surface temperature in the eastern tropical Pacific. *J. Clim.*, **14** (7), 1479-1498.

- Chao S.-Y.**, 1985. Coastal jets in the lower atmosphere. *J. Phys. Oceanogr.*, **15**, 361-371.
- Chui, C. K.**, 1992. An Introduction to Wavelets. Academic Press, New York and Harcourt Brace and Jovanovich, New York, 266pp.
- Cutchin, D. L. and R. L. Smith**, 1973. Continental shelf waves: low frequency variations in seal-level and currents over the Oregon continental shelf. *J. Phys. Oceanogr.*, **12**, 897-913.
- Daubechies, I.**, 1992. Ten Lectures on Wavelets. Society for Industrial and Applied Mathematics, 357pp
- Dayhoff, J. E.**, 1990. Neural network architectures: An introduction. Van Nostrand Reinhold, New York, 259 pp.
- Dengler, A. T.**, 1985. Relationship between physical and biological processes at an upwelling front off Peru. *Deep-Sea Res.*, **32** (11A), 1301-1315.
- Dickey, J. O., M. Ghil and S. L. Marcus**, 1991. Extratropical aspects of the 40–50 oscillation in the length-of-day and atmospheric angular momentum. *J. Geophys. Res.*, **96**, 22643-22658.
- Duncombe Rae, C. M.**, 1991. Agulhas retroflection rings in the South Atlantic Ocean; an overview. *S. Afr. J. Sci.*, **11**, 327-344.
- Duncombe Rae, C. M., F. A. Shillington, J. J. Agenbag, J. Taunton-Clark and M. L. Gründlingh**, 1992. An Agulhas ring in the South Atlantic Ocean and its interaction with the Benguela Upwelling frontal system. *Deep-Sea Res.*, **39** (11-12), 2009-2027.
- Emery, W. J., and R. E. Thomson**, 1998. Data Analysis Methods in Physical Oceanography. Pergamon, Oxford, 500pp

- Enfield, D. B.**, 1981. Thermally driven wind variability in the planetary boundary layer above Lima, Peru. *J. Geophys. Res.*, **86**, 2005-2016.
- Flament, P.**, 2000. Ocean Winds. CERSAT News (Editorial). November Issue No. 11.
- Flament, P., L. Armi and L. Washburn**, 1985. The evolving structure of an upwelling filament. *J. Geophys. Res.*, **90**, 11765-11778.
- Frederiksen, J. S. and P. J. Webster**, 1988. Alternative theories of atmospheric teleconnections and low-frequency fluctuations. *Rev Geophys.*, **26**, 459-494.
- Freilich, M. H. and R. S. Dunbar**, 1999. The accuracy of the NSCAT-1 vector winds: Comparisons with National Data Buoy Centre buoys. *J. Geophys. Res.*, **104**, 11231-11246.
- Gammelsrød, T., C. H. Bartholomae, D. C. Boyer, V. L. L. Filipe and M. J. O'Toole**, 1998. Intrusion of warm surface water along the Angolan-Namibian coast in February-March 1995: the 1995 Benguela Niño. In *Benguela Dynamics*, S. C. Pillar, C. L. Moloney, A. I. L. Payne and F. A. Shillington, eds. *S. Afr. J. Mar. Sci.*, **19**, 41-56.
- Gordon, A. L., J. R. E. Lutjeharms and M. L. Gründlingh**, 1987. Stratification and circulation at the Agulhas retroflection. *Deep-Sea Res.*, **34** (4A), 565-599.
- Hart, T. J. and R. I. Currie**, 1960. The Benguela Current. *Discovery Rep.*, **31**, 123-298.
- Hartmann, D. L., M. L. Michelson and S. A. Klein**, 1991. Seasonal variations of tropical intraseasonal oscillations: A 20–25 day oscillation in the western Pacific. *J. Atmos. Sci.*, **49**, 1277-1289.
- Hewitson, B. C., and R. G. Crane**, 1994. Neural Nets: Applications in Geography. Kluwer Academic Publishers, London.

- Hewitson, B. C., and R. G. Crane**, 2002. Self-organizing maps: Applications to synoptic climatology. *Clim. Res.*, **22** (1), 13-26.
- Heydenrych, C.M.**, 1987. A climatology of the coastal low in the SW Cape, unpublished M.Sc. dissertation, University of Cape Town, 135pp.
- Hickey, B. M.**, 1998. Coastal oceanography of western North America from the tip of Baja California to Vancouver Island. In *The Sea*, Vol. 11, *The Global Coastal Ocean; Regional Studies and Syntheses*, A. R. Robinson and K. H. Brink, eds. Wiley, New York, Chap. 12.
- Hill, A. E., B. M. Hickey, F. A. Shillington, P. T. Strub, K. H. Brink, E. D. Barton and A. C. Thomas**, 1998. Eastern ocean boundaries. In *The Sea*, Vol. 11, *The Global Coastal Ocean; Regional Studies and Syntheses*, A. R. Robinson and K. H. Brink, eds. Wiley, New York, Chap. 2.
- Hormazabal, S., G. Shaffer and O. Pizarro**, 2002. Tropical Pacific control of intraseasonal oscillations off Chile by way of oceanic and atmospheric pathways. *Geophys. Res. Lett.*, **29** (6), 1-4.
- Hunter, I. T.**, 1987. Weather of the Agulhas Bank and the Cape south coast, unpublished M.Sc. dissertation, University of Cape Town, 182pp.
- Huthnance, J. M.**, 1995. Circulation, exchange and water masses at the ocean margin: the role of physical processes at the shelf edge. *Prog. Oceanogr.*, **35**, 353-431.
- Johnson, A. S.**, 2001. Analysis of variable scatterometer wind fields in the Benguela Upwelling Region, unpublished M.Sc. dissertation, University of Cape Town, 53pp.
- Jones, D. A. and I. Simmonds**, 1993. Time and space spectral analyses of Southern Hemisphere sea level pressure variability. *Mon Wea. Rev.*, **121**, 661-672.

- Jury, M. R.**, 1984. Wind shear and differential upwelling along the SW tip of Africa. PhD Thesis, University of Cape Town, 51pp.
- Jury, M. R.**, 1985. Case studies of alongshore variations in wind-driven upwelling in the southern Benguela region. In *South African Ocean Colour Experiment*, edited by L. V. Shannon, pp. 29-46, Sea Fisheries Research Institute, Cape Town, South Africa.
- Jury, M. R., C. MacArthur and G. B. Brundrit**, 1990a. Pulsing of the Benguela upwelling region: Large-scale atmospheric controls. *S. Afr. J. Mar. Sci.*, **9**, 27-41.
- Jury, M. R., C. MacArthur and C. J. C. Reason**, 1990b. Observations of trapped waves in the atmosphere and ocean along the coast of southern Africa. *S. Afr. Geog. J.*, **72**, 33-46.
- Kalnay, E. and Coauthors**, 1996. The NCEP/NCAR 40-year reanalysis project. *Bull. Amer. Meteor. Soc.*, **77**, 437-471.
- Kamstra, F.**, 1985. Environmental features of the southern Benguela with special reference to wind-stress. In *South African Ocean Colour Experiment*, edited by L. V. Shannon, pp. 13-27, Sea Fisheries Research Institute, Cape Town, South Africa.
- Kamstra, F.**, 1987. Interannual variability in the spectra of the daily surface pressure at four stations in the southern hemisphere. *Tellus*, **39A**, 509-514.
- Kaz'min A. S. and G. G. Sutyrin**, 1990. Anticyclone blocking of the Benguela Current: satellite and shipborne data analysis. *Sov. J. Remote Sensing*, **7** (6), 986-995.

- Kiladis, G. N. and K. C. Mo**, 1998. Interannual and Intraseasonal Variability in the Southern Hemisphere, In *Meteorology of the Southern Hemisphere*, D. J. Koroly and D. G. Vincent, eds. American Meteorological Society, Boston, Chap. 8.
- Kiladis, G. N. and M. Wheeler**, 1995. Horizontal and vertical structure of observed tropospheric equatorial Rossby waves. *J. Geophys. Res.*, **100**, 22981-22997.
- Ko, K.-C., and D. G. Vincent**, 1995. A composite study of the quasi-periodic subtropical wind maxima over the south Pacific during November 1984 – April 1985. *J. Clim.*, **8**, 579-588.
- Kohonen, T., J. Hynninen, J. Kangas and J. Laaksonen**, 1995. The self-organizing map program package. Version 3.1. <http://www.cis.hut.fi/research/som-research/nnrc-programs.shtml>
- Large, W. G., and S. Pond**, 1981. Open ocean momentum flux measurements in moderate to strong winds. *J. Phys. Oceanogr.*, **11**, 324-336.
- Lass, H. U., M. Schmidt, V. Mohrholz and G. Nausch**, 2000. Hydrographic and current measurements in the area of the Angola-Benguela front. *J. Phys. Oceanogr.*, **30** (10), 2589-2609.
- Lau, K. -M. and H. -Y. Weng**, 1995. Climate signal detection using wavelet transform: How to make a time series sing. *Bull. Amer. Meteor. Soc.*, **76** (12), 61-78.
- Lee, S. and I. M. Held**, 1993. Baroclinic wave packets in models and observations. *J. Atmos. Sci.*, **50**, 1413-1428.
- Liebmann, B. and H. H. Hendon**, 1990. Synoptic-scale disturbances near the Equator. *J. Atmos. Sci.*, **50**, 1463-1479.

- Lutjeharms, J. R. E., and J. M. Meeuwis**, 1987. The extent and variability of south-east Atlantic upwelling. In *The Benguela and Comparable Ecosystems*, A. I. L. Payne, J. A. Gulland and K. H. Brink, eds. *S. Afr. J. Mar. Sci.*, **5**, 51-62.
- Lutjeharms, J. R. E., F. A. Shillington and C. M. Duncombe Rae**, 1991. Observations of extreme upwelling filaments in the southeast Atlantic Ocean. *Science* (N.Y.), **253** (5021), 774-776.
- Lutjeharms, J. R. E., and P. L. Stockton**, 1987. Kinematics of the upwelling front off southern Africa. In *The Benguela and Comparable Ecosystems*, A. I. L. Payne, J. A. Gulland and K. H. Brink, eds. *S. Afr. J. Mar. Sci.*, **5**, 25-49.
- Lutjeharms, J. R. E., and H. R. Valentine**, 1987. Water types and volumetric considerations of the south-east Atlantic Upwelling Regime. In *The Benguela and Comparable Ecosystems*, A. I. L. Payne, J. A. Gulland and K. H. Brink, eds. *S. Afr. J. Mar. Sci.*, **5**, 63-71.
- Lutjeharms, J. R. E. and R. C. Van Ballegooyen**, 1988. The retroflection of the Agulhas Current. *J. Phys. Oceanogr.*, **18**, 1570-1583.
- Lutjeharms, J. R. E., and N. M. Walters**, 1985. Ocean colour and thermal fronts south of Africa. In *South African Ocean Colour Experiment*, edited by L. V. Shannon, pp. 227-237, Sea Fisheries Research Institute, Cape Town, South Africa.
- Madden, R. A.**, 1986. Seasonal variations of the 40–50 day oscillation in the tropics. *J. Atmos. Sci.*, **43**, 3138-3158.
- Madden, R. A. and P. Julian**, 1971. Detection of a 40–50 day oscillation in the zonal wind. *J Atmos. Sci.*, **28**, 702-708
- Madden, R. A. and P. Julian**, 1972. Description of global-scale circulation cells in the Tropics with a 40–50 period. *J Atmos. Sci.*, **29**, 1109-1123.

- Madden, R. A. and P. Julian**, 1994. Observations of the 40–50 day tropical oscillation – a review. *Mon. Wea. Rev.*, **122**, 814-837.
- Main, J. P. L.**, 1997. Seasonality of Circulation in Southern Africa using the Kohonen Self-Organising Map. unpublished M.Sc. dissertation, University of Cape Town.
- Marcus, S. L., M. Ghil and J. O. Dickey**, 1994. The extratropical 40-day oscillation in the UCLA General Circulation Model. Part 1: atmospheric momentum. *J. Atmos. Sci.*, **51**, 1431-1446.
- Meehl, G. A., J. W. Hurrell and H. van Loon**, 1998. A modulation of the mechanism of the semiannual oscillation in the Southern Hemisphere. *Tellus*, **50A**, 442-450.
- Meeuwis, J. M. and J. R. E. Lutjeharms**, 1990. Surface thermal characteristics of the Angolan-Benguela Front. *S. Afr. J. Mar. Sci.*, **9**, 261-279.
- Mélice, J. L., A. Coron and A. Berger**, 2001. Amplitude and frequency modulation of the Earth's obliquity for the last million years. *J. Clim.*, **14**, 1043-1054.
- Misra, M.**, 1997. Parallel environment for implementing neural networks. *Neural Computing Surveys*, **1**, 48-60.
- Nelson, G.**, 1985. Notes on the physical oceanography of the Cape Peninsula upwelling system. In *South African Ocean Colour Experiment*, edited by L. V. Shannon, pp. 63-95, Sea Fisheries Research Institute, Cape Town, South Africa.
- Nelson, G.**, 1989. Poleward motion in the Benguela area. In *Poleward Flows Along Eastern Ocean Boundaries*, S. J. Neshyba, C. N. K. Mooers, R. L. Smith and R. T. Barber, eds., Coastal and Estuarine Studies, Vol. 34. Springer-Verlag New York, 110-130pp.

- Nelson, G.**, 1992. Equatorward wind and Atmospheric pressure spectra as metrics for primary productivity in the Benguela system *S. Afr. J. Mar. Sci.*, **12**, 19-28.
- Nelson, G. and L. Hutchings**, 1983. The Benguela Upwelling area. *Prog. Oceanogr.*, **3**, 333-356.
- Nelson, G. and A. Polito**, 1987. Information on currents in the Cape Peninsula area, South Africa. In *The Benguela and Comparable Ecosystems*, A. I. L. Payne, J. A. Gulland and K. H. Brink, eds. *S. Afr. J. Mar. Sci.*, **5**, 287-304.
- Park, C. and S. D. Schubert**, 1992. Remotely forced intraseasonal oscillations over the tropical Atlantic *J. Atmos. Sci.*, **50**, 89-103.
- Parker, D. E.**, 1973. Equatorial Kelvin waves at 100 millibars. *Q. J. R. Meteorol. Soc.*, **90**, 116-129.
- Parrish, R. H., A. Bakun, D. M. Husby and C. S. Nelson**, 1983. Comparative climatology of selected environmental processes in relation to eastern boundary current pelagic fish reproduction. In *Proceedings of the Expert Consultation to Examine Changes in Abundance and Species Composition of Neritic Fish Resources, Costa Rica, April 1983*. Sharp, G. D. and J. Csirke (Eds.). *F.O.A. Fish. Rep.*, **291** (3), 731-777
- Perry, K. L.**, 2000. SeaWinds on QuikSCAT Level 3 Daily, Gridded Ocean Wind Vector (JPL SeaWinds Project). http://podaac.jpl.nasa.gov:2031/DATASET_DOCS/qscat_L3.html
- Perry, K. L.**, 2001. QuikSCAT Science Data Product User's Manual. Version 2.0. ftp://podaac.jpl.nasa.gov/pub/ocean_wind/quikscat/doc/QSUG4-4.doc
- Peterson, R. G. and L. Stramma**, 1991. Upper-level circulation in the South Atlantic Ocean. *Prog. Oceanogr.*, **26**, 1-73.

- Portabella, M. and A. Stoffelen**, 2001. Rain detection and quality control of SeaWinds. *J. Atm. and Ocean Techn.*, **18** (7), 1171-1183.
- Preston-Whyte, R. A. and P. D. Tyson**, 1973. Note on pressure oscillations over South Africa. *Mon. Weath. Rev.*, **101** (8), 650-653.
- Preston-Whyte, R. A. and P. D. Tyson**, 2000. The Weather and Climate of Southern Africa. Oxford University Press, Cape Town.
- Reason, C. J. C.**, 2000. Multidecadal climate variability in the subtropics/mid-latitudes of the Southern Hemisphere oceans. *Tellus*, **52A**, 203-223.
- Reason, C. J. C., R. J. Allan, J. A. Lindesay and T. J. Ansell**, 2000. ENSO and climatic signals across the Indian Ocean Basin in the global context: Part 1, Interannual composite patterns. *Int. J. Climatol.* **20**, 1285-1327. 2000.
- Reason, C. J. C. and M. R. Jury**, 1990. On the generation and propagation of the southern African coastal low. *Q. J. R. Meteorol. Soc.*, **116**, 1133-1151.
- Reason, C. J. C. and D. G. Steyn**, 1990. Coastally trapped disturbances in the lower atmosphere: dynamic commonalities and geographic diversity. *Prog. Phys. Geog.*, **14** (2), 178-198.
- Reid, J. L.**, 1989. On the total geostrophic circulation of the South Atlantic Ocean: Flow patterns, tracers and transports. *Prog. Oceanogr.*, **23**, 149-244.
- Rouault, M., P. Florenchie, N. Fauchereau and C. J. C. Reason**, 2002. South-east tropical Atlantic warm events and southern African rainfall. *Geophys. Res Lett.*, in press.
- Roy, C., S. Weeks, M. Rouault, G. Nelson, R. Barlow and C. van der Lingen**, 2001. Extreme oceanographic events recorded in the southern Benguela during the 1999–2000 summer season. *S. Afr. J. Sci.*, **97**, 465-471.

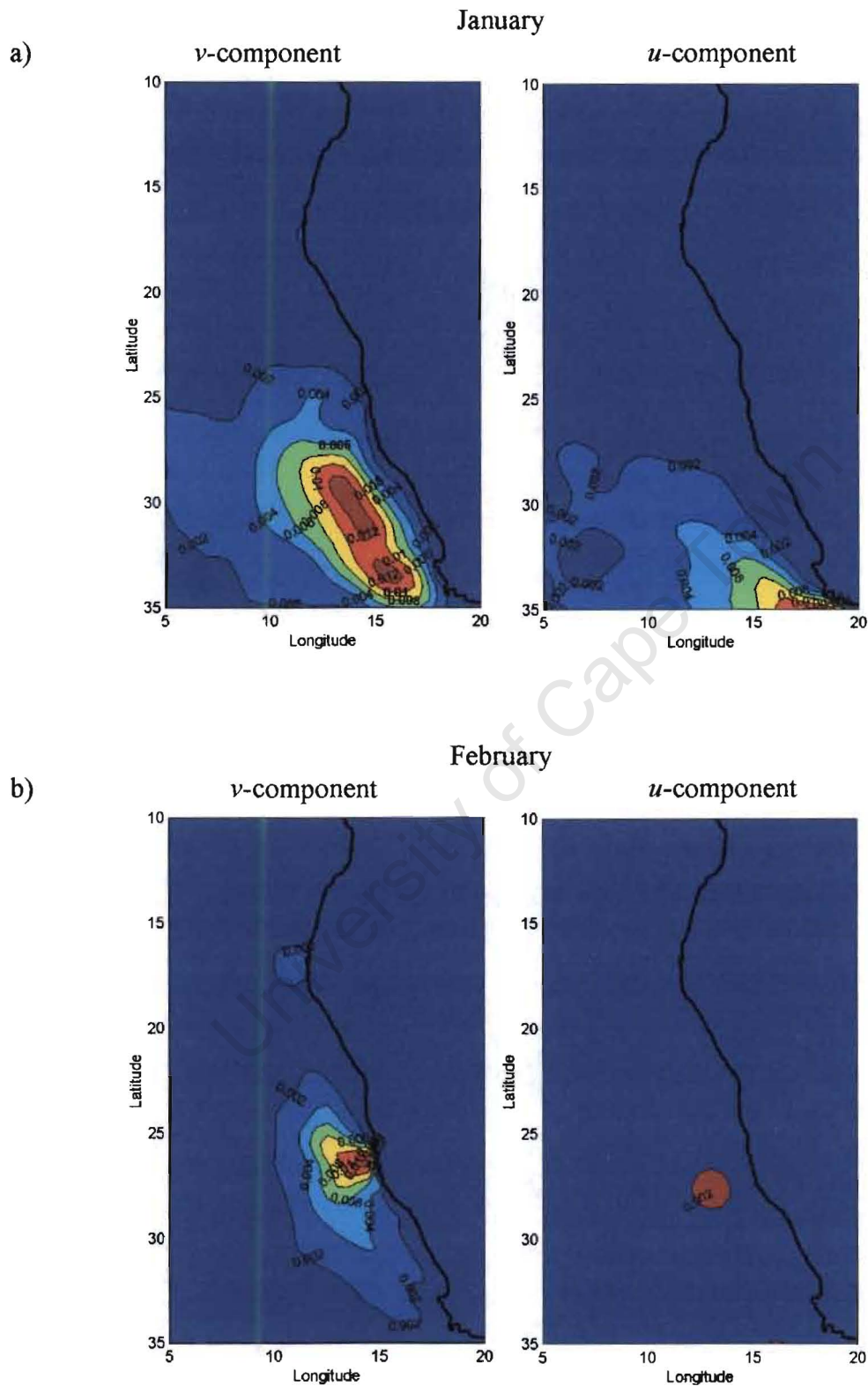
- Schumann, E. H.**, 1983. Long period coastal trapped waves off the south-east coast of southern Africa. *Cont. Shelf Res.*, **2**, 97-107.
- Schumann, E. H. and K. H. Brink**, 1990. Coastal-trapped waves off the coast of South Africa: Generation, propagation and current structures. *J. Phys. Oceanogr.*, **20**, 1206-1218.
- Servain, J., J. Picaut and J. Merle**, 1982. Evidence of remote forcing in the equatorial Atlantic Ocean. *J. Phys. Oceanogr.*, **12**, 457-463.
- Shannon L.V.**, 1985. The Benguela ecosystem. Part 1. Evolution of the Benguela, physical features and processes. *Oceanogr. Mar. Biol. Ann. Rev.*, **23**, 105-182.
- Shannon, L. V., J. J. Agenbag and M. E. L. Buys**, 1987. Large- and mesoscale features of the Angola-Benguela Front. In *The Benguela and Comparable Ecosystems*, A. I. L. Payne, J. A. Gulland and K. H. Brink, eds. *S. Afr. J. Mar. Sci.*, **5**, 11-34.
- Shannon, L. V., J. J. Agenbag, N. D. Walker and J. R. E. Lutjeharms**, 1990a. A major perturbation in the Agulhas retroflexion area in 1986. *Deep-Sea Res.*, **37**, 493-512.
- Shannon, L. V., A. J. Boyd, G. B. Brundrit and J. Taunton-Clark**, 1986. On the existence of an El Niño-type phenomenon in the Benguela system. *J. Mar. Res.*, **44** (3), 495-520.
- Shannon, L. V., J. R. E. Lutjeharms and J. J. Agenbag**, 1989. Episodic input of subantarctic water into the Benguela region. *S. Afr. J. Sci.*, **85** (5), 317-322.
- Shannon, L. V., J. R. E. Lutjeharms and G. Nelson**, 1990b. Causative mechanisms for intra-annual and interannual variability in the marine environment around southern Africa, *S. Afr. J. Sci.*, **86**, 356-373.

- Shannon, L. V., Mostert, S. A. Walters, N. M. and F. P. Anderson, 1983.**
Chlorophyll concentrations in the southern Benguela current region as determined by satellite (*Nimbus-7* coastal zone colour scanner). *J. Plankt. Res.*, **5** (4), 565-583.
- Shannon, L. V. and G. Nelson, 1996.** The Benguela: large scale features and processes and system variability. In *The South Atlantic: Present and Past Circulation*, G. Wefer, W. H. Berger, G. Seidler and D. J. Webb, eds. Springer-Verlag, Berlin, Heidelberg, pp. 163-210.
- Shillington, F. A., 1998.** The Benguela Upwelling System off southwestern Africa. In *The Sea*, Vol. 11, *The Global Coastal Ocean; Regional Studies and Syntheses*, A. R. Robinson and K. H. Brink, eds. Wiley, New York, Chap. 20.
- Shillington, F. A., L. Hutchings, T. A. Probyn, H. N. Waldron and W. T. Peterson, 1992.** Filaments and the Benguela Frontal zone: offshore advection or recirculating loops? In *Benguela Trophic Functioning*, A. I. L. Payne, K. H. Brink, K. H. Mann and R. Hilborn, eds. *S. Afr. J. Mar. Sci.*, **12**, 207-218.
- Shillington, F. A., W. T. Peterson, L. Hutchings, T. A. Probyn, H. N. Waldron and J. J. Agenbag, 1990.** A cool upwelling filament of Namibia, southwest Africa: preliminary measurements of physical and biological properties. *Deep-Sea Res.*, **37** (11A), 1753-1772.
- Silulwane, N. F., A. J. Richardson, F. A. Shillington and B. A. Mitchell-Innes, 2001.** Marine applications of an Artificial Neural Network: Identification and classification of vertical chlorophyll patterns in the Northern Benguela Upwelling System. *S. Afr. J. mar. Sci.*, **23**, 37-51.
- Simpson, J. H., D. J. Edelsten, A. Edwards, N. C. G. Morris and P. B. Tett, 1979.** The Islay front: physical structure and phytoplankton distribution. *Estuar. Coast. Mar. Sci.*, **9** (6), 713-726.

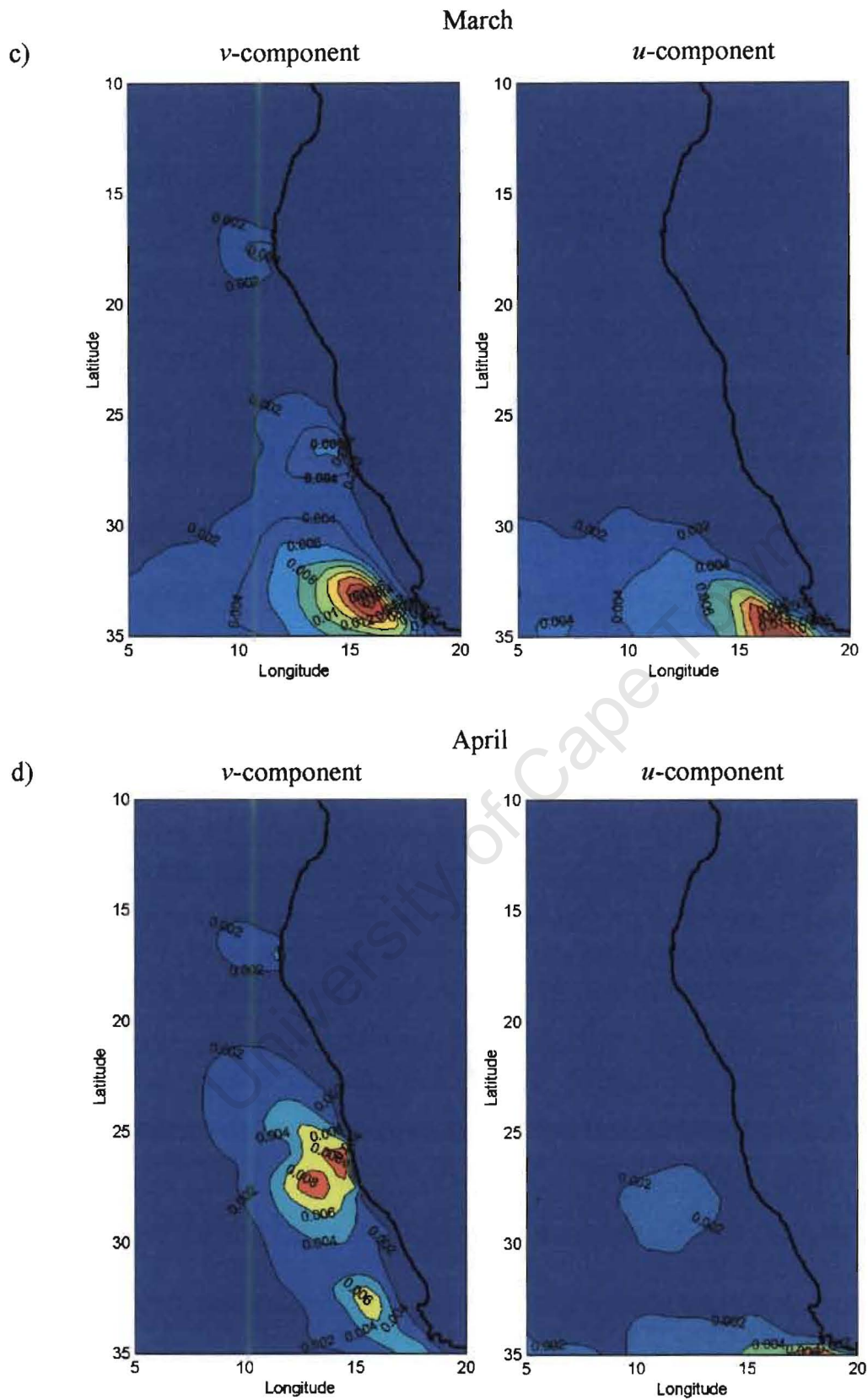
- Smith, R. L., A. Huyer, J. S. Godfrey and J. A. Church**, 1991. The Leeuwin Current off Western Australia, 1986-1987. *J. Phys. Oceanogr.*, **21**, 323-345.
- Steward, C. M.**, 1901. The hot winds of Namaqualand and the Berg winds of the South Coast. *Report of the Met. Commission for the Cape of Good Hope for 1901*, 183-185.
- Stramma, L. and R. G. Peterson**, 1989. Geostrophic transport in the Benguela Current. *J. Phys. Oceanogr.*, **19**, 1440-1448.
- Strub, P. T., P. M. Korso and A. Huyer**, 1991. The nature of the cold filaments in the California Current System. *J. Geophys. Res.*, **96** (C8), 14743-14768.
- Strub, P. T., J. M. Mesías, V. Montecino, J. Rutllant and S. Salinas**, 1998. Coastal ocean circulation off western South America. In *The Sea*, Vol. 11, *The Global Coastal Ocean; Regional Studies and Syntheses*, A. R. Robinson and K. H. Brink, eds. Wiley, New York, Chap. 10.
- Taljaard, J. J.**, 1995. Atmospheric circulation systems, synoptic climatology and weather phenomena of South Africa. Part 4: Surface pressure and wind phenomena in South Africa. Technical Paper No. 30. Government Printer, Pretoria.
- Taunton-Clark, J.**, 1985. The formation, growth and decay of upwelling tongues in response to the mesoscale wind field during summer. In *South African Ocean Colour Experiment*, edited by L. V. Shannon, pp. 47-61, Sea Fisheries Research Institute, Cape Town, South Africa.
- Taunton-Clark, J.**, 1990. Environmental events within the South-east Atlantic (1906-1985) identified by analysis of sea surface temperature and wind data. *S. Afr. J. Sci.*, **86**, 470-472.

- Taunton-Clark, J. and L. V. Shannon**, 1988. Annual and interannual variability in The South-east Atlantic during the 20th century. *S. Afr. J. Mar. Sci.*, **6**, 107-117.
- Tomczak, M., and J. S. Godfrey**, 1994. Regional Oceanography: An Introduction. Pergamon, Oxford, 217pp.
- Torrence, C. and G. P. Compo**, 1998. A practical guide to wavelet analysis. *Bull. Amer. Meteor. Soc.*, **79** (1), 61-78.
- Van Foreest, D., F. A. Shillington and R. Legeckis**, 1984. Large scale, stationary, frontal features in the Benguela Current system. *Cont. Shelf Res.*, **3** (4), 465-474.
- Walker, N. D.**, 1984. Passage of coastal lows on the west coast as shown in coastal weather station data. *Abstracts and Summary of the Coastal Low Workshop*, Institute of Maritime Technology, Simon's Town, 8-11.
- Walker, N. D.**, 1987. Interannual sea surface temperature variability and associated atmospheric forcing within the Benguela system. *S. Afr. J. Mar. Sci.*, **5**, 121-132.
- Weickmann, K. M.**, 1983. Intraseasonal circulation and outgoing longwave radiation modes during Northern Hemisphere winter, *Mon Wea. Rev.*, **111**, 1838-1858.
- Young, R. E. and H. Houben**, 1989. Dynamics of planetary-scale waves during Southern Hemisphere winter. *J. Atmos. Sci.*, **46**, 1365-1383.
- Zemba, J. and C. A. Friehe**, 1987. The marine atmospheric boundary layer jet in the Coastal Ocean Dynamics Experiment. *J. Geophys. Res.*, **92**, 1489-1496.

APPENDIX 1



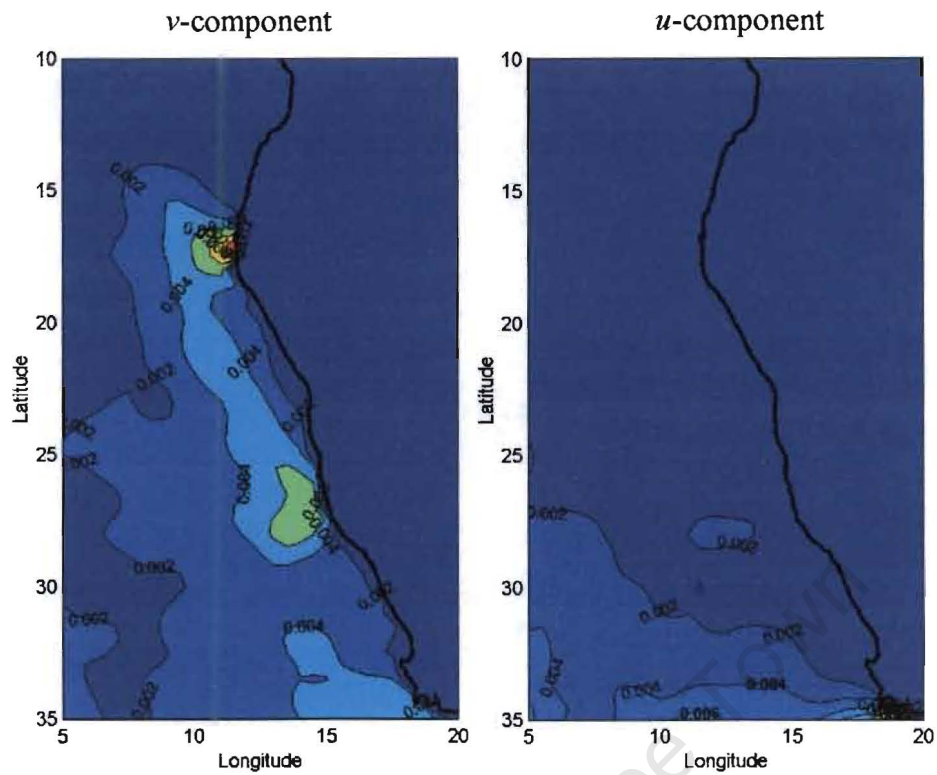
Appendix 1 The Benguela Upwelling Regime: monthly composites of the spatial variance of the u and v -components of the surface wind-stress field. The contour interval is 0.002. (a) January, (b) February.



Appendix 1 (Continued) (c) March, (d) April.

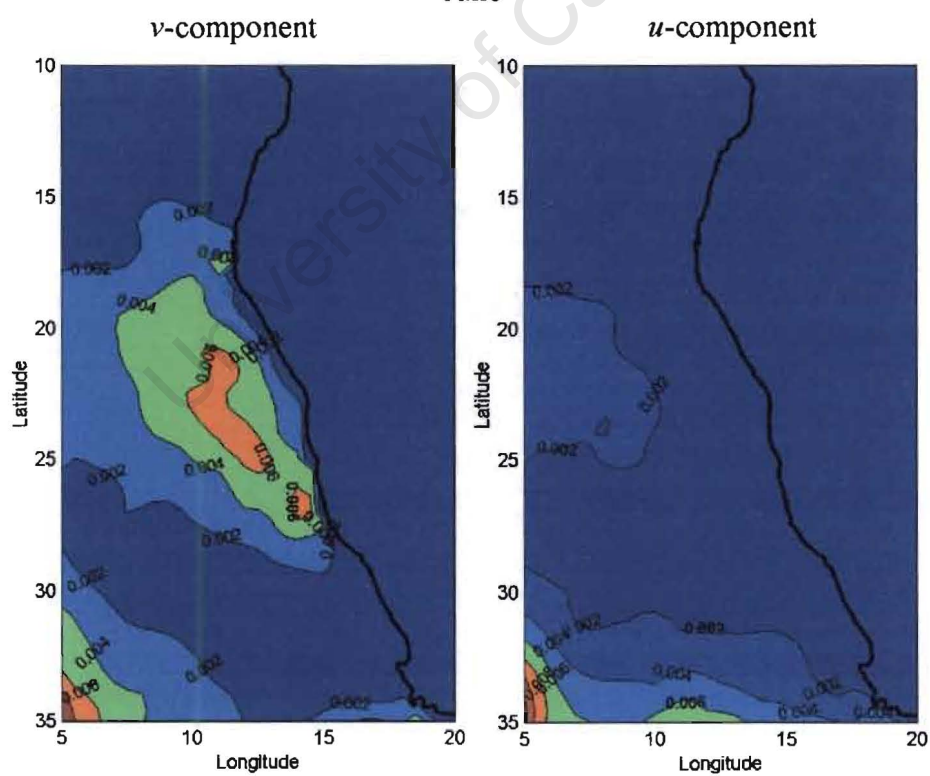
May

e)

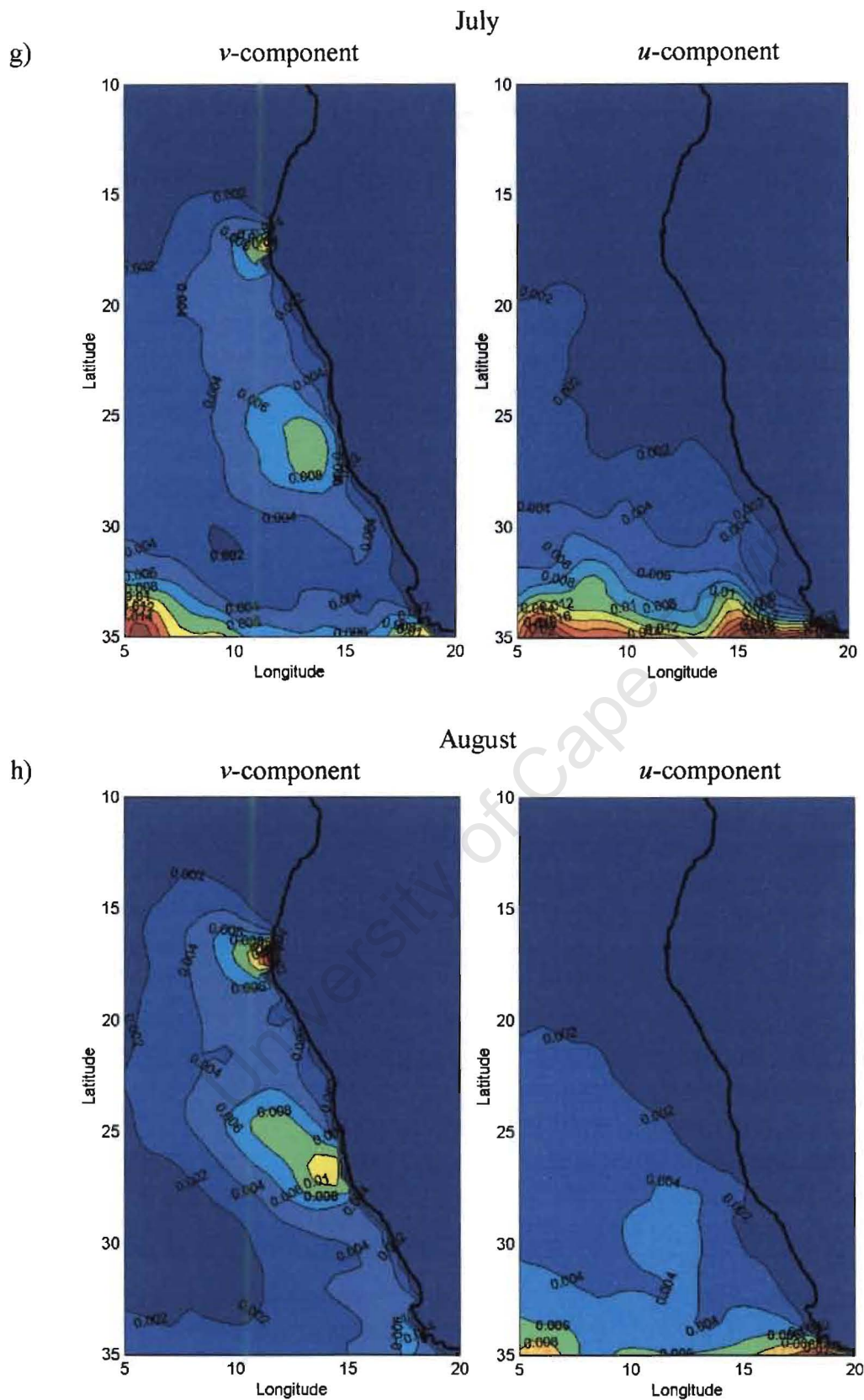


June

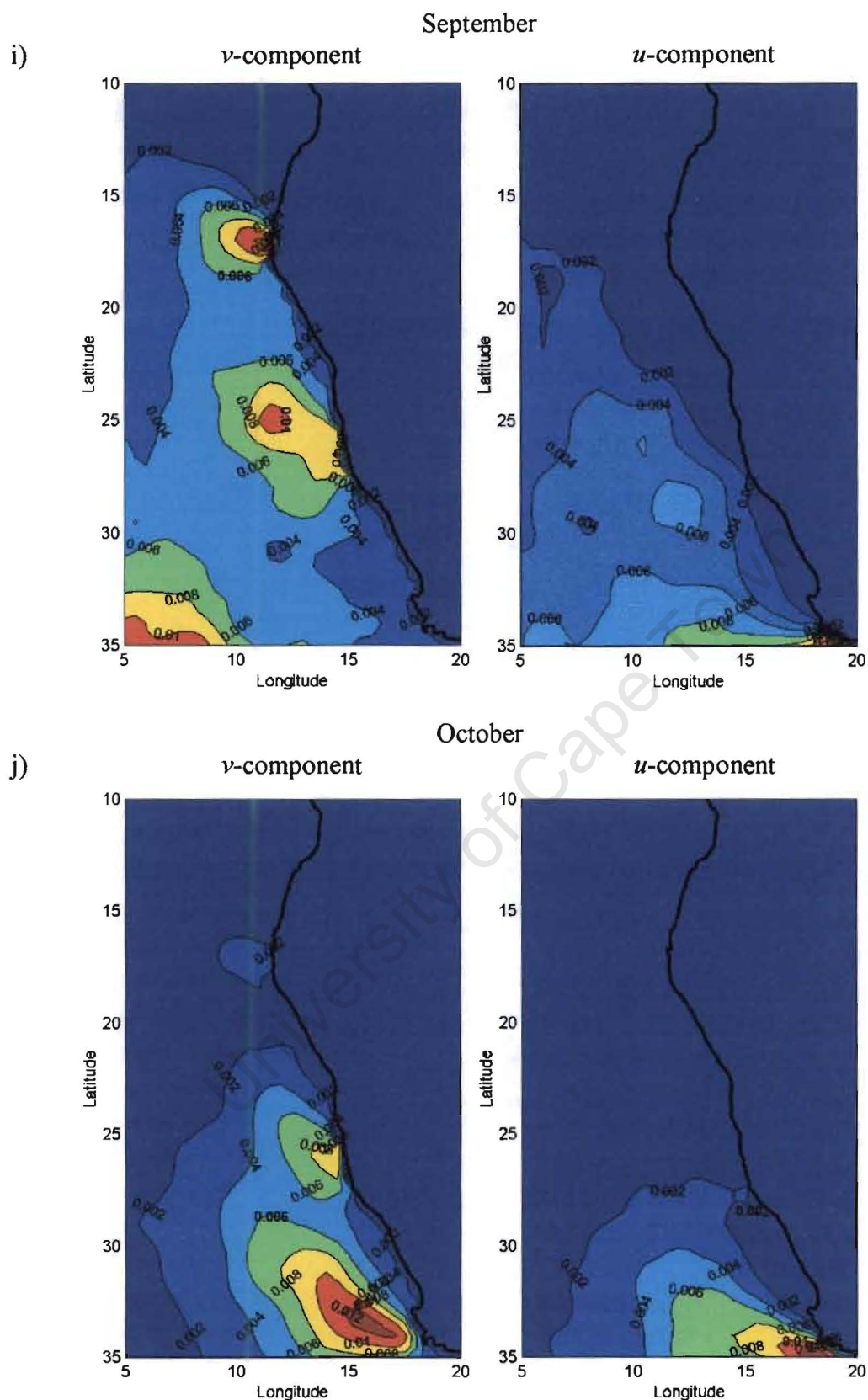
f)



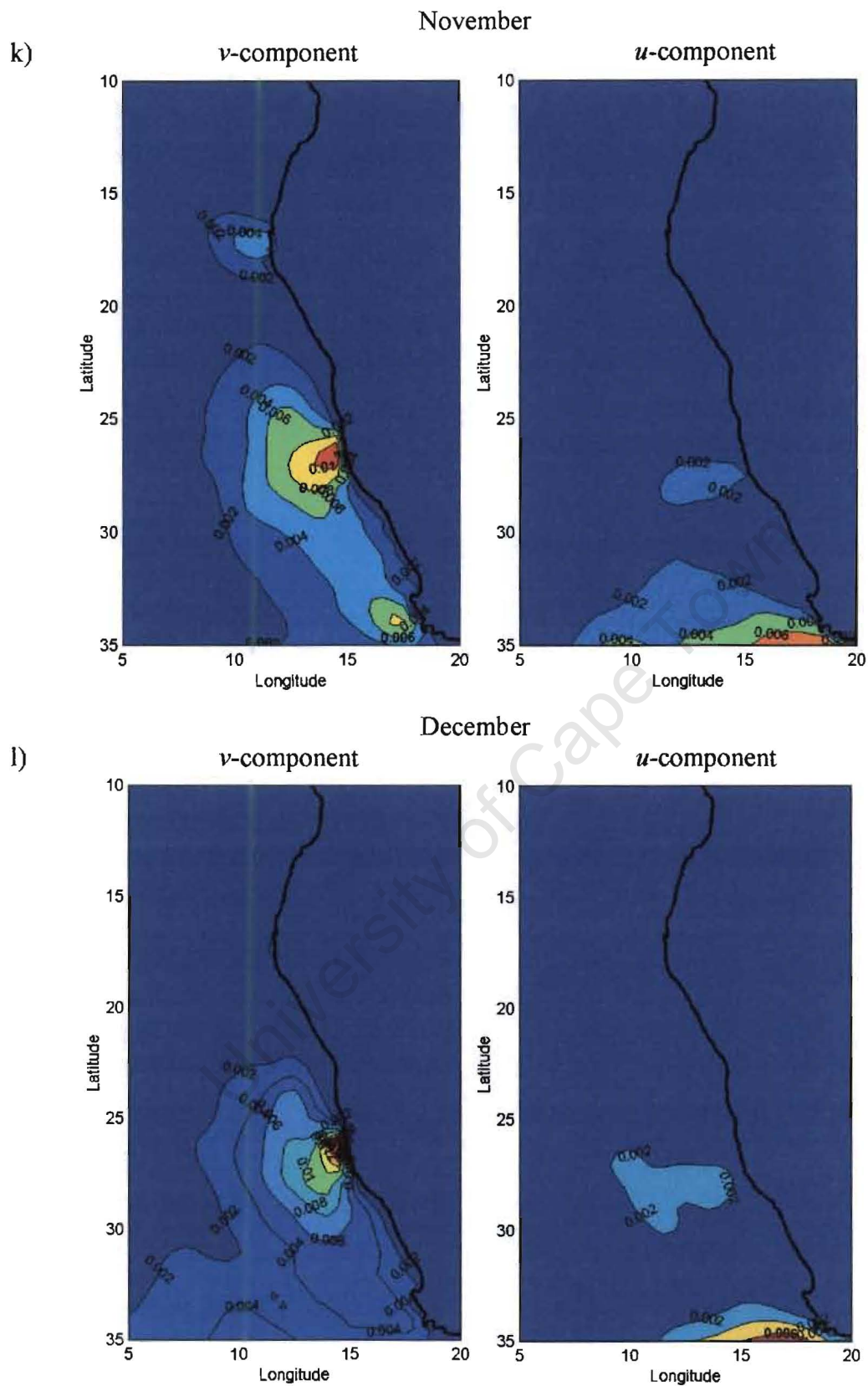
Appendix 1 (Continued) (e) May, (f) June.



Appendix 1 (Continued) (g) July, (h) August.

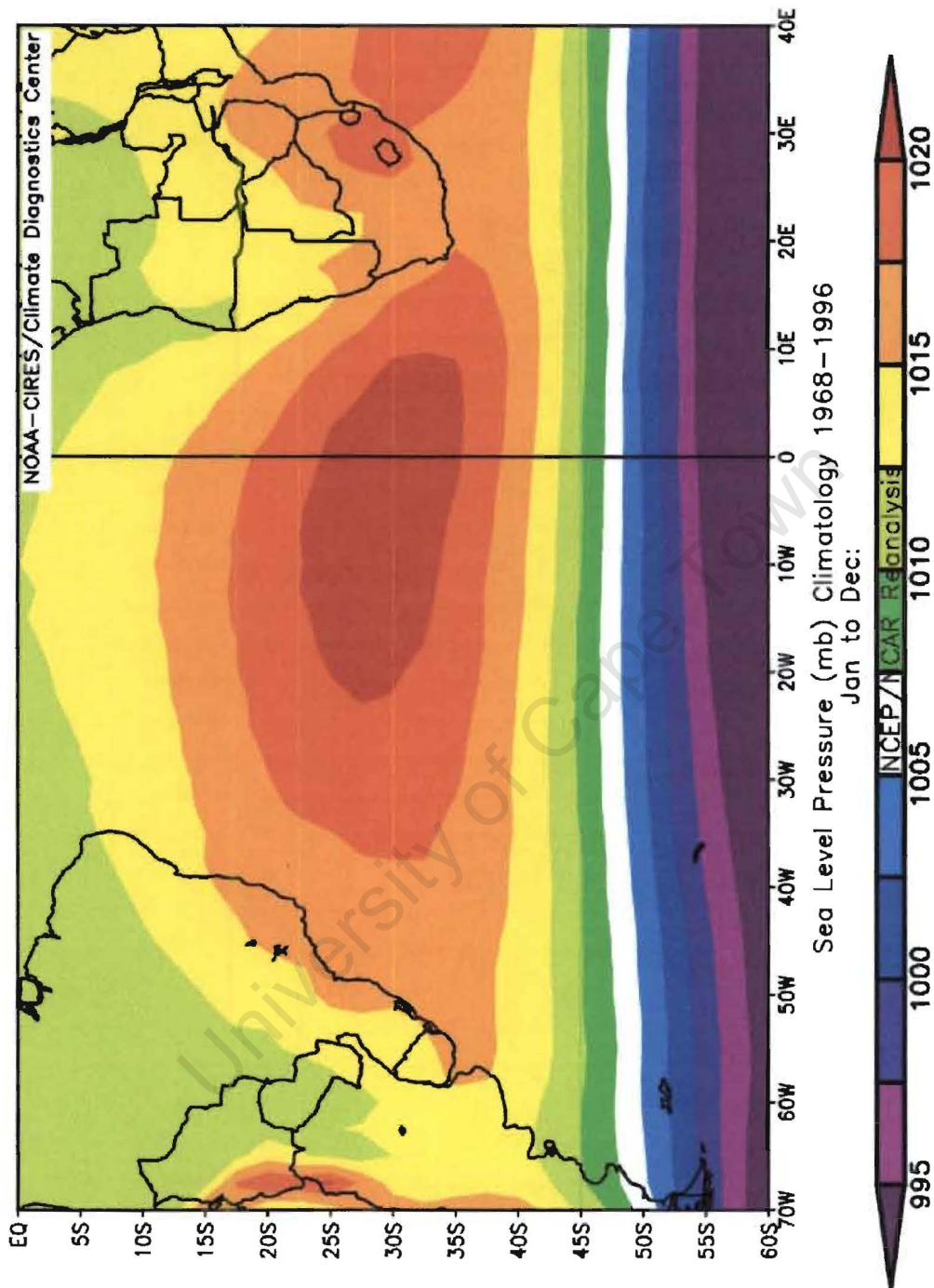


Appendix 1 (Continued) (i) September, (j) October.

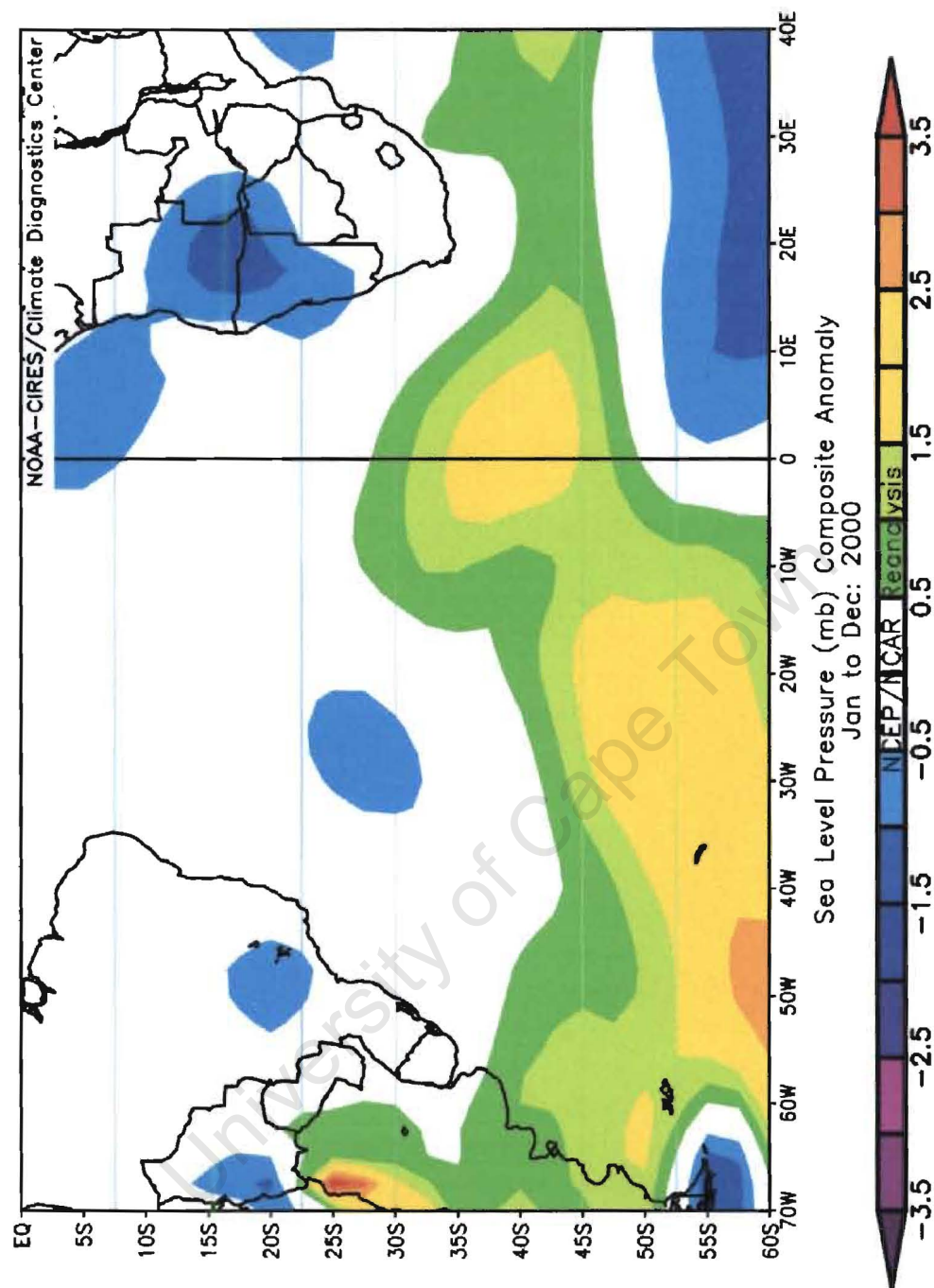


Appendix 1 (Continued) (k) November, (l) December.

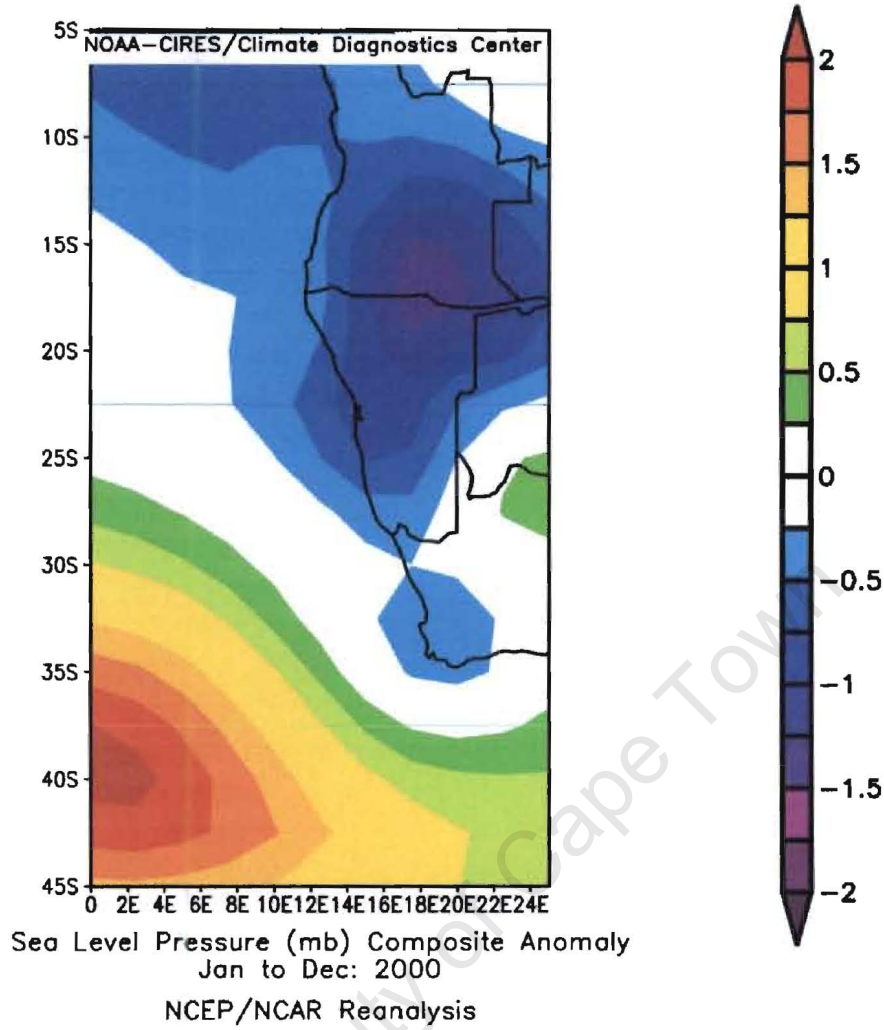
APPENDIX 2



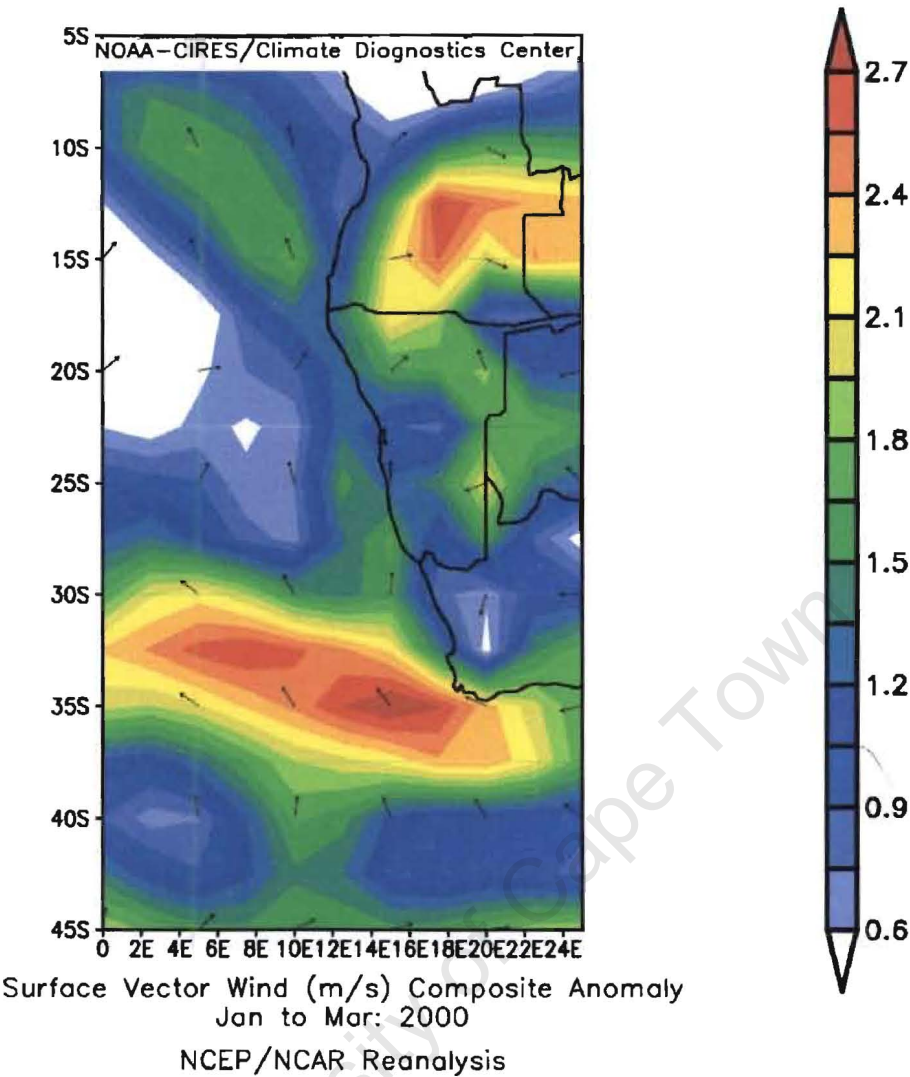
Appendix 2 The long-term climatology (1968–1996) of sea level pressure for the south Atlantic Ocean, as derived from the 2.5° resolution NCEP/NCAR reanalysis dataset².



The composite anomaly (January – December 2000) of sea level pressure for the south Atlantic Ocean, as derived from the 2.5° resolution NCEP/NCAR reanalysis dataset².

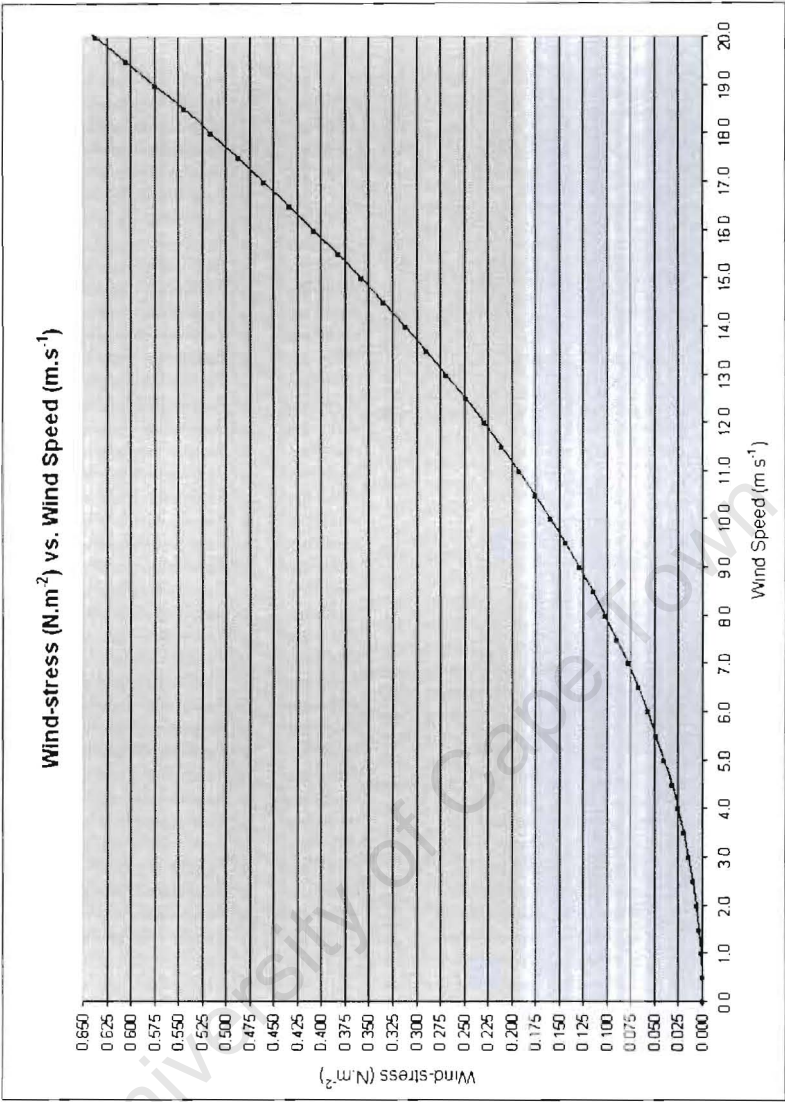


The composite anomaly (January – December 2000) of sea level pressure off the coast of southern Africa, as derived from the 2.5° resolution NCEP/NCAR reanalysis dataset².



The composite anomaly (January – March 2000) of the surface vector wind speed field off the coast of southern Africa, as derived from the 2.5° resolution NCEP/NCAR reanalysis dataset².

APPENDIX 3



Appendix 3 Graph of wind-stress (N.m⁻²) vs. wind speed (m.s⁻¹). The vector wind-stress $\underline{\tau}$ and 10-m neutral stability vector wind $\underline{V}_{10}$ are related using; $\underline{\tau} = \rho C_d \underline{V}_{10} |\underline{V}_{10}|$, where $\underline{\tau}$ is the stress vector, ρ is the density of air, C_d is a dimensionless drag coefficient, $\underline{V}_{10}$ is the wind velocity, and $|\underline{V}_{10}|$ is the wind speed. Following Bakun and Nelson (1991), the parameters ρ and C_d are held constant at 1.224 kg.m⁻³ and 0.0013 respectively.