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Water vapour transport from
the South Atlantic and Indian Oceans
and summer rainfall in southern Africa

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Abstract

Moisture input from the South Atlantic and Indian Oceans over southern Africa is examined through zonal water vapour transport. Along the west coast, variations in intensity and latitudinal position of the South Atlantic anticyclone/modulations of the westerly flow that penetrates from the tropical Atlantic, contribute the most (about 25% and 11% of the variance explained respectively), in January-February to variability in moisture advection from the South/tropical Atlantic, thus affecting rainfall at subtropical/tropical latitudes respectively. The southern extension of the AEJ could play a role in transferring moisture from the tropics southwards during wet phases, while events related to low phases of the Southern Oscillation are marked by an eastward shift of the ascending branch of the Walker circulation, suppressing convection and thus reducing rainfall over the subcontinent. Along the east coast, modulations of moisture advection in the tropics and simultaneous changes in the midlatitude circulation altering convection processes within the SICZ (together contributing to 19% of the variance explained), are found to affect in November-December, rainfall east of the Great Rift escarpment and over central regions of South Africa respectively. Secondly, the Indian anticyclone ridging more/less over the subcontinent is found to modulate convection over southeast South Africa and moisture advection inland from the east coast in the tropics (contributing to more than 15% of variance explained), ultimately influencing local rainfall regimes in October-November. Opposite conditions generally prevail over northern Angola due to simultaneous changes in water vapour transport from the tropical Atlantic. Finally, alterations of the meridional circulation in September-October bring changes in the equatorial easterly flux from the Indian Ocean (contributing to 7% of the variance explained), modulating moisture availability at the east coast and central tropics. Together with shifts of the ascending/subsiding branch of the Walker circulation, it is found to impact rainfall regimes over central regions of southern Africa. In conclusion, the study of zonal water vapour transport may help in explaining southern African rainfall variability and thus in assessing issues such as climate predictability over southern Africa.

Declaration

I declare that this thesis is my own work, and that I have received no assistance except as acknowledged. Part of the work carried out in this thesis has been published.

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"And I saw the four winds (...) and how they position themselves between heaven and earth (...). And I saw the winds on earth which support the clouds, and I saw the paths of the angels."

The Ethiopic Book of Enoch (18:3-5)

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Chapter 1

Introduction

With large and deep convergence zones characteristic of southern African climate in austral summer, the subcontinent and particularly its tropics, is a very energetic place in regard to deep convection processes. Tropical Africa is also one of the less studied areas in terms of climate research, making seasonal predictions, weather forecasts and real-time observations of the hydrological cycle difficult. Recent extreme climatic events have had devastating consequences for many southern African countries, allowing epidemics to spread, causing famine, poverty, lack/destruction of public goods and so much more. Food security is a real issue in Africa, and in most rural areas, agriculture is of the rain-food subsistence type, making the local economy very vulnerable to climatic variations. Climate research is a tool that still needs to be developed further to understand seasonal/interannual variability and to help attain food security and self-sufficiency widely within southern Africa along with other benefits (reducing malaria, natural disasters prevention,...). Considered as the main sources of moisture for the subcontinent, the oceans also exert controls on the large scale circulation, thus impacting rainfall and the water cycle. The aim in this thesis is to describe the contribution of both the South Atlantic and Indian Oceans in southern African summer rainfall variability through zonal moisture transport.

1.1 Elements of southern African climate

The climate of southern Africa has been well documented in other studies (*Preston-Whyte and Tyson, 1988; Lindesay et al., 1986; D'Abreton, 1992; Richard, 1993; Nicholson, 2000*), so a brief overview will be given here.

The most important factor affecting low land temperatures is the orography, which also strongly influences regional rainfall patterns (*Preston-Whyte and Tyson, 1988; D'Abreton, 1992; Fauchereau, 2004*).

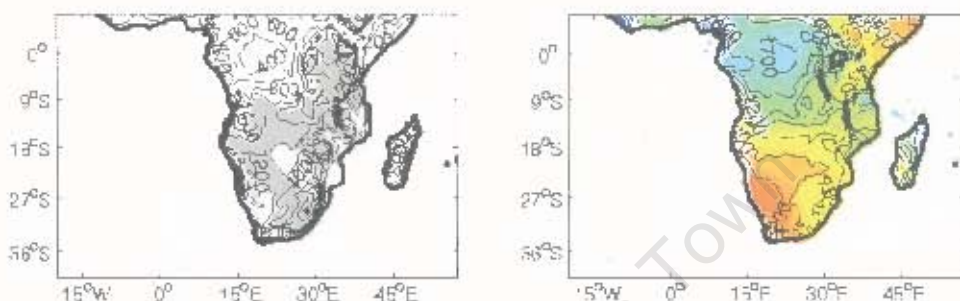


Figure 1.1: Altimetry map of southern Africa with shaded heights above 1000 m (left), and annual mean CRU rainfall over the 1979-2000 period (right).

The overall altimetry of southern Africa and annual mean rainfall are shown in Figure 1.1. Generally rainfall is marked by both zonal and meridional gradients. Highest precipitated volumes are found in the equatorial regions (2000 mm/yr over the Guinea Gulf and 1800 mm/yr over the Congo basin) and a north-south gradient in the isohyets in the intertropical zone illustrates the reduction of rainfall from the Equator to the tropics. In the tropics, most regions have two rainy and one dry season, reflecting the association of rainfall with the migration in position of the Intertropical Convergence Zone (ITCZ), the most important source of variability in southern African climate (*Preston-Whyte and Tyson, 1988*). Overall the isohyets are marked there by an east-west orientation, characterized to the west by maximum rainfall amounts over the Bie plateau (1200 mm/yr) while to the east, the orography within the Great Rift valley results in the highest rainfall, in particular

over the Great Lakes region (2000 mm/yr over Lake Tanganyika for instance). In the subtropics, another east-west gradient shows opposition between the wetter eastern coastal regions and the drier west coast of southern Africa. Most regions of subtropical southern Africa have one rainy season, with maximum rainfall in summer (from November to April). An exception to this is the Western Cape which has a more mediterranean climate, with most rainfall occurring in austral winter (maximum in June).

The second factor affecting southern African climate is the unique oceanographic setting. Strong subtropical gyres related to the surface wind pattern are found in the South Atlantic and Indian Oceans. As a result, southern African climate is affected in the midlatitudes by the cold Benguela Current upwelling system (with waters below 15°C all year around) along the west coast of the subcontinent, and the warm western boundary Agulhas Current to the east (with temperatures varying between 22°C and 28°C in summer). South of 40°S, the Antarctic circumpolar current (ACC) links the South Atlantic and southwest Indian Oceans, transporting huge quantities of cold water eastwards around the globe. On the other hand, large portions of the low-latitude oceans are characterized by zonal currents forming the equatorial currents system. It consists of the westwards flowing North Equatorial Current and South Equatorial Current system (NEC and SEC respectively) which enclose the eastward flowing Equatorial Counter Current (ECC). In the Indian Ocean, such a structure prevails in winter, while in summer the ECC disappears due to the onset of the southwest monsoon. Instead, the so-called Somali Current flows northeastwards along the east African coast, driving coastal upwelling off the Horn of Africa and the Arabian Peninsula. Between the two monsoons, prevailing westerly winds in the equatorial belt give rise to the eastward flowing equatorial surface jet, stronger to the north than to the south of the Equator. Furthermore, the equatorial belt is the region where interannual variations in the easterly winds regime create a situation favorable for wave propagating phenomenon, namely Kelvin and Rossby waves, which can modify the thermocline within the "equatorial wave guide", and

thus play an important role in local interannual climate variability.

Globally, the neighbouring oceans influence air temperatures as well as the moisture budget over southern Africa (*Preston-Whyte and Tyson, 1988; D'Abreton, 1992; Lindesay, 1998*) through heat exchange across the ocean-atmosphere interface. This involves both radiations (short and long wave) and turbulent flux processes (sensible and latent), latent heat transfer (corresponding to evaporation) representing the largest heat loss from the ocean.

Finally, due to the geographical location of the subcontinent, southern Africa climate is influenced by tropical, subtropical and midlatitude pressure systems. A schematic representation of the mean surface pressures and winds in January and July for southern Africa is given in Figure 1.2.

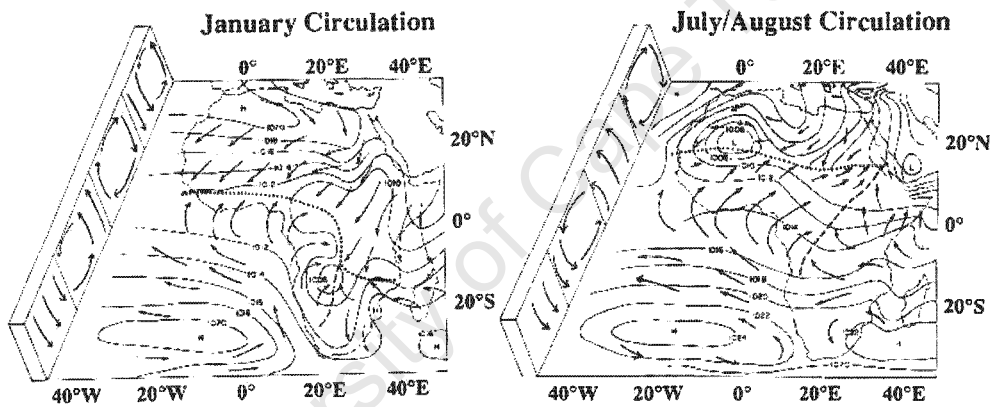


Figure 1.2: Schematic representation of the mean surface pressure systems in January (summer) and July/August (winter) (after *Nicholson, 2000*)

The quasi-stationary anticyclones correspond to the subsiding branch of the meridional circulation (so-called Hadley circulation) and control to a large extent the climate over the subcontinent. They form the so-called semi-permanent subtropical high pressure belt (*D'Abreton, 1992*) which varies in position throughout the year following a semi-annual cycle, referred to as the semi-annual oscillation or SAO. The SAO is reported to explain more than 70% of the annual variance in the surface pres-

sure field (*Preston-Whyte and Tyson, 1988*). The SAO arises due to the differential warming between the Southern Ocean and the Antarctic continent, the dephasing between the two annual temperature cycles resulting in a strong semiannual component in the meridional gradients of temperatures between the two regions. *Walland and Simmonds (1999)* have shown that such modulations were linked to the vertical static stability in the atmosphere, thus producing surface pressure minimums in Spring while meridional pressure gradients are maximum in Autumn. Within the annual cycle, the position of the South Atlantic anticyclone has a variations of about 6° in latitude between February and June (when it is at its northernmost location) and approximately 13° in longitude from October to August (reaching its westernmost position). The South Indian high pressure system on the other hand is characterized by larger annual meridional variations of about 24° between July and December (corresponding to its easternmost location) while it shifts of approximately 5° in latitude from February to October (when it reaches its northernmost location). The prevailing midlatitude circulation consists of strong westerly winds associated with the pressure gradient between the subtropical high pressures and the Antarctic circumpolar trough to the south. The variations within the subtropical anticyclone belt is reproduced to the south by the shifting of storm tracks further north in summer than in winter, and midlatitude disturbances are reported to play a substantial role in southern Africa climate (*Preston-Whyte and Tyson, 1988*). The trades emanate from the eastern flanks of the subtropical high pressure systems and form the lower-tropospheric portion of the tropical Hadley circulation. Over the trade winds region, evaporation is found to exceed precipitation, and the moisture accumulates at lower tropospheric levels below a permanent, continuous and extremely stable layer so-called the trade inversion. As a result, the moisture from the tropical oceans regions is carried by the trade winds airstream into the equatorial trough zone, and thus the trade winds trajectory substantially impacts surface climate over the subcontinent. The equatorial trough zone, located between the high pressure belts of the two hemispheres near the Equator, constitutes the ascending branch of the respective north/south Hadley cells balancing the net radiative energy

between tropical and subtropical latitudes. This is fulfilled by the import of moisture from the trade winds airstream at lower levels and export of geopotential energy and sensible heat in the upper troposphere. Low surface pressures roughly coincide with highest surface temperatures which suggests that the equatorial trough is thermally induced (*Hastenrath, 1985*).

At low tropospheric levels, the meridional circulation is not so strong. Along the western side of southern Africa, a southward component prevails all year around. Along the eastern side, the meridional circulation has a southward/northward orientation north/south of southern Africa in winter. In summer the southward shift of the subtropical high pressure belt supports a southward component south of the tropics while a northward component is part of the Indian monsoon to the north. The zonal circulation (so-called Walker circulation) plays a more substantial role helping to redistribute energy within the equatorial band (*Hastenrath, 1985*). North of 12°S the trades are dominant along the Atlantic coast in winter while they are recurved in an east-west direction in summer. To the south, the South Atlantic trades prevail and south of 32°S, the midlatitude westerly flow, consisting of weak surface winds, is a dominant feature in winter. In summer this flow is replaced by the trade winds regime. Along the Indian coast, north of 15°S, the subcontinent is subject to the monsoon regimes reversal. During northern winter, the air over southern Asia is cooler and denser than over the west Indian Ocean leading to the establishment of a strong pressure gradient which gives rise to the northeast monsoon. Crossing the Equator, it is deflected by the Coriolis force into the southeast trades between 10°S and 20°S. As the year progresses and the Asian landmass heats up, the anticyclonic pressure weakens: a trough develops by June leading to a southwesterly wind regime over the region until September. A major feature of the southwest monsoon regime is the East African low level Jet (EAJ), or Somali Jet (with speed maximas at about 10°S), while strong westerly winds at equatorial latitudes are characteristic of the inter-monsoonal transition period. To the south, easterly winds prevail and recurve while passing south of Madagascar within the eastern escarpment in South Africa. South of 32°S, they connect with the recurved circulation linked with the

South Indian anticyclone. Further south the strong circumpolar westerly circulation is found, where midlatitude depressions and storms develop.

At midtropospheric levels (from 700 mb to 500 mb), the mean circulation is essentially zonal. It consists of easterly/westerly flows within tropical/subtropical latitudes of southern Africa, referred to as the southern extension of the African Easterly Jet (AEJ) and the westerly subtropical jet respectively.

In austral summer the landmass warms up, thermal low pressure systems and convective mechanisms develop over the subcontinent, from the Equator to 20°S, within the so-called Intertropical Convergence Zone (ITCZ) where the recurved southwesterly trades, the northeast monsoon and the tropical easterlies from the Indian Ocean converge. Over oceanic regions, it is characterized by maximums in sea-surface temperature, cloudiness and rainfall coinciding with a trough in surface pressures and a reduction in winds (*Hastenrath, 1985*). The ITCZ is the area where most latent and sensible heat is released and thus play a major role in southern African climate variability (*Preston-Whyte and Tyson, 1988*). In austral winter, solar heating decreases over the subcontinent, the ITCZ migrates north of the Equator and convection is suppressed over the subcontinent. Another convergence zone, referred to as the Zaire Air Boundary in the literature (*Taljaard, 1972; D'Abreton, 1992; Nicholson, 2000*), is defined as the area where the tropical easterlies from the South Indian Ocean and the recurved westerlies from the South Atlantic Ocean meet, typically over southern Angola/northern Namibia. Very little is reported on this feature and opinions are divided. Nevertheless this region is an area of high latent heat release through the formation of convective low pressure cells. Recent works (*Cook, 2000*) have suggested a less permanent convergence zone within the eastern regions of the subcontinent and the southwest Indian Ocean, namely the South Indian Convergence Zone (SICZ). Synoptic systems developing locally have been termed tropical temperate trough (TTTs). They form when a tropical disturbance (easterly low or wave at surface) combines with a westerly wave or a low in the upper atmosphere, thus establishing a connection between the tropics and midlatitudes (*D'Abreton, 1992*) and leading to the formation of

major ^{NW / SE} northeast/southwest orientated cloud bands in conjunction with the tropical temperate trough. With strong convergence/divergence at surface/midtropospheric levels TTTs serve to transfer energy and momentum poleward, which is important for the maintenance of the Hadley circulation (*D'Abreton*, 1992; *Lindesay*, 1998). TTTs have been found to be the most important contributors to rainfall over the interior of South Africa and Zimbabwe during summer (*Harrison*, 1984; *Todd et al.*, 2002).

1.2 Interannual rainfall variability

Variability in southern African rainfall has been a subject of increasing attention in previous studies. In addition to the importance of the summer season (November–April), rainfall variability over the subcontinent is globally characterized during this period by a dipole pattern, with large positive/negative loadings over subtropical regions south of 15°S and eastern Africa. This structure, generally extracted as the first mode of variability in rainfall (*Mutai et al.*, 1998) and OLR (*Jury*, 1992, 1997), may be related to shifts in the location of the SICZ (*Todd and Washington*, 1999; *Cook*, 2000; *Todd et al.*, 2002). Over eastern parts of the subcontinent, temporal aspects of rainfall variability have been investigated at interannual (*Beltrando and Camberlin*, 1993; *Kabanda and Jury*, 1999; *Mapande and Reason*, 2005; *Camberlin and Poccard*, 2001; *Kijazi and Reason*, 2005) and seasonal scales (*Ogallo*, 1989; *Camberlin and Poccard*, 2001; *Ntale and Gan*, 2003). The influence of the neighbouring southeast tropical Atlantic has also been noticed for western tropics of southern Africa: a positive correlation has been established between SST anomalies off Angola, latent heat flux and rainfall at the coast (*Hirst and Hastenrath*, 1983; *Rouault et al.*, 2003). For central regions of southern Africa, earlier studies examined seasonal and interannual variability in rainfall (*Mulenga*, 1998; *Mwale et al.*, 2004; *Hachigonta and Reason*, 2006) and potential relationships with SSTs in the Atlantic and Indian Oceans have been emphasized (*Todd and Washington*, 2004; *Hansingo and Reason*, 2006). Over southern African regions, modulations

in the atmospheric circulation and rainfall variability have been extensively studied (*Harrison, 1984; Lindesay, 1998; D'Abreton and Lindesay, 1993; Reason and Godfred-Spenning, 1998; Nicholson, 2000; Cook, 2000; Todd et al., 2002; Mulenga et al., 2003; Rouault and Richard, 2003; Usman and Reason, 2004; Landmann and Goddard, 2005; Rouault and Richard, 2005; Tennant and Reason, 2005*), and links with the South Atlantic (*Walker, 1990; Jury, 1996; Reason, 1998; Reason and Jagadheesha, 2005a; Reason et al., 2006*) as well as with the Indian Oceans (*Reason and Godfred-Spenning, 1998; Reason and Mulenga, 1999; Reason, 2001a,b; Landmann and Mason, 1999; Rouault et al., 2002; Hansingo and Reason, 2006; Washington and Preston, 2006*) have been evidenced.

The above references, amongst others, further emphasize the fact that climate variability has been less studied in tropical regions than in the subtropics, such as South African regions where it is best documented.

Atmospheric circulation patterns Generally, wet conditions over southern Africa correspond with decreased pressures over the subcontinent at both surface and midtropospheric levels while increased pressures are found over the South Atlantic Ocean (*D'Abreton, 1992*). In addition, the advection of water vapour from the southwest Indian Ocean regions occurs around the subtropical anticyclone off the east coast, thus the modulations in surface pressures over the basin could be of importance (*Preston-Whyte and Tyson, 1988; Taljaard, 1986, 1987*). The Angola low is found to be of particular significance to rainfall variability for tropical and subtropical regions to the west of the subcontinent, also suggesting the importance of the moisture inflow from the tropical southeast Atlantic (*Mulenga, 1998; Cook et al., 2004*). *D'Abreton and Lindesay (1993)* report changes in zonal and meridional transport of water vapour over southern Africa during wet and dry summers, with the neighbouring oceans contributing as moisture sources. *Todd and Washington (1999)* have shown that moisture transport into southern Africa was mainly occurring along the northern/central Indian Ocean, at equatorial latitudes of the South Atlantic and southern Africa, and within the cyclonic circulation

prevailing over the Mozambique channel in summer. *Todd et al.* (2002) present an analysis of tropical temperate troughs (TTTs) in terms of moisture fluxes and find that such dominant rainfall producing systems support meridional transport of water vapour from the tropics to midlatitudes of southern Africa. *Mapande and Reason* (2005) found that wet and dry years over Tanzania were corresponding with weaker/stronger equatorial westerlies linked to anticyclonic/cyclonic anomalies at tropical latitudes of the western Indian Ocean, together with enhanced/reduced westerly moisture input from the tropical Atlantic. Dry conditions over Zambia are found to be linked with low level easterly anomalies and a reduction in moisture input from the southeast Atlantic, while wetter periods correspond with westerly anomalies over south Angola/west Zambia (*Hachigonta and Reason*, 2006). It is worth noting that an increase in the frequency of heavy rainfall has been evidenced for the last 20 years over Angola/Namibia and Tanzania/Mozambique (*Usman and Reason*, 2004).

Teleconnections with the neighbouring oceans Rainfall variability in southern Africa is known to be influenced by sea-surface temperatures (SSTs) in the main oceanic basins (*Reason and Godfred-Spenning*, 1998; *Reason and Mulenga*, 1999; *Richard et al.*, 2001; *Camberlin and Poccard*, 2001; *Fauchereau et al.*, 2003; *Mwale et al.*, 2004). The South Atlantic has been regarded for a long time as a secondary source of moisture over the subcontinent (*Preston-Whyte and Tyson*, 1988; *D'Abreton and Lindesay*, 1993), but recent studies (*Cook et al.*, 2004; *Reason and Jagadheesha*, 2005a; *Reason et al.*, 2006) have emphasized its importance in terms of southern African climate variability. Generally, SST patterns in the South Atlantic are found to be connected with corresponding anomalous atmospheric circulations, affecting southern African climate (*Walker*, 1990; *Mason*, 1990, 1995; *Rocha and Simmonds*, 1997; *Reason*, 1998) and thus leading to rainfall variability in neighbouring countries of southern Africa (*Hirst and Hastenrath*, 1983; *Nicholson and Entekhabi*, 1987; *Camberlin and Poccard*, 2001; *Rouault et al.*, 2003; *Robertson et al.*, 2003; *Cook et al.*, 2004; *Tennant and Reason*, 2005; *Reason et al.*, 2006).

Mason (1995) found evidences that anomalous warm SSTs in the subtropical South Atlantic and to the south of the subcontinent would potentially enhance cyclogenesis. Such changes in the surface ocean could also imply a poleward shift of the midlatitude westerly circulation, associated with wetter conditions over southern regions of the subcontinent, and are thus of substantial importance in regards to southern African climate. Model studies further support the existence of significant atmospheric responses to SSTs variability in the tropical and subtropical South Atlantic (*Robertson et al.*, 2003; *Reason and Jagadheesha*, 2005a). Previous works (*Zebiak*, 1993; *Carton and Huang*, 1994; *Venegas et al.*, 1997; *Chang et al.*, 1997; *Ruiz-Barradas et al.*, 2000; *Sterl and Hazeleger*, 2003; *Colberg*, 2006; *Colberg and Reason*, 2006) have shown the existence of dominant modes of variability within the ocean-atmosphere system over both tropical and South Atlantic regions, which could be of substantial importance to southern African climate.

With warmer waters, the Indian Ocean tends to be considered as the main source of moisture over the subcontinent. Links between southern African climate variability and SSTs in the southwest Indian Ocean have been identified in previous studies (*Walker*, 1990; *Mason*, 1990, 1995; *Reason and Godfred-Spenning*, 1998; *Reason and Mulenga*, 1999; *Landmann and Mason*, 1999; *Reason*, 2001a,b; *Rouault et al.*, 2002; *Washington and Preston*, 2006), some referring to mechanisms linked with meridional and zonal SST gradients in the basin. *Reason and Mulenga* (1999) have shown that warm/cold SSTs in the Indian Ocean are associated with wetter/drier conditions over South Africa through modulations in the convergence of moist air from the Indian Ocean and tropical regions of southern Africa. Using a regional model, *Hansingo and Reason* (2006) found that anomalous warm SSTs in summer over the southwest Indian Ocean regions were resulting in enhanced easterly moisture inflow over southern Africa where above-normal rainfall can be observed. Generally, high surface temperature anomalies to the north of Madagascar are found to lead to dry conditions over southern Africa, through the development of tropical easterly disturbances over equatorial Indian Ocean regions rather than on land (*Mason*, 1995). The same author has shown that anomalously warm SSTs to the south of Madagas-

can result in sensible and latent heat flux anomalies in the tropical easterly inflow and temperate systems leading to wet conditions over summer rainfall regions. A meridional dipole structure has been evidenced through multi-model simulations as most important for southern African rainfall in early austral summer (*Washington and Preston, 2006*): warming/cooling to the south/north of the southwest Indian Ocean is found to correspond with wet conditions over southern Africa while an opposite situation would lead to below-normal rainfall. In their experiments the authors found that cold anomalies to the north of Madagascar could lead on their own to anomalous anticyclonic circulation responsible for low level moisture input at tropical latitudes over eastern regions of southern Africa. Furthermore, ocean conditions related to an east-west SST dipole have been shown to have substantial impacts on rainfall over southern Africa (*Ogallo, 1988; Beltrando and Camberlin, 1993; Richard et al., 1998; Behera and Yamagata, 2001; Behera et al., 2005*). AGCM studies have evidenced links between enhanced summer rains in southern Africa and warm/cold SSTs to the east/west of the Indian Ocean basin through enhanced evaporation to the south of Madagascar resulting in increased advection of moisture over southeastern regions of the subcontinent (*Reason, 2001a*). Recently, anomalous events (1994 and 1997) shed light on ocean-atmosphere variability in the tropical Indian Ocean termed as the Indian Ocean Dipole (*Saji et al., 1999*) or Indian Ocean zonal mode (*Webster et al., 1999*), which could substantially affect rainfall regimes over eastern regions of Africa (*Saji et al., 1999*).

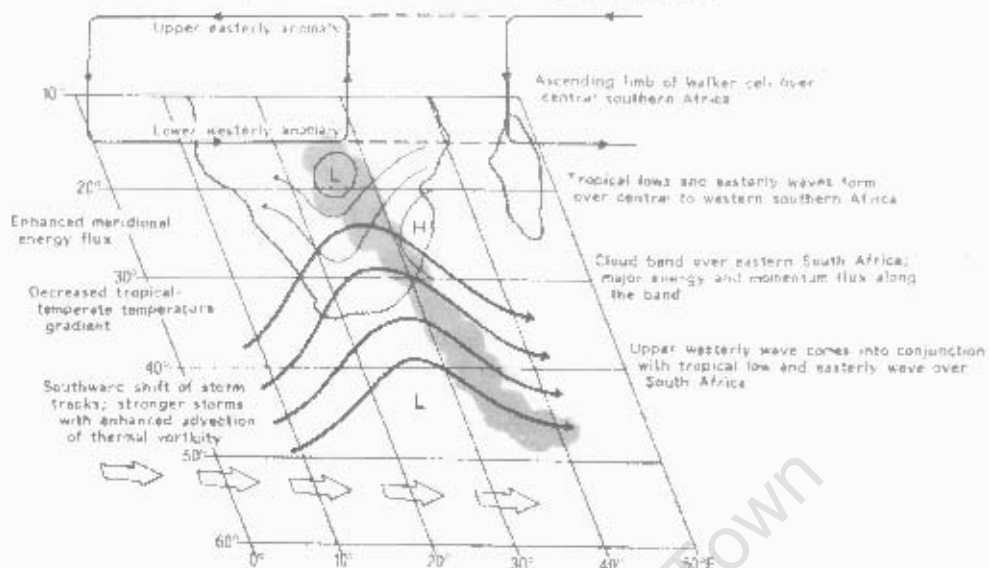
Large scale climatic modes Many studies have been focusing on large scale interactions potentially modulating the climate of southern Africa. Close relationships have been identified with the Southern Oscillation (*Lindesay et al., 1986; Lindesay, 1988; Janowiak and Xie, 1988; Nicholson and Entekhabi, 1986; Rocha and Simmonds, 1997; Makarau and Jury, 1997; Reason et al., 2000; Camberlin and Pocard, 2001; Reason and Jagadheesha, 2005b*), together with impacts on southern African rainfall (*Ropolewski and Halpert, 1987; Preston-Whyte and Tyson, 1988; van Heerden et al., 1988; Jury et al., 1994; Nicholson and Kim, 1997; Richard et al.,*

2000; Cook, 2000, 2001; Reason and Rouault, 2002; Misra, 2002; Kijazi and Reason, 2005).

For low/high phases of the Southern Oscillation, convection is reduced/enhanced over Indonesia, and due to shifts and changes in intensity, wet/dry years over southern Africa are characterized by anomalous westerly/easterly winds at low level and easterly/westerly winds aloft (Figure 1.3). During La Niña years the ascending limb of the Walker circulation is located over tropical Africa leading to strong convection. A subsequent enhanced Hadley cell seems necessary for the development of cloud bands (Lindesay, 1998). El Niño years are characterized by an eastward shift of the ascending limb of the Walker circulation with a concomitant shift of strong convection. Such changes result in a reduction of the meridional energy flux over the subcontinent while convection is increased over Madagascar during dry years over southern Africa. In this thesis we will use the consensual years listed in Appendix A to refer to high and low phases of the Southern Oscillation.

Beltrando and Camberlin (1993) identified substantial influences of the Southern Oscillation on northern autumn rains in Somalia as well as positive/negative relationships with surface pressures in the eastern/western Indian Ocean regions. Moreover, the weakening of the equatorial Walker circulation over the Indian Ocean due to ENSO has been related to abnormally wet short rains in East Africa (Camberlin and Pocard, 2001). The work of Ogallo (1989), Kabanda and Jury (1999), Reason *et al.* (2000) and Kijazi and Reason (2005) have also emphasized further influences of ENSO on the climate over eastern regions of the subcontinent. Links between ENSO events and a north-south dipole in the southwest Indian Ocean has been evidenced in earlier studies involving surface pressures, SSTs and winds changes (Reason and Godfred-Spenning, 1998). Some authors also concluded that SST anomalies in the Indian and Atlantic Oceans are essential in producing rainfall response to ENSO over southern Africa (Nicholson and Kim, 1997; Nicholson, 1997, 2003), while other studies suggest a direct and purely atmospheric mechanism independent of the Indian and Atlantic Oceans (Cook, 2000). Cook (2000) describes this later mechanism as involving Rossby wave generation associated with SST anomalies in

HIGH PHASE southern African rainfall above normal



LOW PHASE southern African rainfall below normal

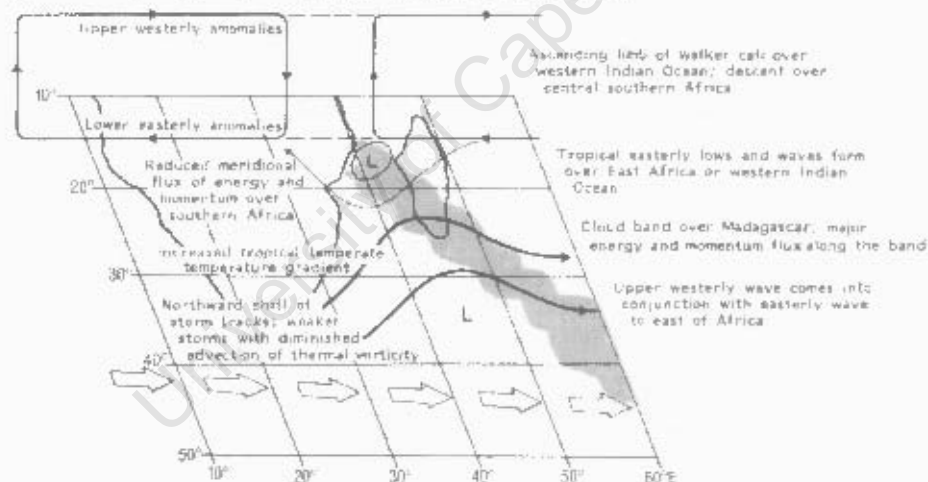


Figure 1.3: The Walker circulation during the low/high phase of the Southern Oscillation (top/bottom) (after Harrison, 1984)

the Pacific resulting in the northeastward shift of the SICZ. Despite the disagreement between these two proposed mechanisms, a rainfall dipole with anomalously wet conditions over eastern tropical Africa and drought over southern Africa, is typical of rainfall response to ENSO (*Ropolewski and Halpert, 1987; Nicholson and Kim, 1997; Nicholson, 2003*).

In addition, rainfall variability over the subcontinent have been shown to result from adjustments to planetary scale circulation patterns (*Todd and Washington, 1999; Reason et al., 2002*). *Todd and Washington (2003, 2004)* evidenced influences of the large scale atmospheric circulation in the North Atlantic region on central equatorial African climate variability during boreal winter/spring, through anomalous westerly midtropospheric moisture inflow over these regions. Furthermore, recent studies (*McCabe and Palecki, 2005*) investigating multidecadal climate variability have shown that the main mode of low frequency changes in sea-surface temperatures in the Pacific Ocean, the Pacific Decadal Oscillation or PDO (*Mantua et al., 1997*), could have substantial impacts on global climate and thus could potentially affect rainfall variability over southern Africa. The Madden-Julian oscillation or MJO (*Madden and Julian, 1971*), an intra-seasonal oscillation (with a period of 30-60 days) of the atmospheric circulation and tropical convection propagating from the equatorial latitudes of the Indian Ocean towards the Pacific warm pool, is also found to influence eastern (*Pohl and Camberlin, 2006a,b*) and southern African rainfall (*Pohl et al., 2007*). *Reason and Rouault (2005)* have emphasized links between rainfall regimes in South Africa and fluctuations in the Antarctic oscillation (AAO) which is identified as the leading principal component of the 850 mb geopotential heights south of 20°S (*Thompson and Wallace, 2000*). Moreover, the Quasi-Biennial Oscillation in equatorial winds or QBO (*Angell and Korshover, 1964*), linked to downward propagating westerly and easterly zonal wind regimes in the equatorial stratosphere, is also reported to affect rainfall regimes over the subcontinent (*Mason and Tyson, 1992; Mason and Lindesay, 1993; Nicholson and Kim, 1997*). The modulation of the subtropical anticyclonic belt is definitely another element of primary importance, and evidences of synchronous variability in

summer within the ocean-atmosphere system formed by both basins (*Fauchereau, 2004; Hermes and Reason, 2004*) would further support large scale influences within climate variability in southern Africa.

In regards of the above cited literature, still few studies have tried to quantify water vapour transport from the neighbouring oceanic basins onto the subcontinent and this will be examined in this document. The relationships between primary modes of variability within each basin regions and rainfall variability over southern Africa will be further investigated. An approach at the scale of the subcontinent would have the advantage to address such issues for both tropical and subtropical regions, and it is thus expected to give further elements of description in particular regarding the tropics which have been less studied.

1.3 Objectives and approaches

The aim of this study is to investigate the role of the South Atlantic and Indian Oceans in modulating rainfall regimes over southern Africa. Given the importance of the summer season, this will be the main period of interest. The questions and hypotheses that will tackled in this thesis are as follow,

- (i) What are the main characteristics of moisture transport over southern Africa?
- (ii) What are the main modes of variability in the South Atlantic and Indian Ocean regions?
- (iii) How do the neighbouring oceans contribute in southern African summer rainfall variability through zonal moisture transport? What are the dynamics and convective mechanisms involved? Are there any links with the ocean-atmosphere systems in each basin?

The corresponding chosen approach can be divided into the following points,

- (i) The key components of water vapour transport over southern Africa are investigated by computing moisture fluxes and convergence. The contribution from each oceanic basin to water vapour over southern Africa is assessed, averaging zonal moisture fluxes along both the west and east coasts.
- (ii) Multivariate analyses are applied to each basin regions in order to extract primary modes of variability prevailing in the respective ocean-atmosphere systems. These modes are used as a basis to describe connections with water vapour transport variability over southern Africa,
- (iii) Dominant modes of variability are extracted for zonal moisture fluxes along the west and east coasts to help identify how moisture input from the neighbouring basin can modulate processes related to convection and thus southern African summer rainfall.

The previously mentioned approaches correspond with the overall structure of this document. Presenting the different data and methodology used, a description of the characteristics of water vapour transport and convergence over southern Africa will be given in the next chapter. Following this, ocean-atmosphere variability in the neighbouring oceanic basins is investigated. Finally the contribution of each ocean to southern African rainfall regimes in summer is considered through zonal moisture fluxes variability along the east and west coasts. Atmospheric processes entering into play as well as potential connections with the ocean-atmosphere variability in both basins are then examined. Conclusions and discussions are gathered in the last section of this thesis.

Chapter 2

Data and Methods

The following section outlines the different data sources and analyses used in this thesis.

2.1 Data sources and characteristics

2.1.1 Precipitation

A substantial limitation to the study of rainfall in Africa is the insufficient number of gauge observations available. Limited and incomplete records over some parts of the continent do not permit a complete description of the local and regional patterns of interannual rainfall variability.

As a reference in this study, rainfall data from the Climate Research Unit at the University of East Anglia (U.K.), CRU TS 2.0 dataset was used. This global gridded dataset based on surface station observations, provides monthly precipitation from 1900 to 2001 at a 0.5 degree resolution. More details can be found in *Mitchell et al.* (2004).

In order to gain confidence in the data, other rainfall estimates with both land (CHARM, GPCC and CAMS-OPI) and global (GPCP, CMAP) coverage were considered. The representation of key features of southern African climate are compared. More details can be found about these datasets in Appendix B.

In regards of climatological means and given the importance of the summer season (November-April), the 1000/500 mm contour lines were used to delineate wetter and more arid regions respectively, in order to examine how the different datasets reproduce wetter/drier regimes for this period. Results are presented in Figure 2.1.

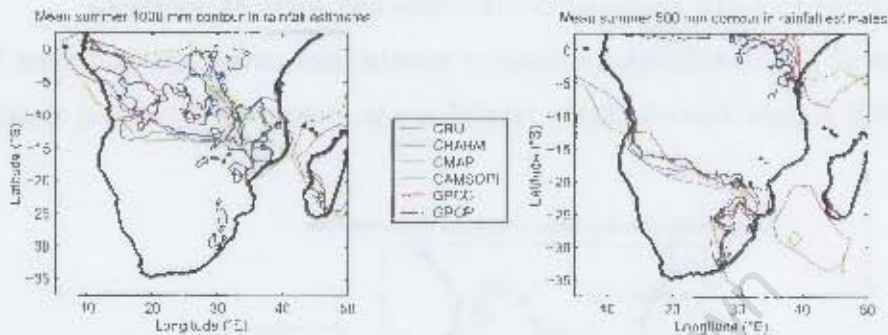


Figure 2.1: Mean 1000 mm and 500 mm contours in summer precipitation as represented by several rainfall estimates (CRU, CHARM, CMAP, CAMSOP1, GPCC, and GPCP). GPCP and CMAP give estimates above the oceans.

The distribution of the 500 mm and 1000 mm isohyets is quite consistent between each dataset although there are some differences. The spatial resolution of each dataset of course must be considered, but most discrepancies are found in the geographical extent of local extremas.

In the west, the 1000 mm contour reflects the influence of moisture input from the equatorial Atlantic, from Gabon inland across the Congo basin and southwards to the northern side of Bie plateau, agreeing with previous results (*Hirst and Hastenrath, 1983; Lindesay, 1998; Nicholson, 2000*). Generally, most differences are found over the central and north Congo as well as further east. CRU and CHARM show the strongest spatial coherence. At this seasonal scale, GPCC estimates seem to differ substantially over latitudes lying in the transition between summer rainfall areas and drier regions to the south (i.e. Bie plateau and north Mozambique).

On the other hand, most of the datasets agree on the position of the 500 mm contour line that delimits semi-arid areas in the southwest and northeast of the subcontinent. Some differences are noticeable for CMAP and CAMS-OPI along the western coast of the subcontinent. To the south and eastern interior, CRU and CHARM agree quite well compared with other datasets regarding seasonal precipitation from Botswana across to north/northeast South Africa and south Mozambique.

In terms of the contribution of summer months into annual totals, Figure 2.2 shows the 90% contour line of summer rainfall as a percentage of the annual amount.

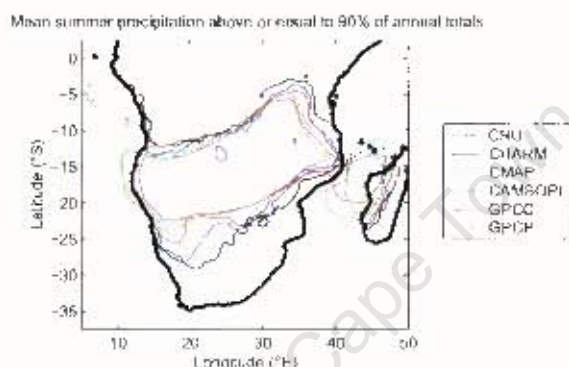


Figure 2.2: Areas of mean summer rainfall above or equal to 90% of annual totals as represented by several rainfall estimates (CRU, CHARM, CMAP, CAMS-OPI, GPCC, and GPCP).

It reveals an area common to all datasets, lying between 12°S and 30°S in the west and between 5°S and 15°S in the east. This area shows the association of rainfall to the latitudinal migration of the ITCZ and the development of the Angola low in summer. The extent of this large band varies slightly from one set to another. To the west for instance, relatively fine resolution datasets (CRU and CHARM) agree on the longitudinal gradient between the Angolan coast and regions further inland, while others do not reflect such delineation. To the south, coarser resolution datasets agree again while other estimates show differences in the amount of summer rain over the semi-arid regions south of the Bie plateau. Along the eastern coast, differences are also found over Tanzania, at the coast and further inland.

Globally, major discrepancies amongst the several precipitation estimates are found over coastal areas, wettest regions of the Congo basin, and transition zones between wetter and drier areas. Still, all datasets exhibit the importance of summer rainfall within tropical and subtropical latitudes of southern Africa.

2.1.2 Land hydrology

The variations between direct rainfall estimates is a motivation to use other datasets related to the water cycle in order to gain confidence in the rainfall data. Information concerning the vegetation cover and subsurface water supply are of great importance regarding rainfall variability. The Normalized Difference Vegetation Index (NDVI) and soil moisture are used in this study to characterize land hydrology. While NDVI is independent of rainfall but actually highly correlated at interannual time-scales, soil moisture is less independent of gauge data as an output of models driven by observed rainfall. The corresponding datasets are described in more detail in the following. In addition, a comparison of mean annual rainfall, annually integrated NDVI and annual monthly mean soil moisture for the summer season is presented in Appendix C. The results confirm substantial coherence between all datasets in reproducing most of extreme wet/dry summer conditions which have had devastating impacts for the populations and the economy of the southern African countries concerned.

2.1.2.1 Normalized Difference Vegetation Index

The principal remote sensing tool for monitoring primary productivity, and for providing an early warning of drought, is the Normalised Difference Vegetation Index (NDVI) (*Tucker, 1979*), derived from Advanced Very High Resolution Radiometer (AVHRR) data from the National Oceanic and Atmospheric Administration (NOAA) polar orbiting satellites. NOAA-AVHRR satellites combine a good time and spatial resolution with at least two passages per day and one kilometre at the nadir. The AVHRR radiometer has 5 channels sensible to different bands of the electromagnetic spectrum. More details can be found in Table 2.1.

AVHRR Channel	Wavelength (μm)	Frequency Domain
1	0.58-0.68	Visible
2	0.72-1.101	Near infra-red
3	3.53-3.93	Medium infra-red
4	10.3-11.3	Thermic infra-red
5	11.5-12.5	Thermic infra-red

Table 2.1: NOAA-AVHRR channels and their respective spectral band and frequency domain.

The study of vegetation using remote sensing information is done by measuring the reflectance of an actual green healthy plant, which is function of the wavelength. Typical reflectance curves for vegetation, soil and water are shown in Figure 2.3.

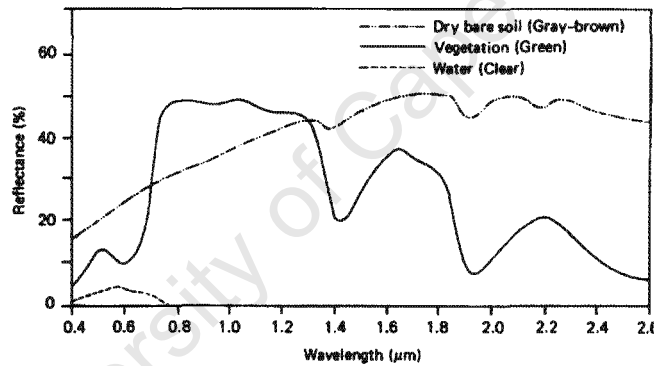


Figure 2.3: Typical reflectance curve for vegetation, soil and water (after *Lillesand and Kiefer*, 1994)

For typical green vegetation, the reflectance curve offers a characteristic shape with low values in the short wavelengths (ultra-violet and visible until $0.50 \mu\text{m}$), and a relative maximum in the green ($0.55 \mu\text{m}$) before it decreases in the red to reach a level of high reflectance (more than 30%) in the near infrared. These foliage optical properties depend essentially on the cellular structure, as most of the incoming solar radiation is absorbed by the foliate pigments (*Guyot*, 1984). The chlorophyll is the

essential pigment and shows two absorption bands, one in the blue and another in the red, which brings this maximum reflectance in the yellow-green at 550 nm. On the other hand, the maximum reflectance in the near infrared is very dependent on the leaves' anatomic structure. Greater than 1.3 μm , in the medium infrared, the spectral behaviour of plants is essentially dependent on their water content. These spectral vegetation characteristics can be modified by the soil type, the density of the actual vegetal cover, shooting angles and lightening conditions (*Malingreau*, 1989).

The NDVI has been developed by *Rouse et al.* (1973) and this approach was first used by *Tucker* (1986). As explained previously, the chlorophyll absorbs in the red (AVHRR channel 1, 0.58-0.68 μm) and foliage reflects light in the near infrared (AVHRR channel 2, 0.72-1.10 μm). Therefore, higher photosynthetic activity will result in lower reflectance in the red channel and higher reflectance in the near-infrared channel: this signature is unique to green plants (see Figure 2.3). Combining the two channels as in Equation (2.1) allows the response to distinguish the vegetation growth from the background signal. The typical formula for NDVI is given by:

$$NDVI = \frac{NIR - RED}{NIR + RED} \quad (2.1)$$

where *RED* and *NIR* correspond to channel 1 (in the visible red) and channel 2 (near infrared) respectively.

The enlightening variations in both frequency bands, principally due to heterogeneous landscape structures and/or different shooting conditions (*Bariau et al.*, 1985), are partly eliminated in the NDVI calculation through the normalisation applied (division by the sum of both channels red and near-infrared). Consequently, typical NDVI takes values between 0 and +1, but it is negative above the sea (due to the weak reflectance of water in the near-infrared) and for some very dry soil conditions such as deserts. The emphasis on contrasts given by the NDVI formula explains why it is the most often used index.

The NDVI is actually very close to the NIR/RED ratio, and can be written as follow,

$$NDVI = \frac{r - 1}{r + 1} \quad (2.2)$$

with $r = NIR/RED$ also known as Ratio of Vegetation Index (RVI).

This index is a very straightforward method to discriminate vegetation types in a given image. Typical values for different cover types are given in Table 2.2.

Cover Type	RED	NIR	NDVI
Dense vegetation	0.1	0.5	0.7
Dry bare soil	0.269	0.283	0.025
Clouds	0.227	0.228	0.002
Snow and ice	0.375	0.342	-0.046
Water	0.022	0.013	-0.257

Table 2.2: Typical NDVI values for various cover types (after *Holben*, 1986).

Thus, in many studies it has been highlighted that the NDVI can be used to investigate the consequences of interannual climate variability on vegetation, such as droughts (*Mounier*, 1990; *Dubreuil and Du*, 1995).

Like all vegetation indices, NDVI suffers certain limitations. First, the thermal channels (4 and 5) of the AVHRR sensor have internal systems of calibration but the visible and near-infrared channels do not. Given that the sensor is degrading over time, consequent errors have been noticed in the direct estimation of NDVI over the years. Calibration formulas for NOAA-7, -9 and -11 satellites have been proposed by *Nagaraja Rao and Chen* (1995) from regression methods over the Libyan desert.

Light scattering through the atmosphere tends to increase the amount of red radiation received by the satellite as red is more readily scattered in the atmosphere than near infrared, thus reducing the NDVI values. Most of the corrections proposed, either direct methods or models, consider the actual water vapour content, as well

as aerosols and other gas concentrations in the atmosphere (*Holben, 1991*). These atmospheric effects vary with the nature of both vegetation and soil. One simple method in use consists of subtracting from the actual reflectance at each pixel the smallest value recorded, observed over marine areas, assuming that the atmospheric diffusion is a constant over the whole image. In some extreme cases of long-term, large scale aerosol events, this method will not work. This is the case for the Mt. Pinatubo event in the Philippines, where the volcano erupted explosively in June 1991. The vertical force of the eruption pushed a massive amount of materials into the stratosphere. These stratospheric aerosols had a great effect on the clarity of the atmosphere and artificially reduced NDVI values, leading to a more elaborated correction being developed by NASA.

The exact position of the sun during the passage of the satellite is known to largely influence the radiometric recorded signals. The actual value, which the sensor must output in a given wavelength depends on the cosinus of the sun zenith angle, and thus for each image the correction, function of the sun zenith angle at the date and time when the picture was taken, must be done before the NDVI calculation.

As it scans across the Earth, only one point is directly underneath the sensor, called the sub-point or nadir located at the centre of the scan. *Gutman (1987)* has shown that the orbital parameters of NOAA satellites describe a 9 day cycle. During this cycle, a single target on the Earth's surface will be viewed once a day at the nadir by the satellite, 2 days for angles greater than 40° , and the rest of the time with angles between these two extremes. *Gutman (1987)* noticed that the off-nadir effects are not negligible and depend also on the atmospheric conditions as well as the nature of the vegetal cover. *Godward et al. (1991)* estimated that above 30° from each side of the nadir, the distortion in the resulting image is such that the NDVI values are too uncertain to be considered with confidence. These off-nadir effects explain why only images taken with a similar shooting angle can be compared.

In practice, a single NDVI pixel rarely covers homogenous vegetation. A NDVI value at a given pixel consists of the sum of the radiations from all the different local

land types, and thus it reflects the general regional conditions and not necessarily that of a specific vegetation present in the area.

Moreover, NDVI is sensitive to soil types: the actual light reflected by the soil can have a considerable effects on the NDVI value. Hence, most of the regional studies using NDVI carefully consider soil and vegetation cover types.

Some other factors also come into play. For instance, the exact position of the satellite compared with its nominal orbit, as well as the non perfect spherical shape of the Earth are known to influence the geometry of the AVHRR image. There is a need to correct this non-homogeneous geometry. Methods which depend only on orbital and satellital parameters, have been developed in order to obtain an image with orthogonal geographic co-ordinates.

The dataset used in this study is the NDVI version 3 (*Myeni et al., 2002*), produced as part of the NOAA-NASA Pathfinder AVHRR Land program. It consists of global monthly composites of NDVI at 1 degree resolution covering the 1981-2001 period. This dataset has been corrected in regards of AVHRR sensor degradations. Additional corrections have been applied to minimise the effects from the atmosphere, the stratospheric aerosols and these linked with non vegetation, using the maximum value composite method.

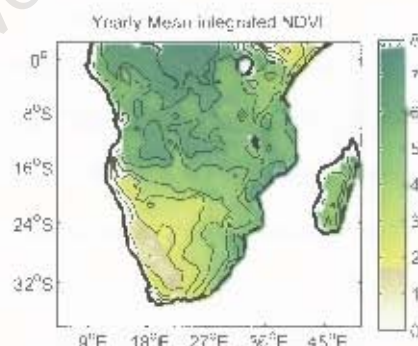


Figure 2.4: Annual mean integrated NDVI for the 1982-2000 period.

surface hydrology processes. Nevertheless, it is difficult to make representative measurements of soil moisture, and existing data is only available in few places and over very restricted time periods. Thus, methods to calculate soil moisture have been developed by integrating a land surface model forward in time over a large scale area. A land surface model typically deals with (*Fan and van den Dool, 2004*),

$$\frac{dw}{dt} = P - E - R \quad (2.3)$$

where w is the soil moisture, P precipitation, E evaporation, and R runoff. w is integrated in time using P , E and R as the main model inputs. P is far better observed than E and R which are parametrized in the models in a way that errors in R and E estimates will not cause accumulated biases in w . This generally involves other observations such as temperature, radiation and wind.

The NOAA NCEP CPC GSM dataset (*Fan and van den Dool, 2004*), which is used in this study, consists of global monthly values from January 1948 to February 2006 at 0.5 degree resolution. Soil moisture in this dataset is estimated by a one-layer hydrological model (*Huang et al., 1996; Van den Dool et al., 2003*), taking observed precipitation and temperature to calculate soil moisture, evaporation and runoff. The potential evaporation is estimated from observed temperatures. Model parameters are spatially constant and the model is tuned to runoff of several small river basins in eastern Oklahoma (resulting in a maximum holding capacity of 760mm of water). Within the CPC PRECipitation REConstruction over Land program, issues such as orographic adjustment/enhancement and inhomogeneity resulting from changes in number of gauges over time are set as the first target for iterations.

Annual means of monthly soil moisture (Figure 2.6) agree well with the distribution of rainfall totals, with maximums located over tropical latitudes, while drier conditions are found south of 15°S (with the exception of southeastern regions of South Africa). Compared to other studies (*Douville and Chauvin, 2000*), values are found to differ substantially over the Congo basin, reflecting inaccuracies that might be linked to biases in the input precipitation field. In addition, the tuning of the

Annual means of NDVI cumulated over the 12 calendar months are presented in Figure 2.4 and reflect well the variations of rainfall with greenest areas in the tropics and semi-arid conditions in the southwest and east of the Great Rift valley.

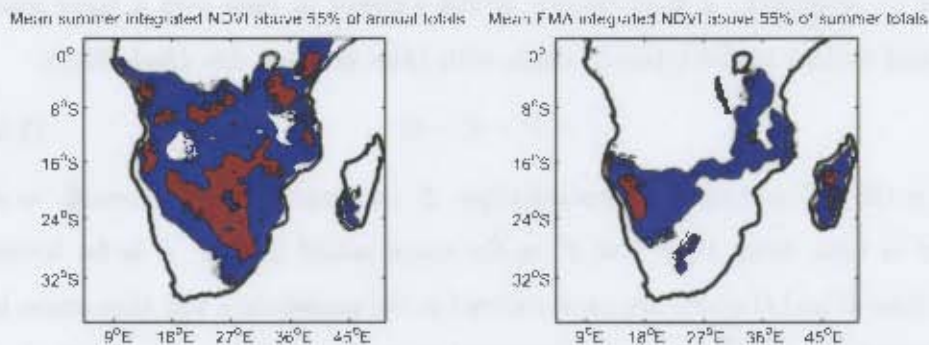


Figure 2.5: Areas of mean summer (left) and February-April (right) integrated NDVI above or equal to 55% (blue) and 60% (red) of annually integrated means over the 1982-2000 period.

When considering the contribution of the summer season to the annual integrated NDVI as shown in Figure 2.5, vegetation growth appears to depend greatly on the early summer period over a large zone extending from the Equator to 24°S/32°S on the western/eastern side of the subcontinent respectively.

On the other hand, late summer rainy season seems important for the vegetation over a region restricted between 16°S and 30°S along the west coast and between 5°S and 16°S along the east coast. Such findings agree reasonably well with the description discussed previously for summer rainfall totals. Furthermore, the contribution of early/late summer rains (November-January versus February-April) appears to affect different areas, thus presumably reflecting different mechanisms characteristic of both periods as reported in earlier studies (*D'Abreton and Tyson, 1995; Nicholson, 2000*).

2.1.2.2 Soil moisture

Soil moisture is also an important parameter used for a number of primary applications such as climatology studies, real-time drought/flood monitoring, and land-

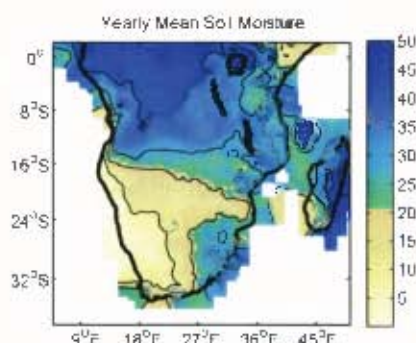


Figure 2.6: Annual monthly mean soil moisture (in mm/mth) from 1982 to 2000.

model over the Oklahoma river basin, which is drier than tropical Africa, could also lower the estimates.

2.1.3 Dynamic state of the atmosphere

As for rainfall, limited and incomplete records of atmospheric parameters over some parts of the continent, do not permit a refined description of the local and regional patterns of atmospheric variability. One alternative is to use re-analyses, providing continuous atmospheric data in time and space. The principle is based on the use of weather forecast model for assimilation of historical data from the World Meteorological Organisation (WMO).

In this study, the newly released 6 hourly NCEP-DOE AMIP Re-analysis dataset (*Kanamitsu et al., 2002*), also called NCEP R2 is used. This dataset provides a various range of atmospheric parameters at $2.5^\circ \times 2.5^\circ$ spatial resolution and vertically interpolated on 28 tropospheric levels (from 1000 mb to 10 mb). A first set of re-analyses, NCEP I, was produced globally, but surface fluxes and hydrologic balance suffer from known biases. Corrections and improvements within both data assimilation and physics of the model were initiated in 1998 through the Atmospheric Model Intercomparison Project (AMIP II) in order to produce updated corrected re-analyses, the NCEP R2 dataset.

ECMWF ERA-40 data were also used. This data consists of 6-hourly fields for a various range of atmospheric parameters at a $2.5^{\circ}\times 2.5^{\circ}$ spatial resolution, but from 1958 to present.

However, the confidence in the different re-analysis data still depends on the quality of assimilated data and the ability of the model to resolve each given variable. As a result, re-analysis variables are divided into three classes: "A" variables are influenced only by observation, "B" variables are influenced by observations and process parametrisations within the model and "C" variables are not assimilated and thus depend entirely on the model parametrisation. An overview of different classification for NCEP-R2 variables is given in Table 2.3.

Variable	Unit	Class	Grid	Level
Zonal and Meridional wind	$m.s^{-1}$	A	Regular ($2.5^{\circ}\times 2.5^{\circ}$)	Pressure levels
Air Temperature	K	A	Regular	Pressure levels
Relative Humidity	%	B	Regular	Pressure levels
Latent Heat Flux	$W.m^2$	C	Gaussian	Surface

Table 2.3: Typical characteristics for 6 hourly atmospheric fields from NCEP R2 re-analyses (after *Fauchereau*, 2004).

2.1.4 Ocean data

To characterize the surface state of the oceans surrounding southern Africa, sea surface temperature (SST) data from the UK Met Office Hadley Centre's sea ice and sea surface temperature data set (HadISST1) have been used. This data supercedes the Global sea Ice and Sea Surface Temperature (GISST) data sets, and is a unique combination of monthly globally-complete fields of SST and sea ice concentration on a 1 degree latitude-longitude grid from 1870 to date.

The SST data are taken from the Met Office Marine Data Bank (MDB), which also includes data received through the Global Telecommunications System (GTS)

from 1982 onwards. In order to enhance data coverage, monthly median SSTs for 1871-1995 from the Comprehensive Ocean-Atmosphere Data Set (COADS) were also used where there were no MDB data. The sea ice data are taken from a variety of sources including digitized sea ice charts and passive microwave retrievals.

HadISST1 temperatures are reconstructed using a two stage reduced-space optimal interpolation procedure, followed by superposition of quality-improved gridded observations onto the reconstructions to restore local detail. The sea ice fields are made more homogeneous by compensating satellite microwave-based sea ice concentrations for the impact of surface melt effects on retrievals in the Arctic and for algorithm deficiencies in the Antarctic, and by making the historical in situ concentrations consistent with the satellite data. SSTs near regions of sea ice are estimated using statistical relationships between SST and sea ice concentration. A more detailed description can be found in *Rayner et al. (2002)*. Prior to multivariate analyses, monthly normalized anomalies were derived from SST data (as well as SLP data for the SVD analyses) by subtracting the monthly mean climatology at each grid point and dividing by the corresponding standard deviation.

2.2 Methodology

As mentioned earlier, the sparse and poor observations network in some regions of southern Africa do not provide consistent long-term datasets with sufficient time resolution to investigate intra and interannual variability of variables related to energy and moisture budgets. We choose as an alternative to compute moisture fluxes and convergence from NCEP R2 re-analyses, which are presented in this section. In the following, the statistical techniques used in this study are also described.

2.2.1 Thermodynamics

Water vapour is of fundamental importance for atmospheric processes. It plays an important role in the formation of rainfall, but also in heat exchange linked to radiative transmission in the atmosphere, and heat release related to phase

changes (*Preston-Whyte and Tyson, 1988*). Evaporation supplies moisture to the atmosphere, which returns to the surface as precipitation.

Unlike rainfall, there are multiple measures of air moisture content: these which have been used for the following calculations are presented briefly in Appendix E.

At a given level, the horizontal moisture flux can be defined as follow,

$$Q_{lvl} = q_{lvl} \cdot v_{lvl} \quad (2.4)$$

where, q_{lvl} , v_{lvl} are the specific humidity and horizontal velocities at the given tropospheric level. Consequently, for each level, moisture convergence was calculated from the divergence of zonal and meridional moisture fluxes.

At monthly or seasonally time scales, the variations of precipitable water in a given volume can be approximated as (*Peixoto and Oort, 1992*),

$$-divQ = P - E \quad (2.5)$$

When integrated over the whole air column, water vapour convergence finally appears as a direct estimate of the (P,E) budget. An excess in precipitation over evaporation locally corresponds to a convergence of humidity fluxes while the reverse leads to divergence.

Furthermore, it is possible to decompose the humidity flux into its stationary and transient component as follows,

$$Q^{tot} = Q^{stat} + Q^{trans} \quad (2.6)$$

When averaging over a long enough time scale,

$$Q^{tot} = \bar{q} \cdot \bar{v} + \overline{q' \cdot v'} \quad (2.7)$$

where, $(\bar{q}, \bar{v})$ and (q', v') represent stationary/transient specific humidity and wind respectively. The first term corresponds to the mean moisture flux through the mean circulation while the second is the contribution of fluctuations from the mean.

January vector moisture flux maps with contours of mean humidity convergence at 850 mb, 700 mb and 500 mb are shown in Figure 2.7. Water vapour from the

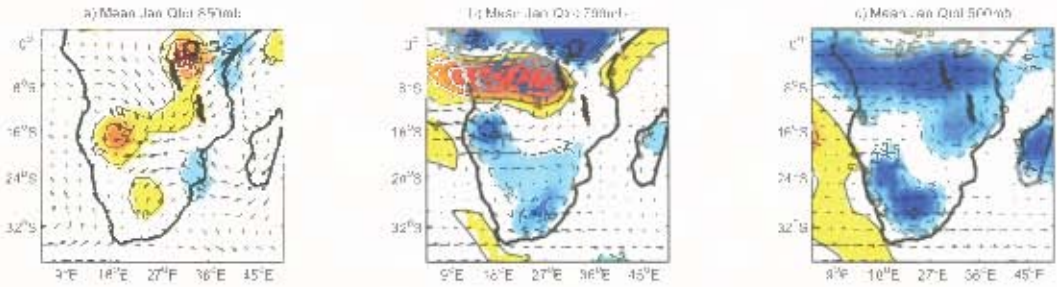


Figure 2.7: Mean January surface moisture fluxes (streamlines in $g.kg^{-1}.m.s^{-1}$ with arrows scaled at 1 unit/degree of latitude) together with contours of moisture convergence (in $g.kg^{-1}.s^{-1}$) at 850 mb (a), 700 mb (b) and 500 mb (c). Positive values contour areas of moisture convergence while negative values refer to moisture divergence at given levels.

neighbouring oceanic basins appears to converge at tropical latitudes within the summer position of the ITCZ, particularly to the east of the Congo basin, around $4^{\circ}S$ and $30^{\circ}E$ where an area of pronounced convergence can be identified at 850 mb. Further south, another marked convergence area to the southeast of Angola, centred at about $17^{\circ}S$ and $18^{\circ}E$ over the Bie plateau, is a local feature known as the Angola low (Mulenga, 1998; Cook et al., 2004; Reason and Jagadheesha, 2005b). Moisture from the tropical Atlantic feeds into the Angola low in austral summer at low-levels. Finally a third low-level convergence zone is found to the south of Botswana at about $27^{\circ}S$ and $25^{\circ}E$ corresponding with the subtropical heat-low location in summer (Preston-Whyte and Tyson, 1988), which receives moisture from the east.

At 700 mb, Figure 2.7b also shows an easterly flux crossing the continent between $10^{\circ}S$ to $25^{\circ}S$ and identified as the southern dependence of the African Easterly Jet (AEJ). It brings moisture to a convergence area lying from central Congo south-westwards from the Equator to $10^{\circ}S$.

Zones of strong humidity divergence at 500 mb (Figure 2.7c) match areas of well defined surface convergence found to the east Congo basin and Botswana: these can

be interpreted as locations of deep convection. Stronger divergence at 700 mb over south Angola also suggests substantial convection mechanisms there, but less deep within the air column.

Using a vertical domain along the west and east coasts of southern Africa for zonal moisture fluxes is a way to quantify, at least zonally, the exchange in moisture at the land-ocean interface and thus the potential role of the Atlantic and Indian Oceans in modulating southern African climate. Figure 2.8 presents the mean January climatological structure for zonal moisture fluxes along the west and east coasts of southern Africa.

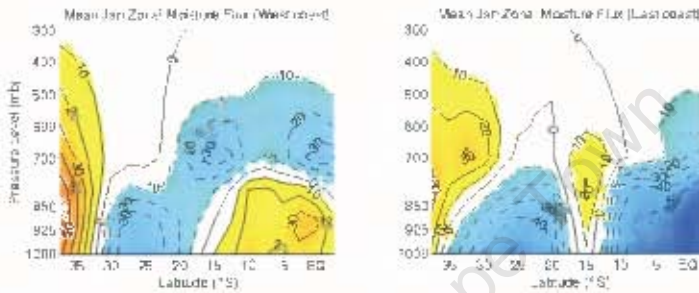


Figure 2.8: Mean vertical structure of zonal moisture fluxes along the west (left) and east (right) Southern African coast (in $g.kg^{-1}.m.s^{-1}$) for January. Positive values correspond to westerly fluxes while negative values refer to easterly fluxes.

Along the west coast of southern Africa, four key features are highlighted in summer: (a) a westerly monsoon-like flux is present from the Equator to $15^{\circ}S$ at surface levels. When this feature is particularly pronounced, it feeds in deep convection east of the Congo basin and within the Angola low,

(b) overlying this westerly flux, the southern extension of the African Easterly Jet (AEJ) is found at midtropospheric levels (just above 700 mb) as described in *Hastenrath (1985)*,

(c) to the south (between $17^{\circ}S$ and $32^{\circ}S$) an easterly flux is driven by the South Atlantic anticyclone. It connects to the southern AEJ in particular during summer, (d) south of $32^{\circ}S$, a westerly moisture flux occupies most of the air column and migrates during the year.

Similarly, along the east coast in summer, there are the following features:

- (a) an easterly flux prevails from the Equator to about 10°S at surface levels corresponding to the zonal component of the northeast monsoon, feeding the deep convection east of the Congo basin,
- (b) between 10°S and 20°S , a westerly flux maximum at 850 mb is characteristic of the cyclonic circulation prevailing over the Mozambique channel,
- (c) to the south (between 20°S and 35°S) an easterly flux linked with the South Indian Ocean anticyclone is found at surface levels,
- (d) further south (south of 35°S), a westerly moisture flux occupies most of the air column but is more marked at midtropospheric levels (700 mb to 500 mb) and corresponds to the westerly midlatitude circulation.

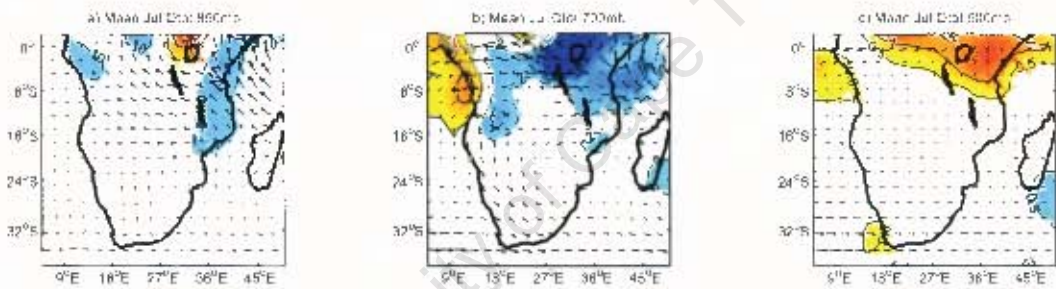


Figure 2.9: Mean July surface moisture fluxes (streamlines in $g.kg^{-1}.m.s^{-1}$ with arrows scaled at 1 unit/degree of latitude) together with contours of moisture convergence (in $g.kg^{-1}.s^{-1}$) at 850 mb (a), 700 mb (b) and 500 mb (c). Positive values contour areas of moisture convergence while negative values refer to moisture divergence at given levels.

Regarding winter, July vector moisture fluxes and convergence are presented in Figure 2.9. Due to the latitudinal migration of the ITCZ towards its northernmost position in winter, deep convection is almost suppressed over southern Africa during this period. The reversal in the monsoon regimes is noticeable along the east coast.

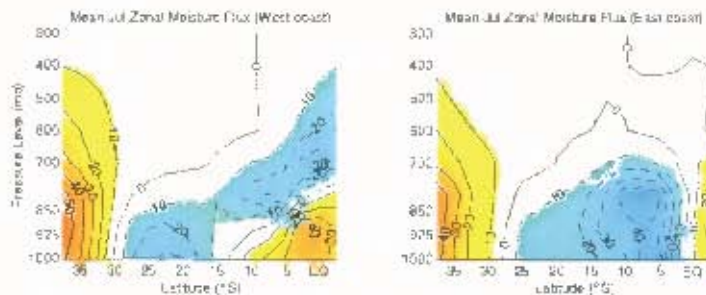


Figure 2.10: Mean vertical structure of zonal moisture fluxes along the west (left) and east (right) Southern African coast (in $g.kg^{-1}.m.s^{-1}$) for July. Positive values correspond to westerly fluxes while negative values refer to easterly fluxes.

The corresponding changes in zonal moisture fluxes (Figure 2.10) can be summarized as follows,

- (a) along the west coast in winter, the westerly monsoon-like flux is reduced in intensity, extending from the Equator to $13^{\circ}S$ at surface levels. Overlying it and lower in height compared to summer, the southern extension of the African Easterly Jet (AEJ) now consists of a low level easterly flux (at 700 mb) rather than a real midtropospheric jet. This overall seasonality agrees with the findings of *Nicholson (2003)* who considers this southern AEJ as clearly defined only from August through to December. To the south, the easterly flux is shifted northwards (between $15^{\circ}S$ and $30^{\circ}S$) and slightly weakened compared to summer. Finally, south of $30^{\circ}S$, the westerly circulation occupying most of the air column is similarly shifted northwards.
- (b) along the east coast, linked with the monsoon reversal in the Indian Ocean, the vertical structure of zonal moisture fluxes in the tropics is completely modified. As a result, north of $20^{\circ}S$, easterly/westerly components linked with the recurvature of the low-level East African Easterly jet (or Somali Jet) are now observed to the north/south respectively. This agrees with the description given in the literature (*Hastenrath, 1985*). South of $30^{\circ}S$, the westerly midlatitude circulation is found to be shifted to the north at surface levels compared to summer.

As reported in earlier studies (*Alexander and Schubert*, 1990; *Trenberth*, 1991, 1997; *Kanamaru and Salvucci*, 2003), large gaps in the data network and systematic errors in post-processing procedures result in mass imbalance within re-analysis data. This constitutes a major problem with column budget considerations and a common approach consists in adjusting winds which are not mass conservative. However, in the context of this study, the archived data are still the most accurate available data at each level. Weak wind regimes in the tropics at least at the surface, should also lower the systematic errors introduced by data assimilation methods on atmospheric energy divergence. Moreover, the 6 hourly sampling within NCEP-R2 allows the effects of surface pressure approximation to be minimized (in particular those linked with semi-diurnal tide variation in mass), while the low elevation in the tropics (except for the Great Rift valley regions) should not lead to anomalous mass convergence as noticed over high mountainous areas.

2.2.2 Statistical methods

As a result of interactions among physical processes within the ocean-atmosphere coupled system, climatic signals show variability on a wide range of temporal and spatial scales. Given the complexity and number of variables characteristic of climatic processes, there is a need to separate noise (physical and instrumental) and signal (determined by system dynamics). Elaborate statistical techniques such as multivariate analyses can be used to identify a climatic signal, which can then be described with more simple methods such as correlations and composites. For more information the reader can refer to the works of *von Storch and Navarra* (1995), *Bjornsson and Venegas* (1997) and *Venegas* (2001). Only the filtering method used in Chapter 4 is not described in the present section.

2.2.2.1 Multivariate analyses

Analyses of large gridded data sets consisting of spatially distributed time series is known as multivariate analysis. Such exploratory methods used for signal detection help to extract dominant spatial/temporal patterns within a particular field, and

involve eigenvalue decomposition. The most popular method is the Empirical Orthogonal Function (EOF) analysis. Two different approaches can be chosen for EOF decomposition: the covariance matrix approach and the singular value decomposition approach. The only difference between the two approaches is the greater degree of sophistication, computational speed and stability of the singular value decomposition approach (Venegas, 2001). EOF decomposition, also known as Principal Component Analysis (PCA), is a useful technique for compressing the information embedded in gridded datasets: it gives a compact description of spatial/temporal variability in terms of orthogonal functions or statistical modes.

The covariance matrix approach For a standadized data matrix F , the covariance matrix R_{FF} given as,

$$R_{FF} = F.F' \quad (2.8)$$

Then the eigen problem needs to be solve as,

$$R_{FF}.E = E.\Lambda \quad (2.9)$$

where Λ is the matrix containing the eigenvalues λ_k of R_{FF} .

The eigenvectors E_k corresponding to the λ_k eigenvalues are given by column in the eigenvector matrix E , and are orthogonal to one another and thus uncorrelated over space. By projecting the original data series in F onto the eigenvector E^k and summing over all locations gives the correponding time evolution $A^k(t)$ of the k^{th} EOF. As a result, the matrix A is obtained as,

$$A = E'.F \quad (2.10)$$

with rows in A representing the time series to which we will refer as Principal Components (PCs). Just as the eigenvectors E_k are uncorrelated in space so the PCs A^k are orthogonal in time.

Each eigenvalue λ_k is proportional to the percentage of variance explained in F by the corresponding k^{th} mode,

$$\% \text{ of variance explained by the } k^{th} \text{ mode} = \frac{\lambda_k}{\sum_{i=1}^K \lambda_i} \cdot 100 \quad (2.11)$$

The relative importance of each EOF mode obtained can be measure by the variance explained: it is then interesting to retained only the firsts principal modes accounting for the largest fraction of the field variance.

The Singular Value Decomposition approach The other EOF method commonly used is the singular value decomposition approach which is generally more stable and robust as it solves the eigenvalue problem through a one step method (no storage of covariance matrices). The singular value decomposition is directly performed on the standardised matrix F using the following concept,

$$F = U \Gamma V' \quad (2.12)$$

where Γ consists of the singular values γ_k on the diagonal and positive or zero elements outside the diagonal. The singular values γ_k are proportional to the eigenvalues λ_k calculated through the covariance approach such as,

$$\lambda_k = \gamma_k^2 \quad (2.13)$$

The columns of U are orthogonal and called left singular vectors of F : they are the EOF patterns associated with each singular value and are equivalent to the eigenvectors E^k calculated previously. The rows of V are also orthogonal and called right singular vectors of F . They are equivalent to the Principal Components (PCs) A^k . In matrix notation,

$$A = \Gamma V' \quad (2.14)$$

This last SVD method can be applied to identify the variability embedded within a single dataset but can also be used to study the variability between two fields. In such a case, it will only identify those modes of behavior in which the variations of

the two fields are strongly coupled. The coupling coefficient characteristic of a given mode is computed as the correlation coefficient between the expansion coefficients of each field for this mode. The relative importance of the SVD mode obtained can be measured by the covariance explained (squared covariance factor or scf), and commonly, such as in this study, a North test is used to determine whether eigenvectors are independent of each other.

EOFs interpretation and representation When interpreting EOFs, one must be careful and consider that even if the EOF techniques offer the most efficient statistical compression of a given field, they do not necessarily correspond to real dynamics or physical behaviour. The physical interpretation of EOFs is also limited by a mathematical constraint, mainly linked with the orthogonality of the geographical patterns and the temporal dependance of their associated time coefficients. Moreover, traditional EOF analyses can only detect standing patterns and EOFs are mostly dependant on the geographical domain of study (Venegas, 2001).

Regarding the presentation of the result, it is common to re-normalize the EOFs and corresponding PCS in order for them to have unit variance,

$$\begin{aligned} A^{k*}(t) &= \frac{A^k(t)}{\sqrt{\lambda_k}} \\ E_m^{k*} &= E_m^k \cdot \sqrt{\lambda_k} \end{aligned} \quad (2.15)$$

where λ_k is the k^{th} eigenvalue, E_m^{k*} and A^{k*} are respectively the re-normalized EOFs and PCs. An alternative chosen in this study is to represent EOF patterns as correlation maps: for the k^{th} mode, a correlation map will be the map of correlation coefficients r_m^k between the PC $A^k(t)$ and the values of the field $F_m(t)$ at each location $m = 1..M$. The distribution of the centers of action in the correlation map are coinciding with the one of the EOF pattern.

2.2.2.2 Statistical teleconnections

To evaluate the degree of linear relation between two variables x and y , the most commonly used method is the linear correlation. Complementary to the linear

correlation between two parameters x and y , the composite analysis gives further elements of description regarding their statistical relationship. It consists in choosing extreme events for y that will be used to sample the variable x . The mean of x for these chosen events is then compared to the mean of x for the whole population. It allows to identify which anomalies in x are associated with extreme events of y . Typically such a method can be used to validate the stability of an existing correlation through symmetry between positive and negative events. The statistical significance can be assessed using for instance a Student t-test, as described below.

A Student t-test is a statistic for measuring the significance of a difference of means between two distributions. The t-tests assume normally distributed data, that each data point is independent of the other measurements (i.e. that there is no serial correlation, or autocorrelation), and that the two data sets are independent of each others. It is thus referred to as a parametric test.

Non-parametric tests (i.e. distribution free) can also be used for statistical testing, such as Monte Carlo methods which are a widely used class of computational algorithms for simulating the behavior of various physical and mathematical systems. Monte Carlo methods are stochastic techniques, meaning they are based on the use of random numbers and probability statistics to investigate problems. The basic idea of Monte Carlo is very simple. Since mean values of stochastic variable can be expressed as the integral of a variable times the probability density function (pdf), it is possible to reverse the process by generating stochastic numbers and computing their mean by simple averaging. In theory, this should then give an approximation to the integral.

Furthermore, a Mann-Withney-Pettitt test can be used to detect changes within the characteristic of a variable before and after some time instance t . The U -test proposed by *Mann and Whitney* (1947) is based on the rank values of a sequence. The inflection point is defined as that point for which the absolute value of a sequence has reached a maximum. Derived from the U -test, the Pettitt-test finally aims at identifying that position within a series by dividing this series into one part with

significant changes of values and the other without changes. This technique is well suited to investigate the stability within climatic time-series (*Vandiepenbeeck*, 1996), thus we used such an approach to prepare data to multivariate analysis and ensure that no breaks in stationarity could introduce statistical discrepancies.

2.2.2.3 Spectral analysis

Most commonly, EOF analyses are followed by a spectral analysis of the principal components for the modes of interest. Out of this context, many methods exist to provide information on the dominant time-scale of variability embedded in a given dataset. The Continuous Wavelet transform (CWT) has been chosen in this study. Compared to traditional spectral techniques, the CWT has the advantage to identify localized variations of power within a time serie, thus allowing to decompose a signal into elementary contributions of both space and time.

Wavelets are an extension of the Fourier analysis, but instead of comparing the whole signal to sinusoides, a wavelet decomposition uses a time-localized oscillatory function referred to as mother wavelet which is continuous in both time and frequency. Consecutive segments of the signal are then compared to parts of the mother wavelet oscillating at different frequencies. Because the size of the window chosen is a critical parameter, Morlet proposed another approach: the principle is to keep the number of oscillations constant within the window but to stretch/compress the window size to capture respectively low and higher frequencies within a given signal. This explains the terminology used for wavelet transformation described as "mathematical microscope": large wavelets give approximate information about the signal, while short wavelets allows to "zoom" on the details. In the CWT, for each value of the scale used, the correlation between the scaled wavelet $g(x)$ and successive segments of the data stream is computed as follow,

$$C(b, a) = \frac{1}{\sqrt{a}} \int \bar{g}\left(\frac{x-b}{a}\right) \cdot f(x) \cdot dx \quad (2.16)$$

where a is the dilatation parameter determining the wavelet size, and b the translating parameter corresponding to the wavelet position along the signal to be analysed.

Practically, the most commonly used CWT wavelet $g(x)$ is the Morlet wavelet, a Gaussian-windowed complex sinusoid given as,

$$g(x) = \pi^{-1/4} \cdot e^{i\omega_0 x} \cdot e^{-x^2/2} \quad (2.17)$$

where ω_0 is the frequency.

It is then possible to show the amplitude versus frequency scale and how it varies along the signal. Thus the local spectrum is of particular interest and that is this component of the CWT analysis that has been used in this study: generally, the equivalent of the Fourier spectrum is the global wavelet spectrum E defined as,

$$E(1/a) = \int |C(b, a)|^2 \cdot db \quad (2.18)$$

where $1/a$ corresponds to the wave number. More details on this method can be found in *Torrence and Compo* (1998) which has been used as reference work for this section.

2.3 Summary

Radiosondes and precipitation gauges networks are sparse in Africa, thus rainfall estimates over some parts of the continent are still suffering from biases. In this respect, several precipitation datasets were compared together with land hydrology data (mainly NDVI and soil moisture) to examine how key features of southern African climate are captured between them. As a consequence, CRU rainfall estimates together with NDVI and soil moisture will be used in this thesis to analyse variability in the water cycle over the subcontinent.

Although less problematic than rainfall estimates, the limited and incomplete range of atmospheric parameters issued from observations do not permit a detailed examination of the regional patterns of atmospheric variability. The use of re-analyses, providing a panel of continuous atmospheric data in time and space, from both ECMWF ERA-40 (from 1962 to 2002) and NCEP R2 (over the 1979-2000 period) datasets, is considered as an alternative for the purposes of this study. Despite the marked influence of energetic features such as the ITCZ, tropical Africa

is still one of the least studied areas in terms of energy of moisture exchange at both continental and global scales. Thus, moisture fluxes and divergence were computed from NCEP R2 in order to identify the key pathways of water vapour transport over the subcontinent, as represented in the re-analyses over the last 20 years.

In the following chapter, multivariate techniques will be used to describe the variability in both the ocean-atmosphere systems prevailing over the South Atlantic and Indian Ocean basins, and zonal moisture fluxes along the east and west coasts of southern Africa. Then more simple methods, such as correlations and composites, will be used to assess potential relationships between the variability in zonal water vapour transport over the subcontinent, the local hydrological cycle and primary modes of ocean-atmosphere variability in the neighbouring ocean basin regions.

The data and methods presented in this chapter constitute a basis for this thesis that attempts to study how southern African rainfall variability can be modulated by water vapour input from the South Atlantic and Indian Oceans, using calculations of moisture fluxes and convergence.

Chapter 3

Variability in the oceanic basins regions

It is known that rainfall variability in southern Africa is influenced by sea-surface temperatures (SSTs) of the Pacific but also of the neighbouring oceanic basins (*Mason, 1990, 1995; Reason and Godfred-Spenning, 1998; Reason and Mulenga, 1999; Saji et al., 1999; Richard et al., 2001; Reason, 2001a,b; Camberlin and Poccard, 2001; Behera and Yamagata, 2001; Rouault et al., 2002; Reason and Jagadheesha, 2005a; Reason et al., 2006; Washington and Preston, 2006*). As such, it is essential to identify the large scale interactions between the ocean and the overlying atmosphere. Empirical Orthogonal Functions (EOF) and Singular Decomposition Values (SVD) analyses are used simultaneously on 40-years (1962-2002) monthly dataset that includes SSTs (from Hadley Centre) and atmospheric parameters (sea level pressures, 850 mb winds) as well as latent heat flux from ECMWF ERA-40 dataset to help determine patterns of variability over southern Africa and the neighbouring oceans. The results presented in this chapter concern first the South Atlantic and then the Indian Ocean basin. Regarding links with the Pacific, the reader is referred to Chapter 1. In the following, ENSO relationships between the prevailing modes of variability identified in both the South Atlantic and South Indian Ocean basin regions are briefly examined.

3.1 The South Atlantic Ocean

The South Atlantic has been regarded for long as a secondary source of moisture for the subcontinent but recent studies have shown its substantial importance to southern African climate variability (*Cook et al.*, 2004; *Reason and Jagadheesha*, 2005a). For instance, anomalous intrusion of warm tropical waters down to the northern coast of Namibia that occurs roughly once or twice per decade, so-called Benguela Niños (*Shannon et al.*, 1986; *Florenchie et al.*, 2003, 2004), can be associated with significant anomalies in the local atmospheric stability and evaporation. A positive correlation has been established between SST anomalies off Angola, latent heat flux and rainfall at the coast (*Hirst and Hastenrath*, 1983; *Rouault et al.*, 2003).

The domain chosen for the following multivariate approach ranges from the Equator to 50°S and from 50°W to 20°E. Because of the known existence of tropical modes within the Atlantic basin (*Zebiak*, 1993; *Hastenrath and Greischar*, 1993; *Carton and Huang*, 1994; *Chang et al.*, 1997; *Ruiz-Barradas et al.*, 2000) this investigation for variability has been extended to a region from 20°N to 20°S within the same longitudinal band. In the following, the chosen dataset have been smoothed using a 3 months-running mean prior to multivariate analysis.

3.1.1 Variability within the South Atlantic Ocean regions

In the South Atlantic Ocean, modes are commonly identified in SSTs as representative of distinct variability in the tropics and in the midlatitudes (*Venegas et al.*, 1997; *Sterl and Hazeleger*, 2003; *Colberg*, 2006; *Colberg and Reason*, 2006). The main patterns in SLPs are characteristic of modulations in intensity as well as the displacement of the South Atlantic anticyclone (*Venegas et al.*, 1997; *Sterl and Hazeleger*, 2003). The differences between the analyses of each fields and their relationships in the different studies reflect mainly the effects of the spatial domain and period considered as well as the statistical methods used (*Palastanga et al.*, 2002; *Colberg*, 2006). However, these investigations reveal that variability in the South Atlantic mainly ranges from interannual to interdecadal time-scales.

The first EOF leading modes of variability in the South Atlantic for SSTs and SLPs are respectively presented in Figure 3.1 and 3.2 together with their superimposed annual climatology and corresponding expansion coefficients. Higher modes being statistically degenerate, we retained only the first two modes in accordance with the scree test (not shown). These modes explain up to about half of the total variance explained within each field.

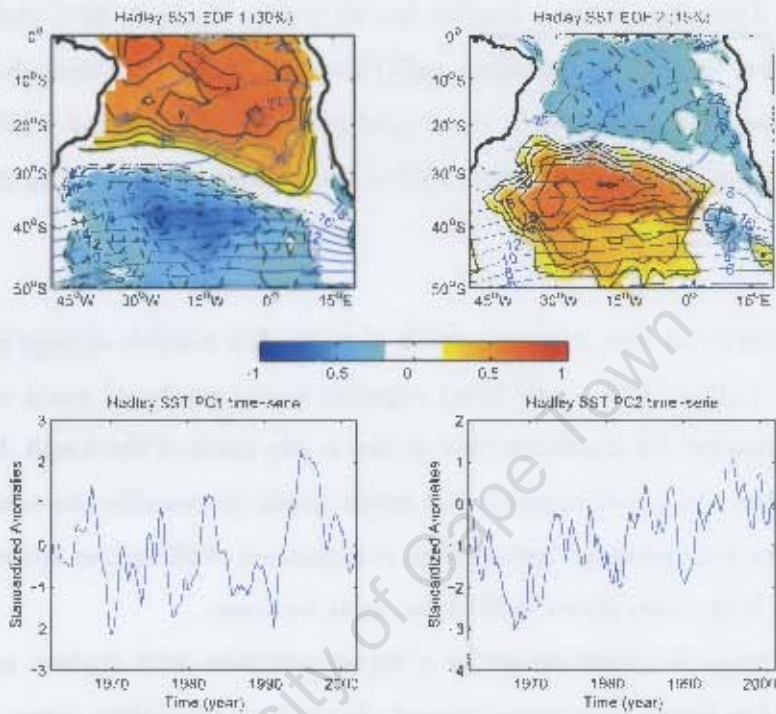


Figure 3.1: First leading EOF modes for SSTs in the South Atlantic over the period 1962-2002. In brackets is indicated the percentage of variance explained for each mode. The spatial patterns (in colors) are presented as correlation maps (top) between the corresponding expansion coefficient (bottom) and anomalies at each grid point. The corresponding annual climatology for this period is shown in blue contours.

The first mode (Figure 3.1) shows largest opposite loadings in SSTs over the northern and southern parts of the South Atlantic basin, in the area where the SST gradient is oriented northwestwards from South Africa to northeast Brazil. The highest positive loadings to the north mean that the EOF representation of warm-

ing/cooling in the tropical Atlantic implies a shift the region of maximum SST gradient. Weaker negative SST loadings are found in the midlatitude South Atlantic. This primary mode of SST variability in the South Atlantic resembles the first mode of *Sterl and Hazeleger* (2003), *Colberg* (2006) and *Colberg and Reason* (2006). It contributes about 30% of the variance explained.

The second mode displays an out-of-phase relationship in SSTs north/south of about 20-25°S. The main positive loading has its centre at about 30°S and 15°W and suggests that this EOF describes variability within the midlatitude South Atlantic and bears resemblance with the second mode of *Venegas et al.* (1997) and (*Sterl and Hazeleger*, 2003). It explains 15% of the variance within South Atlantic SSTs.

For SLP (Figure 3.2), the primary pattern of variability consists of large positive loadings north of about 25°S, while lower negative scores are found south of 35°S. Maximum loadings for this mode are located just to the north of the South Atlantic anticyclone centre. As a consequence, this mode seems to describe the variability in intensity of the high pressure system with modulations of SLPs over the tropical South Atlantic. It explains about 40% of the total variance.

The second mode is characterized by a dipole structure with highest loadings in the south of the basin. The superimposed climatology explicitly shows that it corresponds to a weakening/strengthening of the South Atlantic anticyclone and its simultaneous north-south displacement, due to reduced/enhanced SLPs in the midlatitudes (i.e. to the south of its climatological centre). Despite the absence of anti-correlation loadings to the north and south of the basin in the analyses of *Venegas et al.* (1997) and *Sterl and Hazeleger* (2003), this bears resemblance with the description given for the first mode they identified in SLPs. This mode contributes about 17% of the total variance explained.

Table 3.1 lists the correlation scores between SST and SLP expansion coefficients of each leading EOF modes described above, together with their correlation with ENSO using the SOI. It is worth noting that the low significant thresholds, in Table

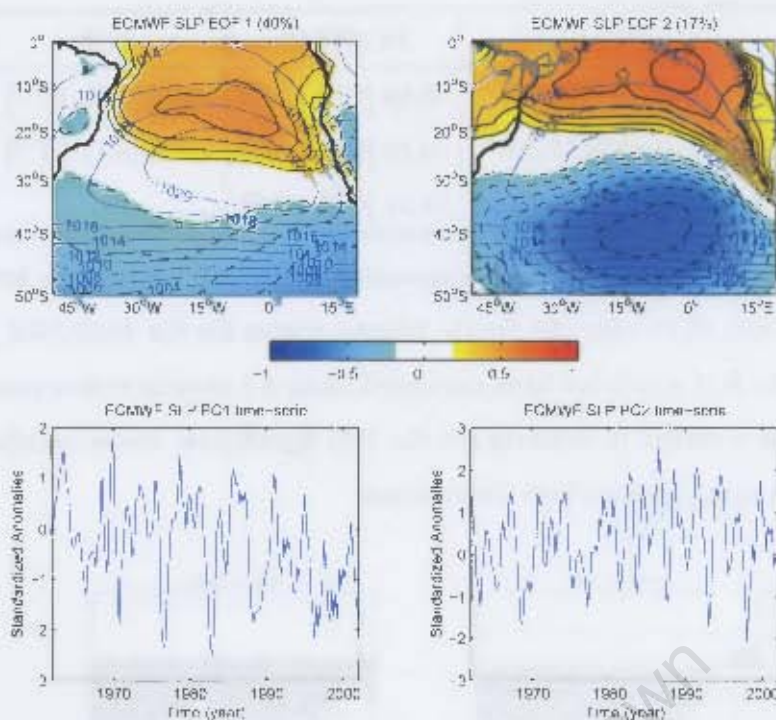


Figure 3.2: First leading EOF modes for SLPs in the South Atlantic over the period 1962-2002. In brackets is indicated the percentage of variance explained for each mode. The spatial patterns (in colors) are presented as correlation maps (top) between the corresponding expansion coefficient (bottom) and anomalies at each grid point. The corresponding annual climatology for this period is shown in blue contours.

3.1 and in the remaining of this chapter, might result from the auto-correlation in the time-series, partly due to the 3-months filtering.

The SLP PC1 and SST PC1 time-series are significantly (-0.50) and best correlated at zero lag (Figure 3.3). SST PC2 and SLP PC2 time-series are negatively correlated without lag (-0.16) and Figure 3.3 shows that the correlation is best when the atmosphere leads the ocean by 2 months: this could suggest that atmosphere-to-ocean forcings link these two modes of variability but the coupling coefficient is rather weak (-0.22).

Very low correlations are found between the SOI and SST PC1 and PC2 at zero lag. Lagged correlations in Figure 3.4 exhibit strongest scores when SST PC1 leads

	PC1(SLP)	PC2(SLP)	SOI
PC1(SST)	-0.50 [8.93.10 ⁻²]	-0.06 [9.01.10 ⁻²]	-0.01 [8.92.10 ⁻²]
PC2(SST)	-0.12 [9.07.10 ⁻²]	-0.16 [9.34.10 ⁻²]	-0.04 [9.17.10 ⁻²]
SOI	-0.27 [9.23.10 ⁻²]	-0.31 [8.97.10 ⁻²]	-

Table 3.1: Correlations between the expansion coefficients of the first two EOF modes in SSTs and SLPs over the South Atlantic region for the 1962-2002 period, together with the SOI which has been smoothed using a 3 months-running-mean for consistency. The numbers in brackets are the 95% significance levels calculated for each correlation using Monte-Carlo simulations.

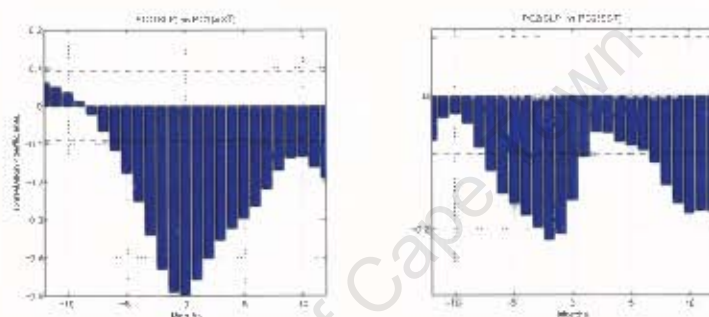


Figure 3.3: Lagged correlations between SST and SLP leading EOF modes.

SSTs in the Pacific by 9 months (0.38). Relationships between the SOI and SST PC2 reveal only weak maximums when SST PC2 leads by 5 months (-0.12) and when the SOI leads by 10 months (-0.1). Both SLP PC1 and PC2 appear to be significantly correlated with ENSO (respectively -0.27 and -0.31) and from lagged correlations (Figure 3.4) the relationship seems best when SLP PC1 leads SSTs in the Pacific by 2 months (-0.35), while stronger correlations are found with SLP PC2 when SSTs in the Pacific lead by 1 month (-0.32), which shows that ENSO has a potential relationship with these modes.

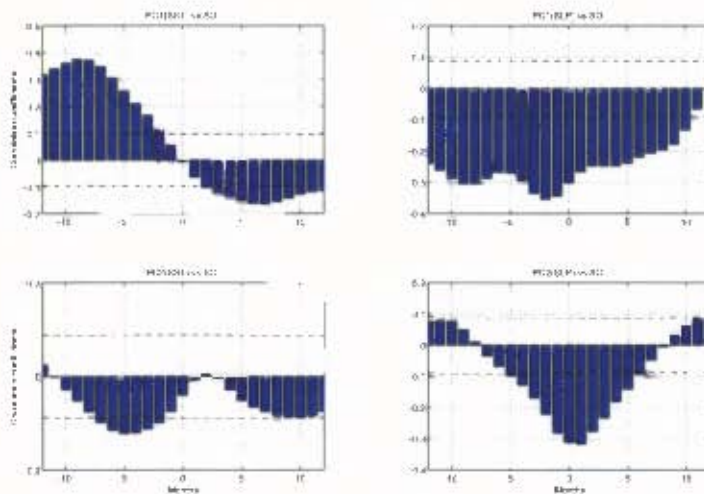


Figure 3.4: Lagged correlations between the SOI and the SST/SLP expansion coefficients (left/right) for the first and second EOF modes (top and bottom respectively).

In conclusion, the first mode in SSTs shows large scale warming/cooling in the tropical Atlantic with an associated shift of the region of maximum SST gradient. The second mode consisting of largest SST anomalies in the southern part of the basin, represents variability in the midlatitude South Atlantic. In SLPs, the first EOF leading mode can be described as the variations in intensity of the South Atlantic anticyclone due to modulations in surface pressures in the tropical South Atlantic. The pattern obtained for the second mode in SLPs suggests the simultaneous weakening/strengthening and north-south displacement of the South Atlantic anticyclone, associated with modulations of surface pressures in the midlatitudes. Given the effects of the spatial domain/time-period considered as well as the statistical methods and data used, differences are found between the primary patterns identified here and those isolated in earlier studies (Venegas *et al.*, 1997; Sterl and Hazeleger, 2003; Colberg, 2006; Colberg and Reason, 2006). Nevertheless, the results are still consistent with the overall description of variability in the South Atlantic reflecting the dual influence of the tropical/midlatitude regions as well as the importance of the circulation linked with the South Atlantic anticyclone.

3.1.2 Ocean-atmosphere variability in the Atlantic Ocean

To refine the study, Singular Value Decomposition (SVD) methods are now applied to identify the variability between SSTs and each one of the three atmospheric parameters, SLPs, 850 mb winds and surface latent heat flux (LHF) respectively. 850 mb winds have been considered as a complex field and the characteristics relative to the zonal component will constitute the main focus of the wind field analysis. To better assess and gain more insights into the relationship between SSTs and SLPs found in the previous section, SVD analyses are first performed on the same domain, covering the whole of the South Atlantic basin. Focusing on the potential mechanisms associated with the variability, lead/lag relationships are investigated as some modes are dependant on the seasonal cycle. Due to differences between tropical dynamics and mechanisms at higher latitudes, SVD analyses have been extended to a tropical sector ($[20^{\circ}\text{N}-20^{\circ}\text{S}; 50^{\circ}\text{W}-20^{\circ}\text{E}]$) in order to isolate modes of variability characteristic of the ocean-atmosphere system in the tropics (*Chang et al.*, 2006).

3.1.2.1 Ocean-atmosphere variability within the South Atlantic basin

In this part, a similar spatial domain to the one chosen for the previous EOF analyses of SST and SLP fields separately is considered for the SVD analyses. For SLPs, 850 mb winds and LHF, the first two modes of variability with SSTs account for 90%, 78% and 74% of the total covariance explained respectively.

	SST/SLP	SST/ U_{850mb}	SST/LHF
Mode 1	-0.50 [9.10^{-2}]	-0.02 [$8.96.10^{-2}$]	-0.16 [9.10^{-2}]
Mode 2	-0.01 [$9.04.10^{-2}$]	-0.14 [$8.57.10^{-2}$]	0.10 [$8.83.10^{-2}$]

Table 3.2: Coupling correlation coefficients between expansion coefficient time-series corresponding to the first two leading SVD modes within the South Atlantic over the 1962-2002 period. In brackets are the 95% significance level for the correlations using Monte-Carlo simulations.

Table 3.2 lists the respective coupling correlation coefficients between the expansion coefficients of both variables considered for each SVD analyses (i.e. between SSTs and SLPs, 850 mb winds and LHF). The first mode exhibits strong significant relationship between SSTs and SLPs (-0.50) and weaker coupling coefficients for surface winds and latent heat flux. For the second mode a particularly low coupling coefficient is found between SSTs and SLPs while the relationship between SSTs and 850 mb zonal winds is dominant but also rather weak.

First mode of ocean-atmosphere variability in the South Atlantic basin

Spatial patterns and expansion coefficients for the first mode of variability within the ocean-atmosphere system over South Atlantic basin are given in Figure 3.5. SST and SLP patterns and expansion coefficient time series obtained for this first mode closely resemble those obtained with the primary EOF mode of SST and SLP fields separately. This mode seems to be related to the significant relationship found for EOF SST PC1 and EOF SLP PC1 over the South Atlantic domain: a correlation of 0.97 between EOF SST PC1 and SVD SST PC1 was found, while it is about 0.99 between EOF SLP PC1 and SVD SLP PC1.

This first mode shows largest loadings in SSTs for all SVD analyses in the northern part of the South Atlantic basin and contributes about 74%, 54% and 55% of the total covariance explained between SSTs and SLPs, 850 mb winds and LHF respectively. As mentioned earlier, it describes a warming/cooling in the tropical South Atlantic associated with a shift the region of maximum SST gradient.

The associated SST expansion coefficients obtained for each SVD analyses are shown in Figure 3.6 and the correlation coefficients between them in Table 3.3: all three time-series appear well correlated ($r > 0.9$) between each other which confirms consistent characteristics of this first mode of variability within the South Atlantic Ocean regions were isolated. A wavelet analysis of the corresponding SST expansion coefficients reveals variability at about 14 years, agreeing with previous studies (Venegas *et al.*, 1997; Sterl and Hazeleger, 2003). Interannual variability seem also characteristic of this mode, agreeing with Colberg (2006) and Colberg and Reason

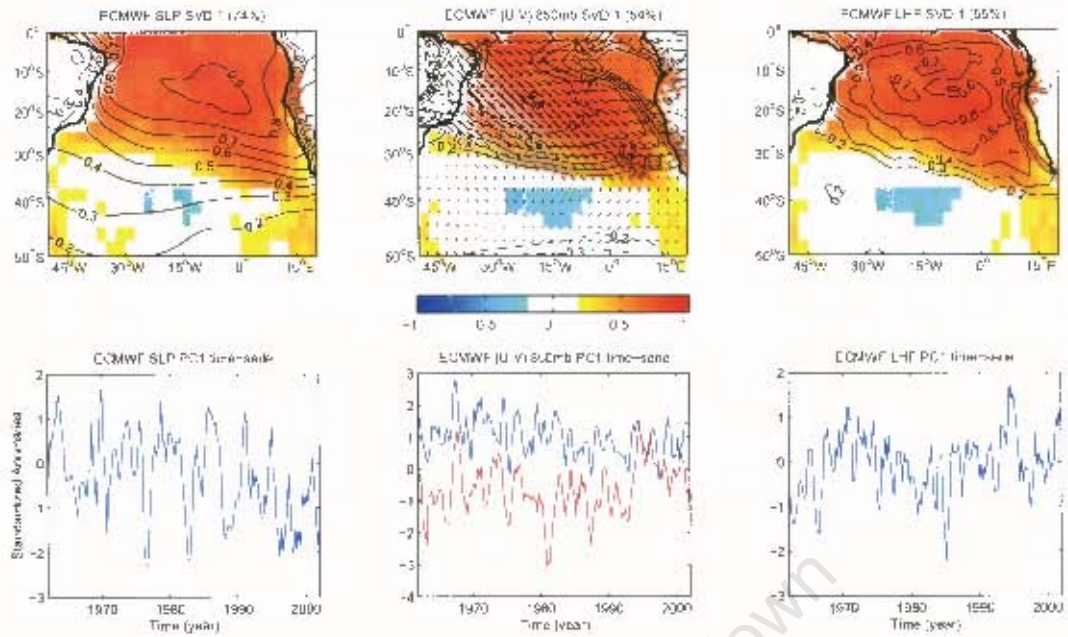


Figure 3.5: First leading modes of SVD analyses in the South Atlantic region, between SSTs (in colors) and SLPs (left), 850 mb winds (middle) and LHF (left) in contours respectively. The squared covariance factor (scf) is given for each mode in brackets. The spatial patterns (top) are presented as correlation maps between the corresponding expansion coefficient and anomalies at each grid point. The contours in 850 mb winds map represent correlation scores for the zonal component. The expansion coefficients for each variable (SLPs, U_{850mb}/V_{850mb} in blue/red respectively, and LHF) are displayed in the bottom panels.

Expansion coefficients	r
$SST_{SLP}/SST_{(U,V)_{850mb}}$	0.94
SST_{SLP}/SST_{LHF}	0.99
$SST_{LHF}/SST_{(U,V)_{850mb}}$	0.96

Table 3.3: Correlation coefficients between SST expansion coefficients as isolated in the first leading SVD modes within the South Atlantic over the 1962-2002 period (between SSTs and SLPs, 850 mb winds and LHF respectively).

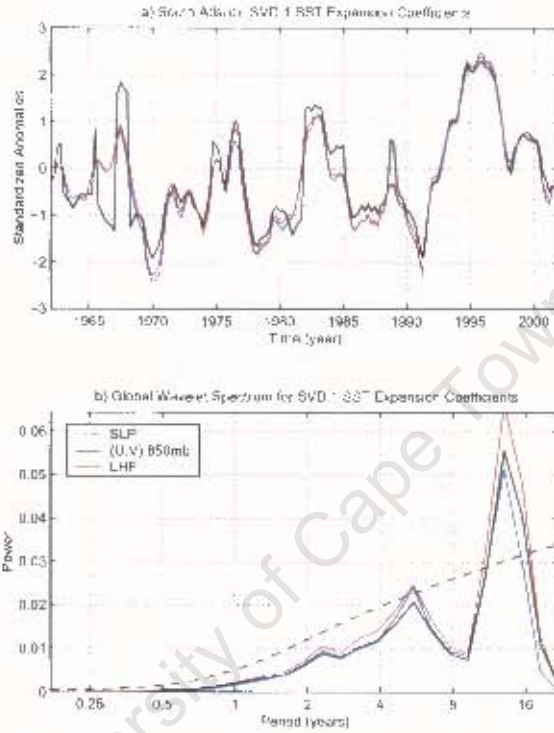


Figure 3.6: (a) SST expansion coefficients for the first leading modes of SVD analyses in the South Atlantic region, between SSTs and SLPs, 850 mb winds and LHF respectively, (b) Continuous Wavelet Spectrum (CWT) for the corresponding SVD SST PC1 expansion coefficients. The 95% confidence level is indicated by the dashed black line.

(2006). In the present analysis, a marked signal is found at about 4-5 years suggesting a potential influence of ENSO within this mode of variability as emphasized in Figure 3.8 bottom panel.

Table 3.2 indicates strong negative correlation between SSTs and SLPs (-0.50). From the patterns in Figure 3.5, the shift of the region of warm SSTs due to the warming/cooling in the tropical Atlantic seems to be associated with a weakened/strengthened subtropical anticyclone due to reduced/enhanced pressure over these regions. The weak coupling coefficient between SSTs and U_{850mb} wind as well as LHF (Table 3.2) suggests no direct relationships.

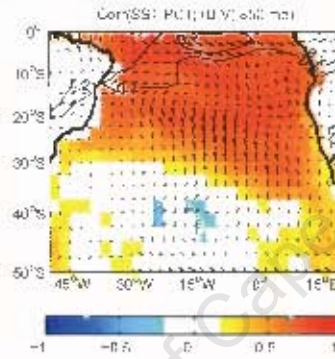


Figure 3.7: Correlations between SVD 1 SST_{SLP} expansion coefficient and the $(U, V)_{850mb}$ field. The contours represent correlations for the zonal wind component. The pattern in SSTs characteristic of this mode is plotted in colors.

The projection of the SST expansion coefficient time-series characteristic of the SST/SLP SVD analysis on the 850 mb wind field is presented in Figure 3.7. Northerly anomalies seem to prevail in the central basin and could be a response of the wind field to the pressure gradient associated with reduced SLPs in the tropical Atlantic. Moreover, reduced trades are found in the equatorial Atlantic and could act to support SST warming locally. The relationships are rather weak and the effect of winds is not so pronounced over the southeastern parts of the tropical Atlantic where anomalous SSTs also develop. This is consistent with the results of Colberg (2006) and Colberg and Reason (2006) who found for a similar mode, that despite significant correlations with surface winds in the northwestern/western

parts of the basin, warming in the southeastern South Atlantic tropics was not due to local wind stress anomalies but rather to non-local dynamics such as the propagation of equatorial Kelvin waves as proposed in previous studies (Carton and Huang, 1994; Florenchie et al., 2003, 2004).

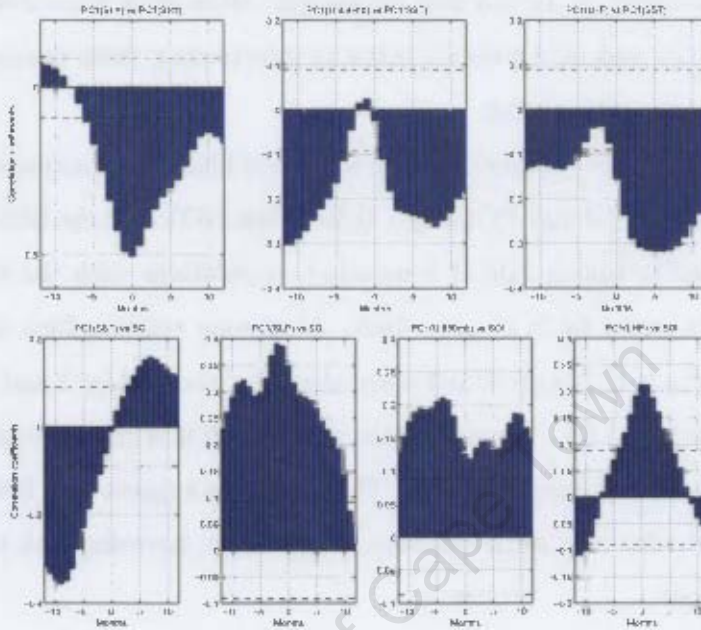


Figure 3.8: On top are presented lagged correlations between SVD 1 SST and SVD 1 SLP (left), SVD 1 U_{850mb} (middle), and SVD 1 LHF (right) expansion coefficients. In the bottom panel are shown lagged correlations between the SOI and SVD 1 SST (far left), SVD 1 SLP (left), SVD 1 U_{850mb} (right) and SVD 1 LHF (far right). Significance level at 95% using Monte-Carlo simulations are plotted in dashed lines.

Lagged correlations (Figure 3.8 top panel) exhibit a synchronous negative relationship between SLPs and SSTs, as discussed previously, while weaker maximum correlation (-0.3) when U_{850mb} winds lead SSTs by 12 months could further indicate non-local influences. Another weak maximum is found when SSTs lead U_{850mb} winds by about 6-9 months. These maximum correlation scores around 12 and 6 months could support the idea that this mode is phase locked on the seasonal cycle. In addition, a maximum relationship is found between SSTs and latent heat flux (-0.32) when SSTs lead LHF by about 4 to 8 months. The patterns in Figure 3.5 suggest a

reduction/enhancement of winds which are particularly pronounced in the western central Atlantic together with reduced/enhanced evaporation locally, which could lead to a further warming/cooling of SSTs. As noted in *Venegas et al.* (1997), SST anomalies inducing changes in the atmospheric circulation could in turn strengthen the existing SST anomaly by air-sea heat exchanges. More work is needed to investigate such hypothesis and care must be taken in interpreting these results as some of the relationships are rather weak.

Relationships with ENSO (Figure 3.8 bottom panel) illustrate maximum negative correlation between the SOI and PC1 SST (-0.35) when SSTs lead the SOI by about 9 months. At a similar timing (about 8 months), correlations with the SOI are of about 0.32 in SLPs and 0.19 in U_{850mb} winds. Maximum relationships with SLPs (0.39) and U_{850mb} winds (0.21) are found when they lead the SOI by 2 and 4 months respectively. Concerning LHF, the correlation with the SOI is strongest at zero lag and when the SOI leads by 1 month (0.22). These findings suggest that ENSO could have a potential but relatively weak influence in this mode, agreeing with the results from spectral analysis.

Second mode of ocean-atmosphere variability in the South Atlantic basin

Figure 3.9 shows spatial patterns and corresponding expansion coefficients for the second leading mode of variability within the ocean-atmosphere system over the South Atlantic basin. SST and SLP patterns and expansion coefficients obtained for this second mode closely resemble those obtained with the second EOF of SST and SLP fields separately. A correlation of -0.92 is found between EOF SST PC2 and SVD SST PC2, while between EOF SLP PC2 and SVD SLP PC2 it is about 0.96.

This second mode shows an out-of-phase relationship in SSTs north and south of 25°S with largest negative loadings in SSTs in the midlatitude South Atlantic. Thus, this mode can be considered as representing the variability in the midlatitude South Atlantic. It contributes about 16%, 23% and 19% of the total covariance explained between SSTs and SLPs, 850 mb winds and LHF respectively.

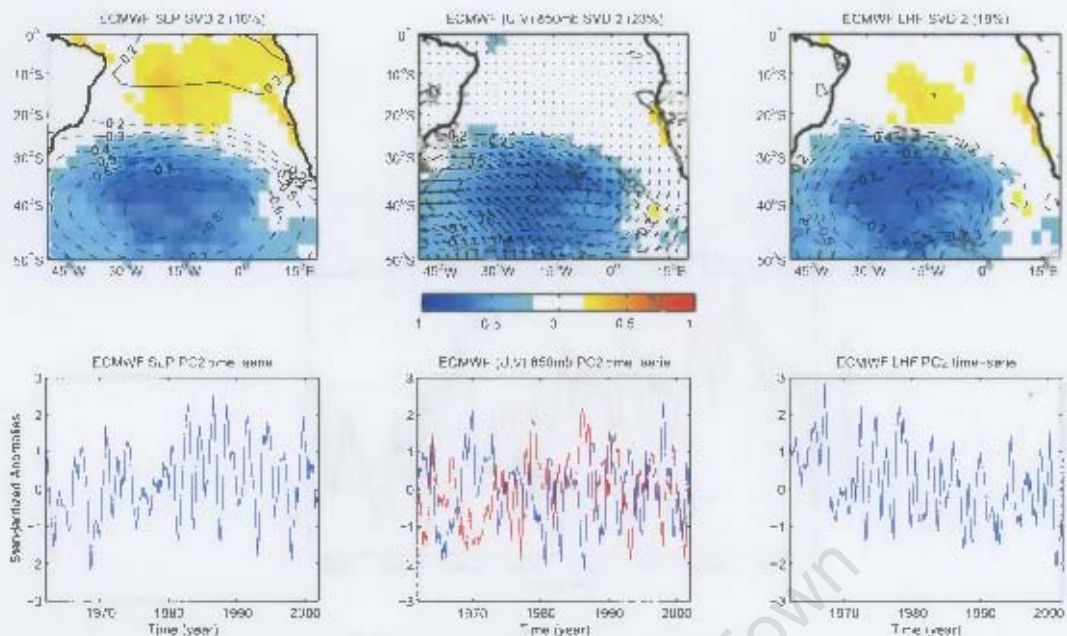


Figure 3.9: Second leading modes of SVD analyses within the South Atlantic between SSTs (in colors) and SLPs (left), 850 mb winds (middle) and LHF (left) in contours respectively. The scf is given for each mode in brackets. The spatial patterns (top) are presented as correlation maps between the corresponding expansion coefficient and anomalies at each grid point. The contours in 850 mb winds map represent correlation scores for the zonal component. The expansion coefficients for each variable (SLPs, U_{850mb}/V_{850mb} in blue/red respectively, and LHF) are displayed in the bottom panels.

Expansion coefficients	r
$SST_{SLP}/SST_{(U,V)_{850mb}}$	0.87
SST_{SLP}/SST_{LHF}	0.96
$SST_{LHF}/SST_{(U,V)_{850mb}}$	0.93

Table 3.4: Correlation coefficients between SST expansion coefficients as isolated in the second leading SVD modes within the South Atlantic over the 1962-2002 period (between SSTs and SLPs, 850 mb winds and LHF respectively).

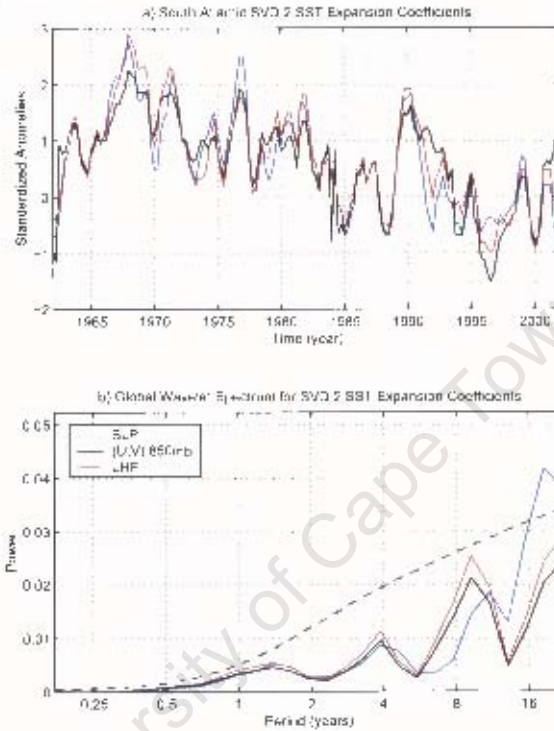


Figure 3.10: (a) SST expansion coefficients for the second leading modes of SVD analyses in the South Atlantic region, between SSTs and SLPs, 850 mb winds and LHF respectively, (b) Continuous Wavelet Spectrum (CWT) for the corresponding SVD SST PC2 expansion coefficients. The 95% confidence level is indicated by the dashed black line.

The associated expansion coefficients obtained for each SVD analysis are shown in Figure 3.10 and the correlation coefficients between them in Table 3.4: all three time-series appear well correlated ($r > 0.8$) between each other, and support the fact that consistent characteristics of this second mode of variability in the South Atlantic region have been isolated. The CWT spectra associated with the corresponding SST expansion coefficients show variability at about 4 years together with interdecadal variations, agreeing with previous studies (*Sterl and Hazeleger, 2003*). Lagged correlations (not shown) exhibit only a weak correlation between the SOI and SSTs maximum with a lag of 3 months.

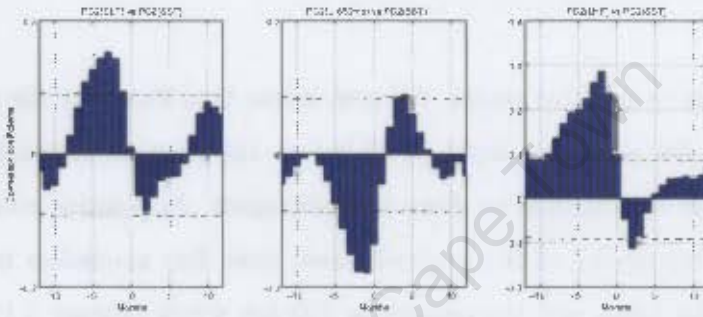


Figure 3.11: Lagged correlations between SVD 2 SST and SVD 2 SLP (left), SVD 2 U_{850mb} winds (middle), and SVD 2 LHF (left) expansion coefficients. Significance level at 95% using Monte-Carlo simulations are plotted in dashed lines.

Figure 3.11 shows a maximum positive relationship between SSTs and SLPs (0.15) when SLPs lead SSTs by 3 months. The pattern in SLPs (Figure 3.9) suggests a weakened/strengthened subtropical anticyclone due to modulations of surface pressures in the midlatitudes (i.e. to the south of its climatological position) together with its north-south displacement, in agreement with *Sterl and Hazeleger (2003)*, but the correlation is rather weak. In addition, the strongest relationships with U_{850mb} winds (-0.18) and LHF (0.28) are found when they lead SSTs by 1-2 months.

The projections of the SST expansion coefficient time-series characteristic of the SST/SLP SVD analysis on the 850 mb wind field and LHF are presented in Figure

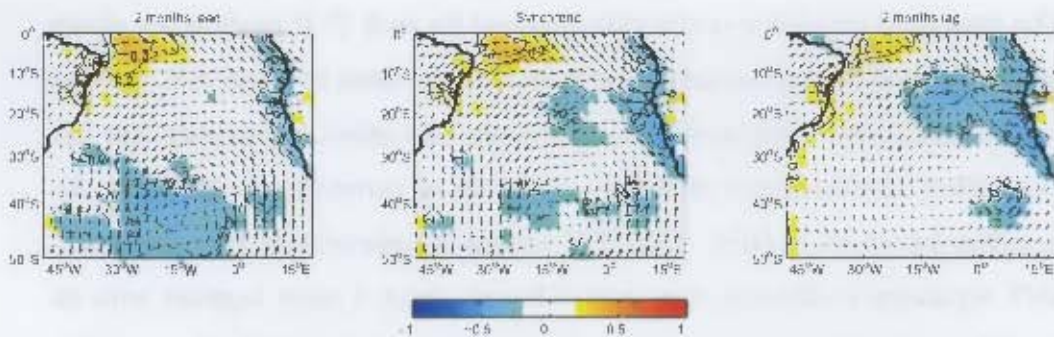


Figure 3.12: Correlations between SVD 2 SST_{SLP} expansion coefficient and 850 mb winds (streamlines) and LHF (colors and contours) at 2 months lead (left), synchro (middle) and at 2 months lag (right).

3.12. It shows prior to positive events, reduced latent heat flux over the southwest midlatitude South Atlantic that might be linked to the anomalous southwestward advection of warmer and moister air from the subtropics. As positive events of this mode develop, a dampening of the original latent heat flux anomalies is found to the southwest of the basin and the pattern in 850 mb winds (Figure 3.12) suggest that it could be due to stronger-than-normal northerlies locally, thus acting to lower midlatitude SSTs. Enhanced southeasterlies over the southeastern parts of the basin (north and south of 20°S) are accompanied by negative latent heat flux anomalies off the west coast of southern Africa that seem to persist and be most developed after the positive events peak (at 2 months lag).

3.1.2.2 Ocean-atmosphere variability within the tropical Atlantic sector

In the tropical Atlantic, two main modes are identified in the literature: a zonal mode often referred to as the "Atlantic Niño" or equatorial mode (Merle, 1980; Zebiak, 1993; Carton and Huang, 1994; Ruiz-Barradas *et al.*, 2000), and a meridional mode termed the interhemispheric or dipole mode (Hastenrath and Greischar, 1993; Chang *et al.*, 1997; Servain *et al.*, 1999). The signature of the zonal mode is confined to the eastern equatorial Atlantic region bearing resemblance to El Niño: the warm-

ing of SSTs locally is reported to be accompanied by a relaxation of the trade winds in the central basin leading to a southward shift of convection (*Zebiak, 1993; Carton and Huang, 1994; Ruiz-Barradas et al., 2000*). Nevertheless, recent studies have found a similar mode in the tropical Atlantic but with differences in the dynamics involved (*Chang et al., 2006*). Interannual variability is reported to dominate this mode (*Ruiz-Barradas et al., 2000; Chang et al., 2006*). The meridional mode on the other hand is characterized by a stronger-than-normal northward oriented SST gradient which is found to be accompanied by a cross-equatorial flow leading to a meridional shift of the maximum surface wind convergence and thus of the ITCZ (*Hastenrath and Greischar, 1993*). As this cross-equatorial flow is deflected by the Coriolis force, it results in stronger/weaker-than-normal trades to the south/north of the Equator that would act to strengthen the original warm/cold anomalies in the northern/southern tropics (*Ruiz-Barradas et al., 2000; Chang et al., 2006*). Such mechanisms are characteristic of a positive tropical feedback referred to as the Wind-Evaporation-SST (WES) feedback (*Xie and Philander, 1994*). Interdecadal variability is reported to prevail within this mode of variability (*Ruiz-Barradas et al., 2000; Chang et al., 2006*).

In order to integrate these modes, SVD analyses were extended to a tropical domain in the Atlantic ranging from 20°N to 20°S and from 50°W to 20°E. For SLP, 850 mb winds and LHF fields, the first two leading modes of variability with SSTs are retained in accordance with the scree test (not shown) and account for 98%, 93% and 90% of the total covariance explained respectively.

Table 3.5 lists the respective coupling correlation coefficients between the expansion coefficients of both variables (i.e. between SSTs and SLPs, U_{850mb} winds and LHF). The values imply that there may be significant relationships between SSTs and SLPs for both modes, and to a certain extent with 850 mb zonal winds for the second mode, while for the SST/LHF analysis the coupling coefficients are closer to their 95% significance level.

	SST/SLP	SST/ U_{850mb}	SST/LHF
Mode 1	-0.44 [8.77.10 ⁻²]	-0.07 [9.20.10 ⁻²]	-0.1 [8.97.10 ⁻²]
Mode 2	-0.68 [8.82.10 ⁻²]	0.23 [8.93.10 ⁻²]	-0.05 [8.92.10 ⁻²]

Table 3.5: Coupling correlation coefficients between expansion coefficient time-series corresponding to the first two leading SVD modes in the tropical Atlantic over 1962-2002 period. In brackets are the 95% significance level for the correlations using Monte-Carlo simulations.

First mode of ocean-atmosphere variability in the tropical Atlantic Spatial patterns and expansion coefficients for the first mode of variability within the ocean-atmosphere system over tropical Atlantic regions is given in Figure 3.13. This first mode consists of maximum loadings in SSTs over eastern equatorial regions of the domain and bears resemblance with the "Atlantic Niño" or equatorial mode found in earlier studies (Zebiak, 1993; Carton and Huang, 1994; Ruiz-Barradas *et al.*, 2000). This mode contributes about 88%, 71% and 73% of the total covariance explained between SSTs and SLPs, 850 mb winds and LHF respectively.

The associated SST expansion coefficients obtained for each SVD analysis are shown in Figure 3.14 and the correlation coefficients between them in Table 3.6: all three time-series appear well correlated ($r > 0.9$) between each other, and a time-series analysis using Continuous Wavelet Transform (CWT) reveals both interdecadal and interannual variability (at about 14 years and 2.5-3/5 years). Lagged correlations with the SOI (not shown) do not exhibit strong relationship between this mode and ENSO.

Table 3.5 implies relatively strong relationship between SSTs and SLPs (-0.44) while only weaker correlations are found with 850 mb zonal winds and LHF. Lagged correlations in Figure 3.15 show a maximum relationship when U_{850mb} winds (-0.23) and SLPs (0.11) lead SST anomalies by 12 months, which could suggest non-local influences but the correlations are rather weak.

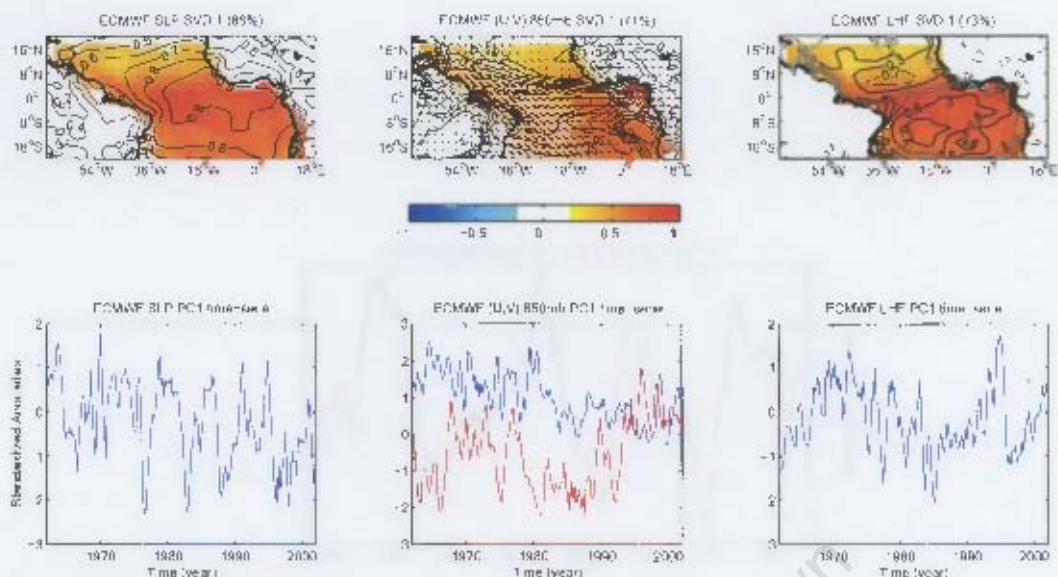


Figure 3.13: First leading modes of SVD analyses in the tropical Atlantic sector, between SSTs (in colors) and SLPs (left), 850 mb winds (middle) and LHF (left) in contours respectively. The scf is given for each mode in brackets. The spatial patterns (top) are presented as correlation maps between the corresponding expansion coefficient and anomalies at each grid point. The contours in 850 mb winds map represent correlation scores for the zonal component. The expansion coefficients for each variable (SLPs, U_{850mb}/V_{850mb} in blue/red respectively, and LHF) are displayed in the bottom panels.

Expansion coefficients	r
$SST_{SLP}/SST_{(U,V)_{850mb}}$	0.99
SST_{SLP}/SST_{LHF}	0.99
$SST_{LHF}/SST_{(U,V)_{850mb}}$	0.99

Table 3.6: Correlation coefficients between SST expansion coefficients as isolated in the first leading SVD modes in the tropical Atlantic sector over the 1962-2002 period (between SSTs and SLPs, 850 mb winds and LHF respectively).

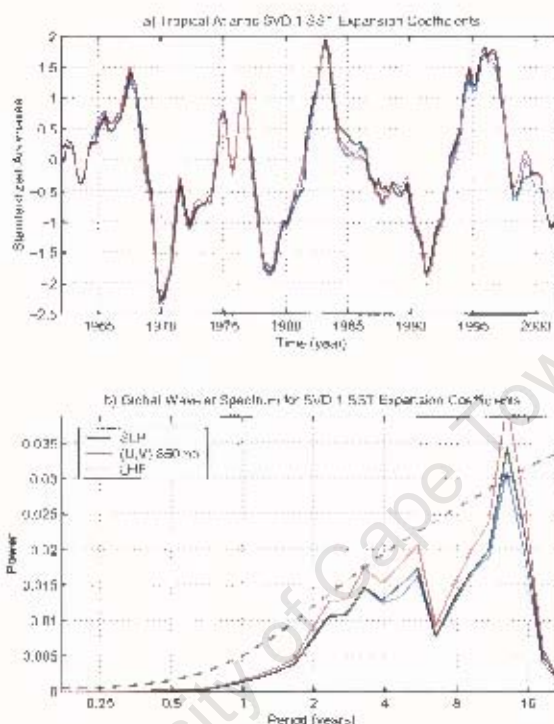


Figure 3.14: (a) SST expansion coefficients for the first leading modes of SVD analyses within the tropical Atlantic sector, between SSTs and SLPs, 850 mb winds and LHF respectively, (b) Continuous Wavelet Spectrum (CWT) for the corresponding SVD SST PC1 expansion coefficients. The 95% confidence level is indicated by the dashed black line.

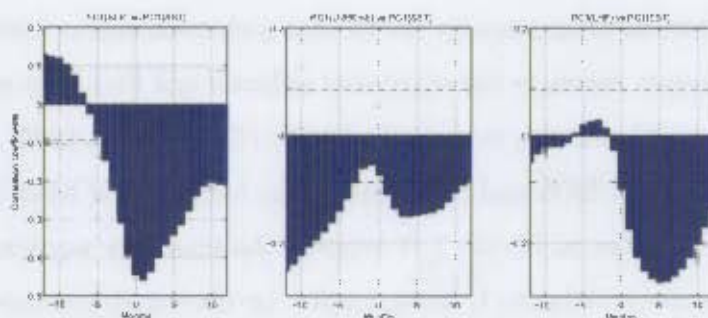


Figure 3.15: Lagged correlations between SVD 1 SST and SVD 1 SLP (left), SVD 1 U_{850mb} winds (middle), and SVD 1 LIIF (left) expansion coefficients. Significance level at 95% using Monte-Carlo simulations are plotted in dashed lines.

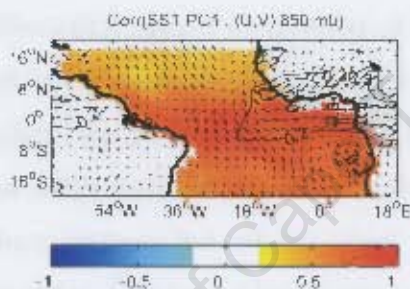


Figure 3.16: Correlations between SVD 1 SST_{SLP} expansion coefficient and the $(U, V)_{850mb}$ field. The contours represent correlations with the zonal wind component. The pattern in SSTs characteristic of this mode is plotted in colors.

The projection of the SST expansion coefficient time-series characteristic of the SST/SLP SVD analysis on the 850 mb wind field is presented in Figure 3.16. Reduced trades in the eastern equatorial Atlantic at zero lag could act to support SST warming locally, but the relationships seem quite weak. However, this is consistent with the description given in earlier studies (Merte, 1980; Zebiak, 1993; Carton and Huang, 1994; Ruiz-Barradas et al., 2000) concerning anomalous warm events in the equatorial Atlantic: equatorial wave dynamics and associated thermocline/SSTs/winds feedbacks have been proposed, in which a relaxation of the trade winds and a deepening of the equatorial thermocline prior to boreal summer warming drive an eastward current just to the south of the Equator. Nevertheless recent

observational analyses have increasingly shown that such mechanisms were partly in existence during certain events in the equatorial Atlantic and that they were much weaker in strength and less stable than in the Pacific (*Chang et al.*, 2006). Stronger relationships exist between SSTs and SLPs, maximum (-0.46) when SSTs lead SLPs by 1 month. The patterns in Figure 3.13 support the idea that warming/cooling in the equatorial eastern Atlantic is accompanied by decreased/increased surface pressures most pronounced over western regions of the basin. Such changes in zonal winds and surface pressures could suggest potential impacts on the latitudinal position of the ITCZ as proposed in *Ruiz-Barradas et al.* (2000).

Another maximum correlation is found when SSTs lead the 850 mb zonal wind field by 3-5 months (-0.15). The patterns in Figure 3.13 could illustrate changes in wind regimes most pronounced over the western parts of the tropical South Atlantic (particularly off Brazil) where it would result in reduced/enhanced northeasterlies. The maximum negative relationship found between SSTs and LHF (-0.24) with a similar timing (when SSTs lead LHF by 4-6 months) could indicate that such changes in the wind field are accompanied by reduced/enhanced evaporation acting to further warm/cool SSTs locally. It is worth noting that the different relationships found around 6 and 12 months could suggest a potential phase locking of this mode on the seasonal cycle, but here again the correlations are weak and might not be so meaningful.

Second mode of ocean-atmosphere variability in the tropical Atlantic
Spatial patterns and expansion coefficients for the second mode of variability within the ocean-atmosphere system over tropical Atlantic regions are given in Figure 3.17. This second mode consists of a dipole structure in SSTs with maximum gradient near the thermal equator, resembling the interhemispheric or dipole mode (*Hastenrath and Greischar*, 1993; *Chang et al.*, 1997; *Ruiz-Barradas et al.*, 2000). The SST pattern obtained from the SST/850 mb winds SVD analysis shows substantial discrepancies with weaker positive loadings and negative loadings extending over both hemispheres. The mode isolated between SSTs and SLP as well as LHF,

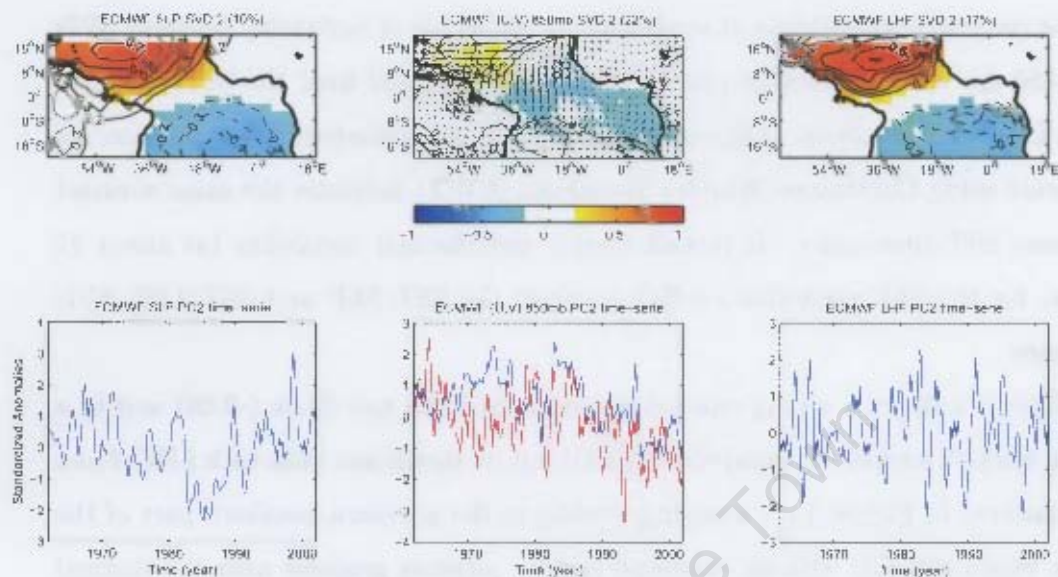


Figure 3.17: Second leading modes of SVD analyses in the tropical Atlantic sector, between SSTs (in colors) and SLPs (left), 850 mb winds (middle) and LHF (left) in contours respectively. The scf is given for each mode in brackets. The spatial patterns (top) are presented as correlation maps between the corresponding expansion coefficient and anomalies at each grid point. The contours in 850 mb winds map represent correlation scores for the zonal component. The expansion coefficients for each variable (SLPs, U_{850mb}/V_{850mb} in blue/red respectively, and LHF) are displayed in the bottom panels.

contributes about 10% and 17% of the total covariance explained respectively.

The associated expansion coefficients obtained for each SVD analysis are shown in Figure 3.18 and the correlation coefficients between them in Table 3.7. SST time-series for the SST/SLP and SST/LHF SVD analyses are well correlated with each other ($r > 0.9$) while very weak scores are found with the SST expansion coefficient issued from the SST/850 mb winds analysis. Together with the differences obtained in the geographical patterns, it suggests that the mode of variability between SSTs and 850 mb winds cannot be related to the one identified from the SST/SLP and SST/LHF SVD analyses. Time-series analysis of the corresponding expansion coefficients using Continuous Wavelet Transform (CWT) indicates the same contrast between SST time-series. It reveals mainly interdecadal variability (at about 12 years) for the SST expansion coefficients from the SST/SLP and SST/LHF SVD analyses

Table 3.5 shows a strong relationship between SSTs and SLPs (-0.68) and to a lesser extent with 850 mb zonal winds (0.23) but no significant links with LHF. From the patterns in Figure 3.17, warming/cooling in the northern/southern part of the basin would coincide with an enhanced surface pressure gradient near the thermal equator.

The projection of the SST expansion coefficient time-series characteristic of the SST/SLP and SST/LHF SVD analyses on the 850 mb wind field (Figure 3.19) exhibits a cross equatorial flow which could induce changes in the meridional position of maximum surface convergence and thus of the ITCZ (*Hastenrath and Greischar, 1993*). Figure 3.19 suggests a reduction/increase of wind speed in the hemisphere of warm/cold SSTs, that could be linked to the deflection of this cross-equatorial flow by the Coriolis force in both hemispheres (*Chang et al., 2006*).

Lagged correlations (Figure 3.20 top panel) show a maximum negative correlation when SSTs lead LHF by 3 months (-0.35) further suggesting mechanisms that could correspond with the Wind-Evaporation-SST (WES) feedback evidenced in earlier studies (*Ruiz-Barradas et al., 2000; Kushnir et al., 2002; Chang et al., 2006*): the increase of wind speed in the hemisphere of cold SSTs could lead to further

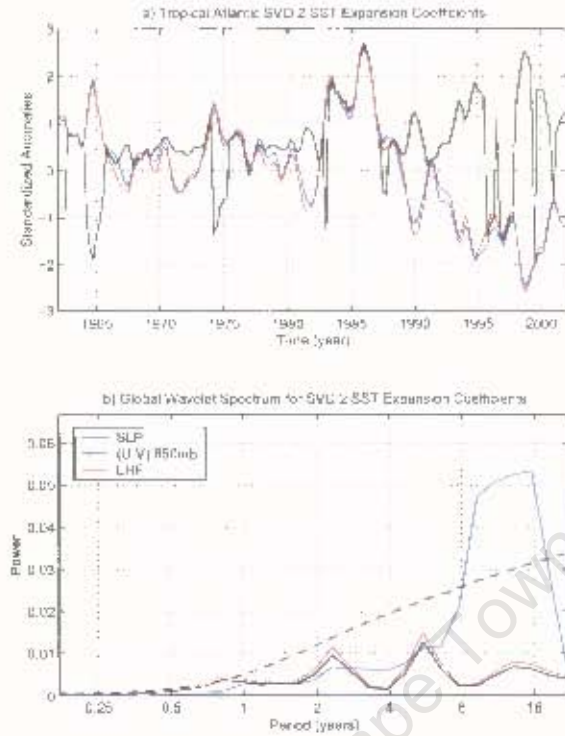


Figure 3.18: (a) SST expansion coefficients for the second leading modes of SVD analyses in the tropical Atlantic sector, between SSTs and SLPs, 850 mb winds and LHF respectively, (b) Continuous Wavelet Spectrum (CWT) for the corresponding SVD SST PC2 expansion coefficients. The 95% confidence level is indicated by the dashed black line.

Expansion coefficients	r
$SST_{SLP} / SST_{(U,V)850mb}$	0.10
SST_{SLP} / SST_{LHF}	0.99
$SST_{LHF} / SST_{(U,V)850mb}$	0.10

Table 3.7: Correlation coefficients between SST expansion coefficients as isolated in the second leading SVD modes within the tropical Atlantic sector over the 1962-2002 period (between SSTs and SLPs, 850 mb winds and LHF respectively).

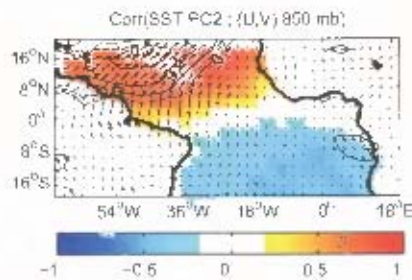


Figure 3.19: Correlations between SVD 2 SST_{SLP} expansion coefficient and $(U, V)_{850mb}$. The contours represent correlations with the zonal wind component. The pattern in SSTs characteristic of this mode is plotted in colors.

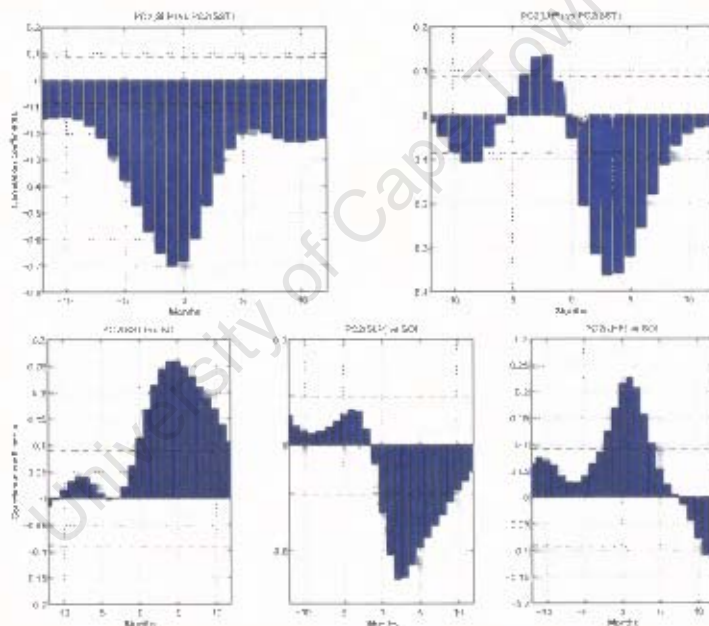


Figure 3.20: On top are presented lagged correlations between SVD 2 SST expansion coefficients and SVD 2 SLP (left) as well as SVD 2 LHF (right) expansion coefficients. In the bottom panel are shown lagged correlations between the SOI and SVD 2 SST (left), SVD 2 SLP (middle) and SVD 2 LHF (right). Significance level at 95% using Monte-Carlo simulations are plotted in dashed lines.

cooling through enhanced evaporation locally, while the decrease in wind speed in the hemisphere of warm SSTs (less pronounced in Figure 3.19 projections) would act to warm the ocean surface further.

Maximum but rather weak positive relationship between the SOI and SST SVD 2 (Figure 3.20 bottom panel) is found when warming in the Pacific leads the increased temperatures in the tropical Atlantic by 4-5 months (0.26), consistent with results from *Ruiz-Barradas et al.* (2000). Maximum correlation is found with SLPs (-0.25) and U_{850mb} winds (-0.4) when the SST anomalies in the Pacific lead by approximately 2-3 months (-0.25) while a shorter lead (1 month) characterizes the relationship between the SOI and LHF (0.22). It appears that ENSO has a relatively weak influence within this mode.

3.1.3 Summary

Using multivariate techniques, leading EOF modes of variability of the individual SST and SLP fields have been identified in the South Atlantic region (from the Equator to 50°S and from 50°W to 20°E). The principal features of these individual modes are found again in the SVD analysis between both fields which was extended to 850 mb winds and latent heat flux.

The first mode of variability describes a warming/cooling in the tropical Atlantic resulting in a shift of the region of maximum SST gradient, consistent with the findings of *Wallace et al.* (1990), *Kushnir* (1994), *Venegas et al.* (1997), *Colberg* (2006) and *Colberg and Reason* (2006). It is accompanied by a weakening/strengthening of the subtropical anticyclone modulated by reduced/enhanced pressures in the tropical South Atlantic (i.e. to the north of its climatological position). Northerly/southerly anomalies in the central basin could be the response of the wind field to the induced pressure gradient. Reduced trades particularly pronounced in the western equatorial Atlantic could be responsible for SST anomalies there but the absence of relationship over the southeastern tropics where SST anomalies also develop rather suggest non-local influences such as the propagation of equatorial Kelvin wave as proposed in earlier studies (*Carton and Huang*, 1994;

Florenchie et al., 2003, 2004; *Colberg*, 2006; *Colberg and Reason*, 2006). The induced changes in the atmospheric circulation by the anomalous SST could in turn strengthen the existing SST anomalies by air-sea heat exchanges as suggested by *Venegas et al.* (1997). This mode shows variability mainly at interdecadal and interannual time-scales (at 14 years and 4-5 years).

The second mode of variability in the South Atlantic basin illustrates the variability in the midlatitude SSTs. For SLPs it describes a weakening/strengthening and north-south displacement of the South Atlantic high pressure system due to reduced/enhanced surface pressures in the midlatitudes (i.e. to the south of its climatological position). Such changes would result in the anomalous advection of warmer and moister/colder and drier air to southwestern regions of the South Atlantic, thus reducing/enhancing latent heat flux from the ocean to the atmosphere. As the events develop, the original latent heat flux anomalies are dampened and this could be due to stronger/weaker-than-normal northerly winds in the southwest midlatitude South Atlantic, thus cooling/warming SSTs locally until positive/negative events peak. On the other hand, enhanced/reduced southeasterlies over the southeastern parts of the basin are accompanied by negative/positive latent heat flux anomalies off the west coast of southern Africa which are most developed after the events. Interdecadal as well as interannual variations are found to prevail in this mode of variability. A marked variability is found at about 5-6 years and a maximum relationship with the SOI when warming in the Pacific lead the SST signal by about 4 months emphasize links with ENSO.

Given the existence of dominant modes of variability within the tropical Atlantic sector (*Zebiak*, 1993; *Hastenrath and Greischar*, 1993; *Carton and Huang*, 1994; *Chang et al.*, 1997; *Ruiz-Barradas et al.*, 2000), SVD analyses were extended to a tropical domain in the Atlantic ranging from 20°N to 20°S and from 50°W to 20°E.

The first mode shows warming/cooling of SSTs in the equatorial Atlantic, most pronounced to the east of the domain, associated with a decrease/increase in surface pressures over western regions of the basin and reduced equatorial trades to the east.

This mode bears resemblance with the Atlantic Niño or equatorial mode (*Merle, 1980; Zebiak, 1993; Carton and Huang, 1994; Ruiz-Barradas et al., 2000*). These changes could correspond to the thermocline/SSTs/winds feedback accompanied by the latitudinal shift of the ITCZ proposed in earlier studies (*Ruiz-Barradas et al., 2000; Carton and Huang, 1994*), but the relationships identified here are rather weak which might reflect the fact that such mechanisms are much weaker and less stable than in the Pacific (*Chang et al., 2006*). Potential changes in equatorial wind regimes are also found at few months lag over western parts of the basin, particularly pronounced in the southern tropics where reduced/enhanced northeasterlies could lead to reduced/enhanced evaporation. This could act to sustain the original SST anomalies particularly in the western part of the tropical Atlantic, but the statistical relationships are also quite weak. Both interdecadal (14 years) and interannual (2.5 years) scales dominate this mode.

The second mode resembles the interhemispheric or dipole mode (*Hastenrath and Greischar, 1993; Chang et al., 1997; Ruiz-Barradas et al., 2000*). It exhibits a dipole in SSTs that coincides with a maximum pressure gradient near the thermal equator. Such changes result in cross-equatorial flow, and could then impact the meridional position of the ITCZ (*Hastenrath and Greischar, 1993*). Enhanced winds in the hemisphere of negative SST anomalies would further favour the cooling of the surface ocean there through enhanced evaporation, while reduced winds in the hemisphere of warm SSTs would lead to further warming locally. Such mechanisms have been identified in earlier studies as the WES feedback (*Ruiz-Barradas et al., 2000; Kushnir et al., 2002; Chang et al., 2006*). This mode is characterized mainly by interdecadal variability (12 years).

3.2 The Indian Ocean

The Indian Ocean with warm SSTs along the east coast of southern African is known to influence southern African climate and used to be considered as the first source of moisture over the subcontinent (*Preston-Whyte and Tyson, 1988; D'Abreton, 1992*). Nevertheless, the variability in SSTs has been less studied than in the South Atlantic.

Multivariate analyses are first applied on SST and SLP fields separately, over a domain ranging from the 20°N to 50°S and from 20°E to 120°E. Because of the known existence of tropical modes within the Indian Ocean, as documented in previous studies (*Saji et al., 1999; Webster et al., 1999*), this investigation for variability in the ocean-atmosphere system is divided between a region covering the South Indian Ocean basin (from the Equator to 50°S and from 20°E to 120°E) and a tropical sector (from 20°N to 20°S and within the same longitudinal band). As for the South Atlantic, the chosen dataset have also been smoothed using a 3 months-running-mean prior to multivariate analysis.

3.2.1 Variability within the Indian Ocean regions

In the South Indian Ocean, two main modes of variability are commonly identified in the literature. A first mode consisting of a monopole most pronounced over the central tropics is reported to be linked with ENSO thus illustrating warming in the Indian Ocean basin during El Niño events (*Wallace et al., 1998; Behera and Yamagata, 2001*). A second mode referred to as the subtropical Indian dipole (SID) is characterized by warm/cold anomalies to the south of Madagascar/off Australia respectively. Enhanced evaporation to the northeast of the basin seems to be responsible for the cooling of SSTs there while stronger-than-normal winds to the southern edge of the subtropical South Indian high pressure system would lead to reduced evaporation and thus warm SSTs. Interannual variability is found to dominate this mode which is strongly phase locked on the seasonal cycle (*Behera and Yamagata, 2001*).

The first EOF leading modes of SSTs and SLPs over the Indian Ocean region are respectively presented in Figure 3.21 and 3.22 together with their superimposed annual climatology and corresponding expansion coefficients. In accordance with the scree test (not shown), only the first two modes in SSTs/SLPs were retained. They contribute to about half of the total variance explained for each field.

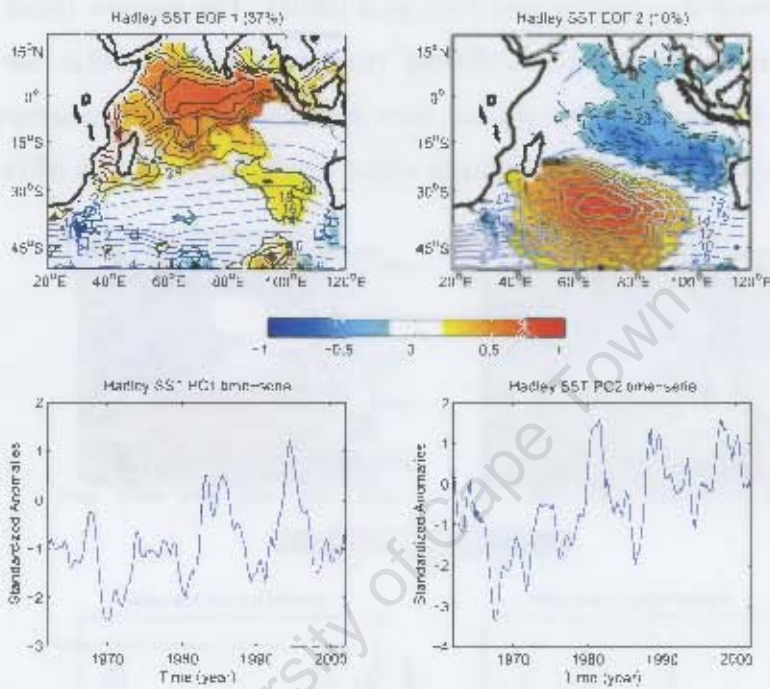


Figure 3.21: First leading EOF modes for SSTs in the Indian Ocean over the period 1962-2002. In brackets is indicated the percentage of variance explained for each mode. The spatial patterns (in colors) are presented as correlation maps (top) between the corresponding expansion coefficient (bottom) and anomalies at each grid point. The annual climatology for this period is shown in blue contours.

The first mode in SSTs (Figure 3.21) shows mostly values of single polarity over the whole domain with maximum positive correlation scores in tropical latitudes particularly in the central northern and western parts of the basin. The positive trend in the corresponding expansion coefficient time-series further suggests that this mode illustrates the recent warming in the Indian Ocean as found in earlier

studies (*Behera and Yamagata, 2001*). Strong interdecadal variability is also to be noticed in the expansion coefficient time-series of this mode. It contributes about 37% of the variance explained in SSTs.

The second mode displays an out-of-phase relationship in SSTs north and south of about 20-25°S where negative/positive anomalies are found respectively. The resulting northeast-southwest dipole structure resembles the subtropical Indian dipole mode identified in *Behera and Yamagata (2001)*. The positive trend noticed in the corresponding expansion coefficient time-series suggests that the recent warming signal noticed in the basin has been split between the two modes. This second mode explains 10% of the variance within South Indian Ocean SSTs.

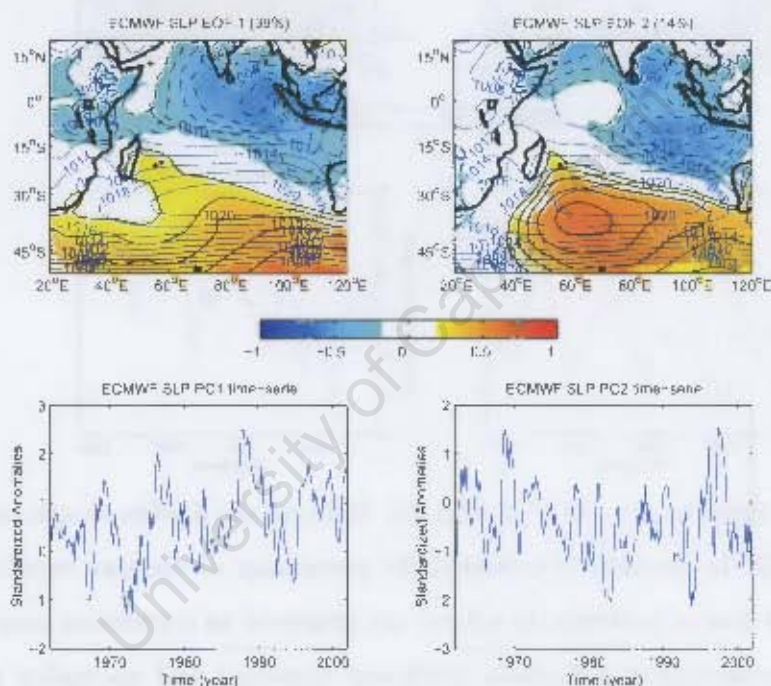


Figure 3.22: First leading EOF modes for SLPs in the Indian Ocean over the period 1962-2002. In brackets is indicated the percentage of variance explained for each mode. The spatial patterns (in colors) are presented as correlation maps (top) between the corresponding expansion coefficient (bottom) and anomalies at each grid point. The annual climatology for this period is shown in blue contours.

For SLP (Figure 3.22), the primary pattern of variability consists of maximum positive loadings to the northeast of the basin (north of about 15°S) and weaker loadings of opposite sign south of 30°S. This mode seems to reflect changes in surface pressures particularly pronounced to the northeast of the basin. It explains about 39% of the total variance in SLPs.

The second mode is characterized by a northeast-southwest dipole structure across the basin, with negative/positive anomalies to the northeast/southeast of the basin respectively. The vicinity its centres of influence to the southwest/northeast of the subtropical anticyclone centre, suggests its strengthening/weakening from the western midlatitudes/northeastern tropics, as well as its simultaneous northeast-southwest displacement. This mode contributes about 14% of the total variance explained.

Table 3.8 lists the correlation scores between SST and SLP expansion coefficients of each leading EOF modes described above, together with their correlation with ENSO using the SOI.

	PC1(SLP)	PC2(SLP)	SOI
PC1(SST)	-0.40 [8.78.10 ⁻²]	0.11 [8.67.10 ⁻²]	-0.37 [9.21.10 ⁻²]
PC2(SST)	-0.07 [8.94.10 ⁻²]	0.01 [8.96.10 ⁻²]	0.12 [9.14.10 ⁻²]
SOI	0.68 [8.88.10 ⁻²]	0.35 [9.10.10 ⁻²]	-

Table 3.8: Correlations between the expansion coefficients of the three first EOF modes in SSTs and SLPs over the Indian Ocean regions for the 1962-2002 period, together with the SOI which has been smoothed using a 3 months-running-mean for consistency. The numbers in brackets are the 95% significance levels calculated for each correlation using Monte-Carlo simulations.

The time-series of SLP PC1 and SST PC1 are significantly correlated (-0.40) and lagged correlations (Figure 3.24) show that maximum relationship (-0.55) is found when SLP's lead SSTs by about 5 months. SST PC2 and SLP PC2 are weakly correlated at zero lag (Table 3.8) and Figure 3.24 exhibits only weak maximum relationships when SSTs lead SLP's by 6-7 months (-0.2) and when SLP's lead SSTs by 3-4 months (0.14).

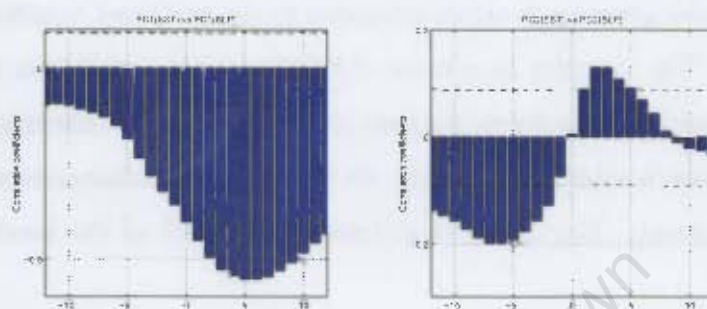


Figure 3.23: Lagged correlations for each mode between SST and SLP expansion coefficients.

The high scores obtained for SST PC1 (-0.37) and SLP PC1 (0.68) with the SOI definitely show that the Indian Ocean warming does occur in phase with ENSO. Lagged correlations (Figure 3.24) between the SOI and SST PC1 show maximum relationship (-0.50) when the warming in the Pacific leads changes in Indian Ocean SSTs by 3-4 months, which agrees with previous studies (Cadet, 1985). For SLP, maximum correlation with Pacific SSTs (0.74) is found when SLP PC1 leads by 1 month, further supporting the idea that this first mode is linked with ENSO. Regarding the second mode, only weak maximum correlations are found between the SOI and SST PC2 when the SOI leads SSTs by 3-4 months (0.20) and when SSTs lead the SOI by 6-7 months (-0.19). This could correspond to some extent with the lead exerted by the subtropical Indian dipole on ENSO at about 9 months as found by Terray and Dominiak (2005). On the other hand, the relationship between the SOI and SLP PC2 is maximum (0.30) at zero lag.

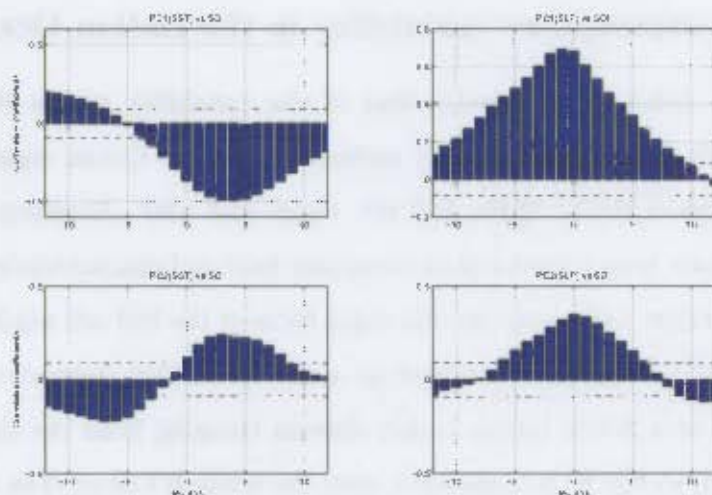


Figure 3.24: Lagged correlations between the SOI and the SST/SLP expansion coefficients (left/right) for the first and second modes (top and bottom respectively).

In conclusion, the first mode shows in SSTs a signal characteristic of the recent warming in the Indian Ocean. In phase with the warming in the Pacific, this mode is significantly correlated with ENSO, agreeing with earlier studies (Wallace *et al.*, 1998; Behera and Yamagata, 2001). A second mode in SSTs is found to consist of opposite loadings in the northeast/southwest Indian Ocean which resembles the subtropical Indian dipole identified in Behera and Yamagata (2001). The primary mode of variability in SLPs is representative of surface pressures changes particularly pronounced to the northeast of the basin. It is also significantly correlated with ENSO further emphasizing its influence on Indian Ocean regions. A second mode is found to be associated with the simultaneous strengthening/weakening of the subtropical anticyclone in the western midlatitudes/northeastern tropics and its simultaneous northeast-southwest displacement. Such findings are consistent with the overall variability in the South Indian Ocean regions described in earlier studies (Wallace *et al.*, 1998; Behera and Yamagata, 2001).

3.2.2 Ocean-atmosphere variability in the Indian Ocean

As for the South Atlantic, an investigation of the variability within the ocean-atmosphere system is presented in this section for Indian Ocean regions using SVD analyses between SSTs, SLPs, 850 mb winds and LHF. Similarly, 850 mb horizontal winds have been considered as a complex field and characteristics relative to the zonal component will constitute the main focus of the 850 mb wind analysis. Lead/lag relationships are also examined as some modes are dependant on the seasonal cycle. First a South Indian Ocean domain (ranging from the Equator to 50°S and from 20°E to 120°E) is considered, then the study is extended to a tropical sector (from 20°N to 20°S in the same latitudinal band) to isolate separately known modes of variability within the tropics (*Saji et al.*, 1999; *Webster et al.*, 1999).

3.2.2.1 Ocean-atmosphere variability within the South Indian Ocean basin

In this part, a domain slightly different from the one chosen for the separate EOF analysis of SST and SLP fields has been used, restricted to regions south of the Equator this time. In accordance with the scree test (not shown), only the first two leading modes of variability have been retained. For SLP, 850 mb winds and LHF fields, the two modes of variability with SSTs account for 88%, 70% and 69% of the total covariance explained respectively.

	SST/SLP	SST/ U_{850mb}	SST/LHF
Mode 1	0.25 [8.67.10 ⁻²]	0.08 [9.03.10 ⁻²]	-0.09 [8.66.10 ⁻²]
Mode 2	0.01 [9.52.10 ⁻²]	0.01 [8.53.10 ⁻²]	0.34 [8.30.10 ⁻²]

Table 3.9: Coupling correlation coefficients between expansion coefficient time-series corresponding to the first three leading SVD modes in the South Indian Ocean region over the 1962-2002 period. In brackets are the 95% significance level for the correlations using Monte-Carlo simulations.

Table 3.9 lists the respective coupling correlation coefficients between the expansion coefficients of both variables (i.e. between SSTs and SLPs, 850 mb winds and LHF). The first mode exhibits significant correlations between SSTs and SLPs (0.25). SST/SLP and SST/ U_{850mb} SVD 2 are characterized by particularly weak relationships, while a stronger correlation is found between SSTs and LHF (0.34).

First mode of ocean-atmosphere variability in the South Indian Ocean

Spatial patterns and expansion coefficients for the first mode of ocean-atmosphere variability over the South Indian Ocean basin are given in Figure 3.25. SST and SLP patterns and expansion coefficients time series obtained for this first mode closely resemble those obtained with the primary EOF mode of SST and SLP fields separately, despite the differences in the domains chosen for analysis. This mode seems to be related to the significant relationship found for EOF SST PC1 and EOF SLP PC1: we found a correlation of 0.98 between SST EOF 1 and SST SVD 1, while it is about -0.92 for the SLP field.

This first mode has largest loadings in SSTs for all SVD analyses over the central South Indian Ocean and is characteristic of the recent warming observed in this basin, particularly pronounced to the northeast, as reported in the previous section. This mode contributes about 75%, 50% and 51% of the total covariance explained between SSTs and SLPs, 850 mb winds and LHF respectively.

The associated SST expansion coefficients obtained for each SVD analysis are shown in Figure 3.26 and the correlation coefficients between them in Table 3.10: all three time-series appear well correlated ($r > 0.8$) between each other confirming that consistent characteristics of this first mode of variability in the South Indian Ocean region were isolated. A wavelet analysis of the corresponding SST expansion coefficients reveals variability at about 3-4 years and a decadal signal is also to be noticed.

Table 3.9 indicates significant correlation between SSTs and SLPs (0.25). From the patterns in Figure 3.25, the warming within the basin seems to coincide with enhanced pressures particularly pronounced to the northeast. Lagged correlations

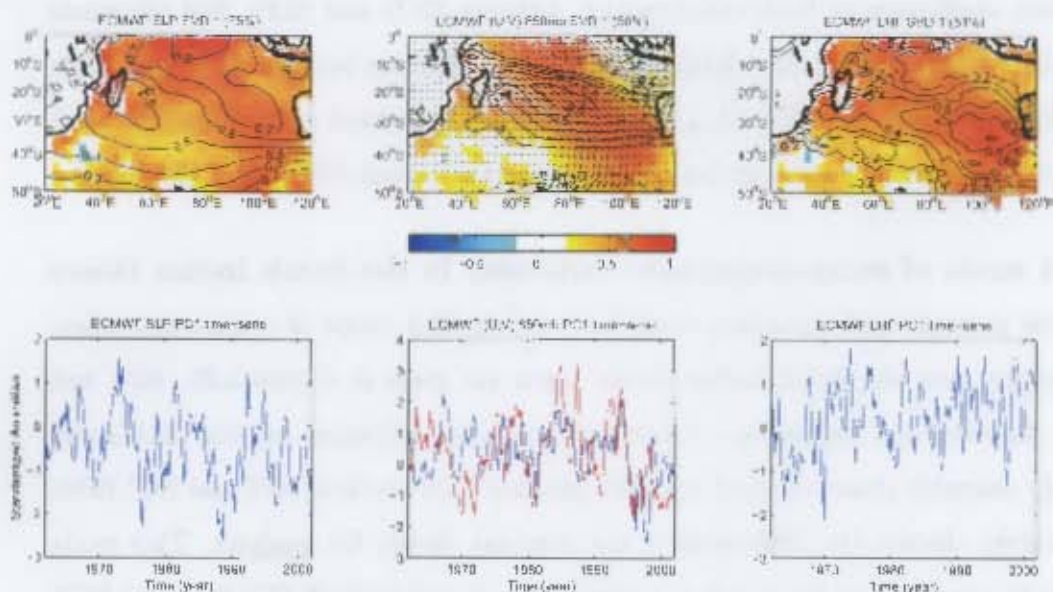


Figure 3.25: First leading modes of SVD analyses in the South Indian Ocean region, between SSTs (in colors) and SLPs (left), 850 mb winds (middle) and LHF (left) in contours respectively. The scf is given for each mode in brackets. The spatial patterns (top) are presented as correlation maps between the corresponding expansion coefficient and anomalies at each grid point. The contours in 850 mb winds map represent correlation scores for the zonal component. The expansion coefficients for each variable (SLPs, U_{850mb}/V_{850mb} in blue/red respectively, and LHF) are displayed in the bottom panels.

Expansion coefficients	r
$SST_{SLP}/SST_{(U,V)_{850mb}}$	0.83
SST_{SLP}/SST_{LHF}	0.96
$SST_{LHF}/SST_{(U,V)_{850mb}}$	0.90

Table 3.10: Correlation coefficients between SST expansion coefficients as isolated in the first leading SVD modes within the South Indian Ocean region over the 1962-2002 period (between SSTs and SLP, 850 mb winds and LHF respectively).

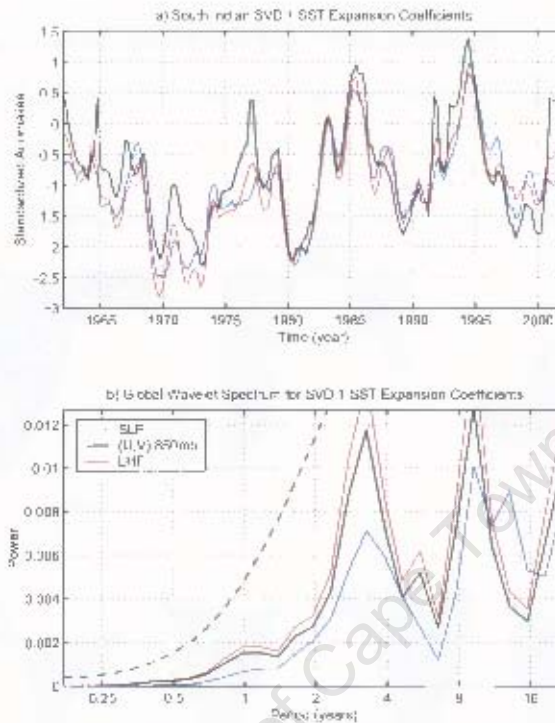


Figure 3.26: (a) SST expansion coefficients for the first leading modes of SVD analyses within the South Indian Ocean, between SSTs and SLPs, 850 mb winds and LHF respectively, (b) Continuous Wavelet Spectrum (CWT) for the corresponding SVD SST PC1 expansion coefficients. The 95% confidence level is indicated by the dashed black line.

(Figure 3.27 top panel) suggest that the relationship between the two fields is maximum (0.42) when SLP's lead SSTs by 7 months. Only weak relationships are found at zero lag between SSTs and both 850 mb zonal winds and LHF (Table 3.9), suggesting no direct effect of winds and LHF. Figure 3.27 exhibits weak maximum correlations (-0.18) when U_{850mb} winds lead SSTs by about 12 months. On the other hand, stronger maximum relationships are found when SSTs lead by 12 months modulations in 850 mb zonal winds (-0.36) and LHF (-0.38) in the central tropics. Links at 6 and 12 months could illustrate the phase locking of this mode on the seasonal cycle.

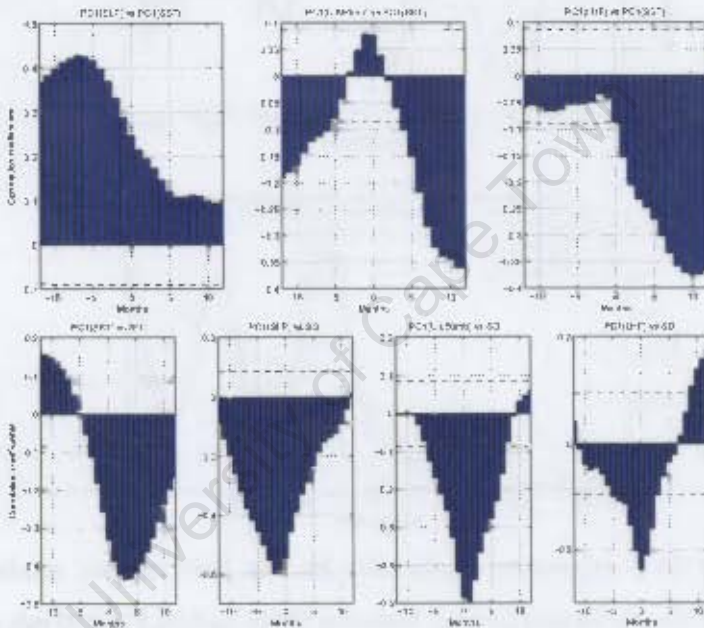


Figure 3.27: On top are presented lagged correlations between SVD 1 SST and SVD 1 SLP (left), SVD 1 U_{850mb} (middle), and SVD 1 LHF (right) expansion coefficients. In the bottom panel are shown lagged correlations between the SOI and SVD 1 SST (far left), SVD 1 SLP (left), SVD 1 U_{850mb} winds (right) and SVD 1 LHF (far right). Significance level at 95% using Monte-Carlo simulations are plotted in dashed lines.

Lagged correlations (Figure 3.27 bottom panel) between the SOI and SST SVD 1 indicate maximum relationship when the warming in the Pacific leads increased

surface temperatures in the Indian Ocean by 3-4 months (-0.43), which agrees with results from EOF analysis. For SLP, maximum correlation with Pacific SSTs is found when SLPs lead by 2 months (-0.60). The relationship between the SOI and surface winds is maximum with a month lead while it is strongest, but weak, with LHF at zero lag. These results further support the idea that this mode of variability in the South Indian Ocean is strongly linked to ENSO.

Second mode of ocean-atmosphere variability in the South Indian Ocean

Spatial patterns and expansion coefficients for the second mode of variability within the ocean-atmosphere system over the South Indian Ocean basin are given in Figure 3.28. SST and SLP patterns and expansion coefficients time series obtained for this second mode closely resemble those obtained for the second EOF modes in SSTs and SLPs separately, with correlations of 0.96 between SST EOF 2 and SST SVD 2, and 0.88 for the SLP field.

This second mode shows in SSTs, for all SVD analyses except with 850 mb winds, a dipole structure characterized by cold/warm SSTs in the northeastern/southwestern parts of the basin (i.e. off northwest Australia and to the southeast of Madagascar), resembling the subtropical Indian Ocean dipole (SID) of *Behera and Yamagata* (2001). Discrepancies are found with the SST/850 mb winds SVD analysis for which the pattern obtained in SSTs only consists of positive loadings, most pronounced to the south of Madagascar. The second mode of variability between SSTs and SLPs, 850 mb winds and LHF contributes about 13%, 20% and 18% of the total covariance explained respectively.

The associated expansion coefficients obtained for each SVD analysis are shown in Figure 3.29 and the correlation coefficients between them in Table 3.11. SST time-series for the SST/SLP and SST/LHF SVD analyses are well correlated ($r > 0.9$). The main discrepancy emerging from the SST time-series associated with the SSTs/850 mb winds SVD analysis ($r \leq 0.5$), further emphasizes that the mode of variability between SSTs and 850 mb winds cannot be related to the one isolated from the SST/SLP and SST/LHF analysis. The CWT spectra of the SST expansion

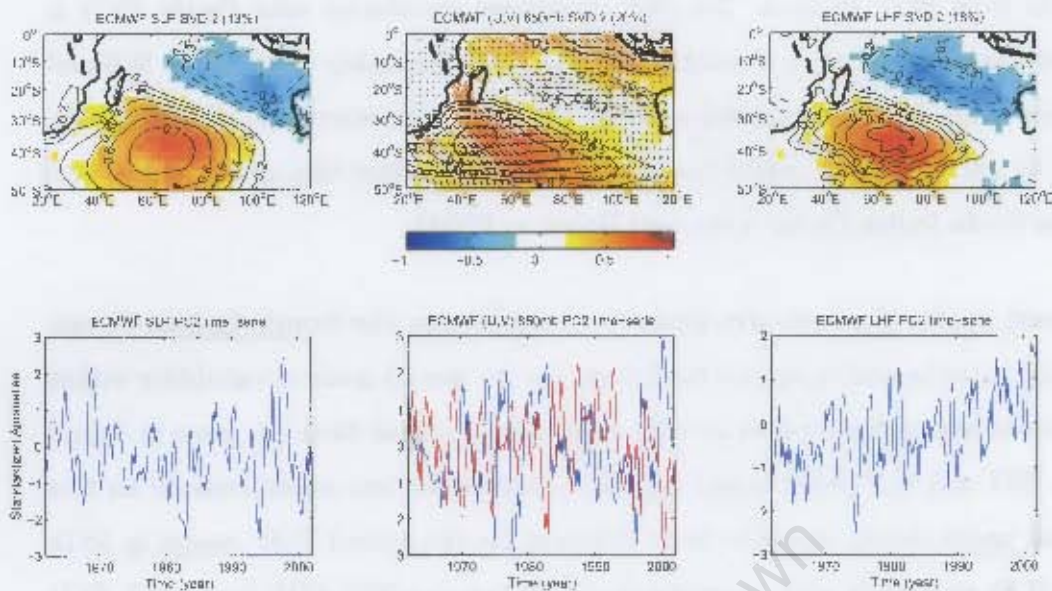


Figure 3.28: Second leading modes of SVD analyses in the South Indian Ocean region, between SSTs (in colors) and SLPs (left), 850 mb winds (middle) and LHF (left) in contours respectively. The scf is given for each mode in brackets. The spatial patterns (top) are presented as correlation maps between the corresponding expansion coefficient and anomalies at each grid point. The contours in 850 mb winds map represent the correlation scores for the zonal component. The expansion coefficients for each variable (SLPs, U_{850mb}/V_{850mb} in blue/red respectively, and LHF) are displayed in the bottom panels.

Expansion coefficients	r
$SST_{SLP}/SST_{(U,V)_{850mb}}$	0.45
SST_{SLP}/SST_{LHF}	0.93
$SST_{LHF}/SST_{(U,V)_{850mb}}$	0.51

Table 3.11: Correlation coefficient between SST expansion coefficients as isolated in the second leading SVD modes within the South Indian Ocean region over the 1962-2002 period (between SSTs and SLPs, 850 mb winds and LHF respectively).

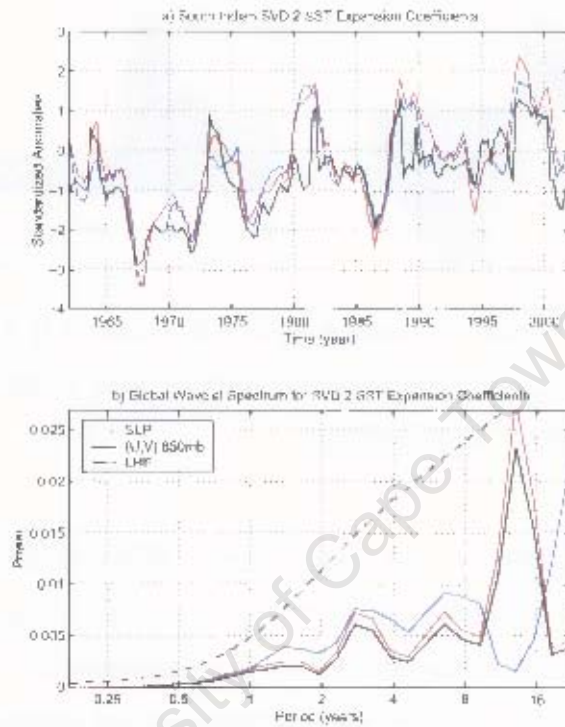


Figure 3.29: (a) SST expansion coefficients for the second leading modes of SVD analyses within the South Indian Ocean, between SSTs and SLPs, 850 mb winds and LIIF respectively, (b) Continuous Wavelet Spectrum (CWT) for the corresponding SVD SST PC2 expansion coefficients. The 95% confidence level is indicated by the dashed black line.

coefficients from the SST/SLP and SST /LHF SVD analyses reveals interannual (at about 1.5, 3 and 7 years) and interdecadal variability (at about 13 years).

Table 3.9 shows a strong relationships between SSTs and LHF (0.34) at zero lag while the other coupling coefficients are rather weak, suggesting the potential role played by evaporation in the overall mechanisms.

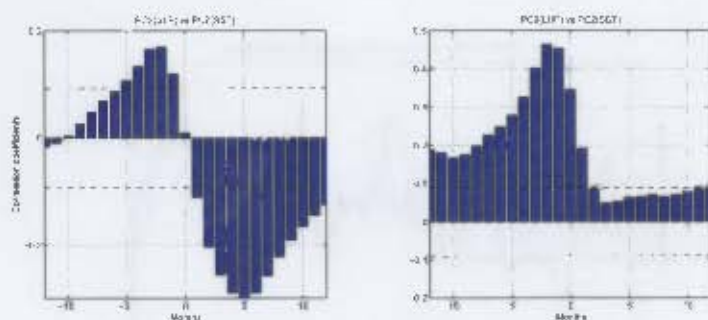


Figure 3.30: Lagged correlations between SVD 2 SST and SVD 2 SLP (left) and SVD 2 LHF (right) expansion coefficients. Significance level at 95% using Monte-Carlo simulations are plotted in dashed lines.

Positive relationships (Figure 3.30) are found between SSTs and SLPs (0.18) as well as LHF (0.48) when they lead the signal in SSTs by 2 months, suggesting that prior to positive events, the South Indian anticyclone is strengthened to the southwest of its centre. This could correspond with the establishment and migration of the west coast trough over western coastal regions of Australia in austral summer (when the Australian continent heats up), leading to a pressure gradient with the South Indian anticyclone which is shifted eastward at this period (*Behera and Yamagata, 2001*).

The projections of the SST expansion coefficient time-series characteristic of the SST/SLP SVD analysis on LHF and 850 mb wind fields (Figure 3.31), suggest for positive events, that this pressure gradient could drive strong southeasterlies off Australia, bringing anomalous colder and drier air to the northeast of the basin where evaporation and upper ocean mixing are favoured cooling SSTs locally, while to the southwest of Madagascar reduced midlatitude westerlies would support the advection of anomalously warmer and moister air implying a decrease in latent heat flux from the ocean to the atmosphere and a reduction of the equatorward

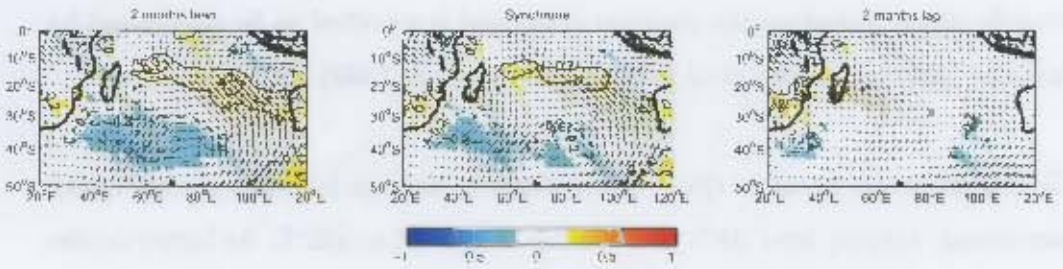


Figure 3.31: Correlations between SVD 2 SST_{SLP} expansion coefficient and U_{850mb} (streamlines) and LHF (colors and contours) at 2 months lead (left), synchronous (middle) and at 2 months lag (right).

Ekman transport of high latitudes cold waters, acting to warm SSTs there. Such findings are in accordance with the mechanisms proposed by *Behera and Yamagata* (2001). After the events peak, Figure 3.31 suggests changes in the prevailing winds accompanied by a dampening of the original LHF signal. Figure 3.30 shows negative correlations evolving between SLPs and SSTs after the events, maximum when SSTs lead SLPs by 5 months (-0.30), indicating enhanced/reduced surface pressures to the northeast/southwest of the basin. Considering the timing described in *Behera and Yamagata* (2001), this could correspond with the overall conditions at the end of austral summer when the Australian landmass cools and high pressures establish over the continent, while the subtropical high migrates northwestwards. Inferring changes in the prevailing easterlies and westerlies this could terminate the events (*Behera and Yamagata*, 2001).

3.2.2.2 Ocean-atmosphere variability within the tropical Indian Ocean sector

In tropical regions of the South Indian Ocean, recent studies have evidenced a mode of variability which is referred to as the Indian Ocean dipole (IOD) mode (*Saji et al.*, 1999) or Indian Ocean zonal (IOZ) mode (*Webster et al.*, 1999). It consists of a dipole-like pattern with cold SSTs off Sumatra and warm SSTs in the western parts of the basin. *Saji et al.* (1999) found that the intensity of the SST dipole

is linked to the strength in zonal wind anomalies in the central basin. This mode is strongly phase locked on the seasonal cycle and is reported to be dominated by interannual variability (*Saji et al.*, 1999; *Webster et al.*, 1999).

To integrate this mode in this study, a tropical domain has been chosen in the Indian Ocean, ranging from 20°N to 20°S and from 20°E to 120°E. As higher modes appeared to be statistically degenerate, only the first leading mode has been retained. It accounts for 68%, 56% and 45% of the total covariance explained between SSTs and SLP, 850 mb winds and LHF respectively.

	SST/SLP	SST/ U_{850mb}	SST/LHF
Mode 1	-0.49 [8.54.10 ⁻²]	-0.18 [8.76.10 ⁻²]	0.07 [8.83.10 ⁻²]

Table 3.12: Coupling correlation coefficients between expansion coefficient time-series corresponding to the first leading SVD mode in the tropical Indian Ocean over the 1962-2002 period. In brackets are the 95% significance level for the correlations using Monte-Carlo simulations.

Table 3.12 exhibits negative strong relationships between SSTs and SLPs and a weaker correlation with 850 mb zonal winds while the scores obtained with LHF are closer to the 95% significance level value.

Spatial patterns and expansion coefficients for this first mode of variability within the ocean-atmosphere system in the tropical Indian Ocean are given in Figure 3.32. This mode consists in SSTs of positive/negative anomalies off the east coast of Africa and to the south and west of Sumatra respectively. It resembles the tropical Indian Ocean dipole mode of *Saji et al.* (1999), but some differences are to be noticed regarding the extent of the eastern negative loadings which are not so pronounced to the west of Sumatra in the present investigation. Discrepancies are also found for the SST/850 mb SVD analysis, showing relatively weaker negative loadings in SSTs to the east of the basin.

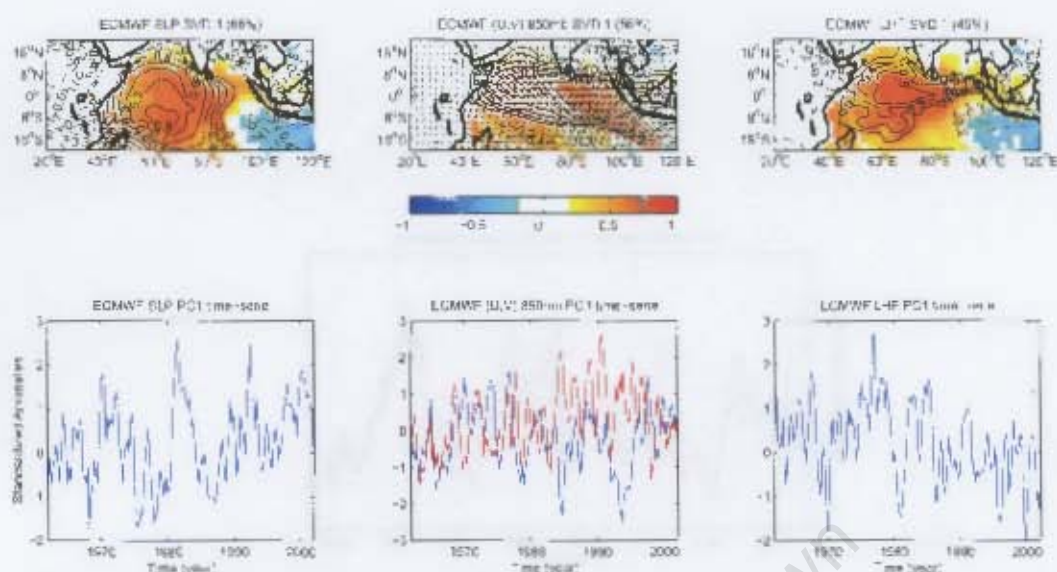


Figure 3.32: First leading modes of SVD analyses in the tropical Indian Ocean sector, between SSTs (in colors) and SLPs (left), 850 mb winds (middle) and LHF (left) in contours respectively. The scf is given for each mode in brackets. The spatial patterns (top) are presented as correlation maps between the corresponding expansion coefficient and anomalies at each grid point. The contours in 850 mb winds map represent correlation scores for the zonal component. The expansion coefficients for each variable (SLPs, U_{850mb}/V_{850mb} in blue/red respectively, and LHF) are displayed in the bottom panels.

	SST_{SLP}	$SST_{(U,V)_{850mb}}$	SST_{LHF}	SOI
SST_{SLP}	-	0.80	0.97	-0.36
$SST_{(U,V)_{850mb}}$	0.80	-	0.84	-0.41
SST_{LHF}	0.97	0.84	-	-0.52
DMI	0.61	0.32	0.64	-0.36

Table 3.13: Correlation coefficients between SST expansion coefficients of the first leading SVD modes in the tropical Indian Ocean over the 1962-2002 period (between SSTs and SLPs, 850 mb winds and LHF respectively) and with DMI and the SOI.

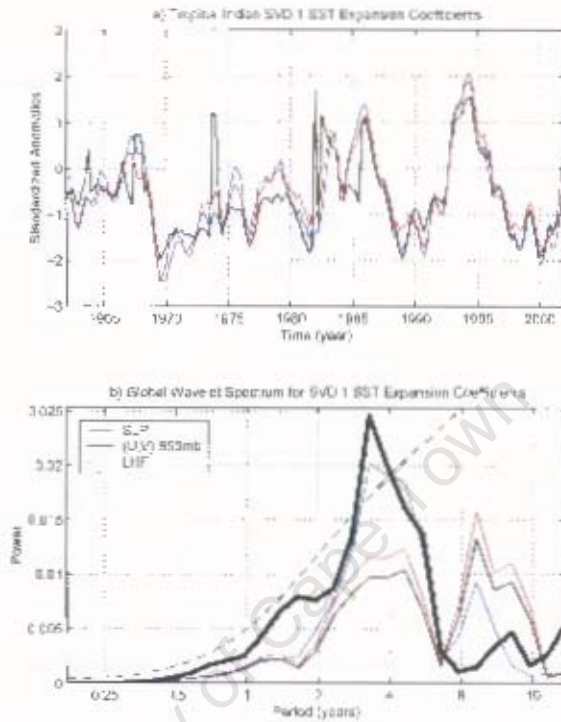


Figure 3.33: (a) SST expansion coefficients for the first leading modes of SVD analyses within the tropical Indian Ocean, between SSTs and SLPs, 850 mb winds and LHF respectively, (b) Continuous Wavelet Spectrum (CWT) for the corresponding SVD SST PC1 expansion coefficients and for the Indian Ocean Dipole Index Mode (DMI) in thick black line. The 95% confidence level is indicated by the dashed black line.

The associated expansion coefficients obtained for each SVD analysis are shown in Figure 3.33 and the correlation coefficients between them in Table 3.13: all three time-series appear well correlated ($r > 0.8$) between each other, confirming that consistent characteristics of this first mode of variability within the tropical Indian Ocean regions were isolated. Nevertheless, lower correlation coefficients with the SST expansion coefficient and discrepancies in the SST loadings from the SST/850 mb winds SVD analysis suggest that this mode of variability between SSTs and 850 mb winds cannot be closely related to the one identified from SST/SLP and SST/LHF SVD analyses.

The reversal in sign of SST anomalies across the basin is identified by *Saji et al.* (1999) using a simple index between the tropical Indian Ocean [10°S - 10°N ; 50°E - 70°E] and tropical southeastern regions of the basin [Equator- 10°S ; 90°E - 110°E]. Table 3.12 shows significant correlations between the dipole mode index (DMI) and the SST time-series associated with the SST SVD 1 expansion coefficients ($r > 0.6$ except for the SST/850 mb winds SVD analysis). Wavelet analysis of the corresponding SST expansion coefficients reveals variability at about 3.5 years corresponding well with the computed DMI signature. Nevertheless, the spectral analysis failed to recover the quasi-biennial signal at 2.5 years identified by *Saji et al.* (1999) which has been contested. A weaker but non-significant peak is also found just above 8 years.

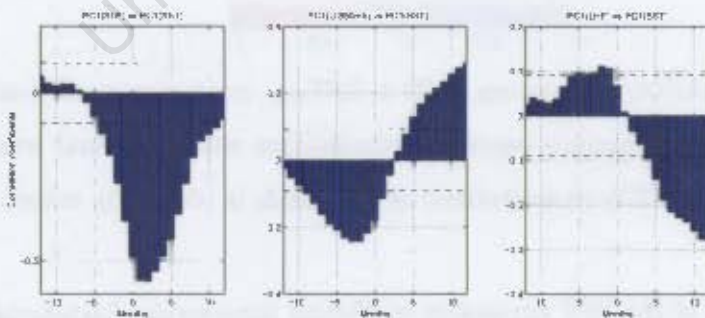


Figure 3.34: Lagged correlations between SVD 2 SST and SVD 2 SLP (left), SVD 2 U_{850mb} winds (middle), and SVD 2 LHF (right) expansion coefficients. Significance level at 95% using Monte-Carlo simulations are plotted in dashed lines.

Table 3.12 exhibits a significant correlation between SSTs and SLPs (-0.49) at zero lag. From the patterns in Figure 3.32, the warming/cooling within the western/eastern parts of the basin would coincide with reduced/enhanced surface pressures respectively.

Lagged correlations (Figure 3.34) show maximum relationship between SSTs and SLPs (-0.55) when SLPs lead by 1-2 months, suggesting the establishment of a pressure gradient across the basin prior to the events. Weaker negative scores characterize the correlation between SSTs and 850 mb zonal winds at zero lag (-0.18) as shown in Table 3.12. Nevertheless, maximum relationships between SSTs and U_{850mb} winds as well as LHF are found when they lead SSTs by about 2 months (-0.22 and 0.12 respectively) and when SSTs lead by 12 months (0.30 and -0.27 respectively). The relationships are rather weak, but the later could illustrate a substantial phase locking of this mode on the seasonal cycle. Previous studies have emphasized similar characteristics for the Indian Ocean dipole which is reported to peak in austral spring (Webster et al., 1999; Saji et al., 1999; Basquero-Bernal et al., 2002; Fauchereau, 2004).

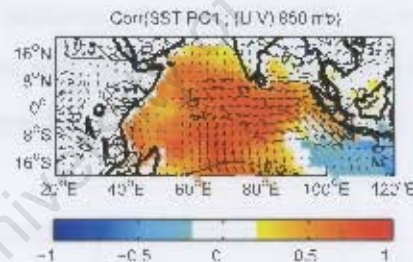


Figure 3.35: Correlations between SVD 1 SST_{SLP} expansion coefficient and the $(U, V)_{850mb}$ field. The contours represent correlations with the zonal wind component. The pattern in SSTs characteristic of this mode is plotted in colors.

The projection of the SST expansion coefficient time-series characteristic of the SST/SLP SVD analysis on the 850 mb wind field is presented in Figure 3.35. Enhanced/reduced southeasterlies, which could be driven by the gradient in surface pressures across the basin, are found to the southeastern part of the tropical Indian Ocean for positive/negative dipole events, agreeing with Saji et al. (1999).

The results presented in Figure 3.36 would further support the idea that the intensity of the SST dipole and strength in the zonal wind anomalies are dependant one from another, in accordance with the findings of *Saji et al. (1999)*. Finally, the correlation with the SOI shown in Table 3.13 for the DMI (-0.36), suggests that this mode of variability is to some extent also dependant of ENSO, consistent with previous studies (*Basquero-Bernal et al., 2002; Fauchereau, 2004*).

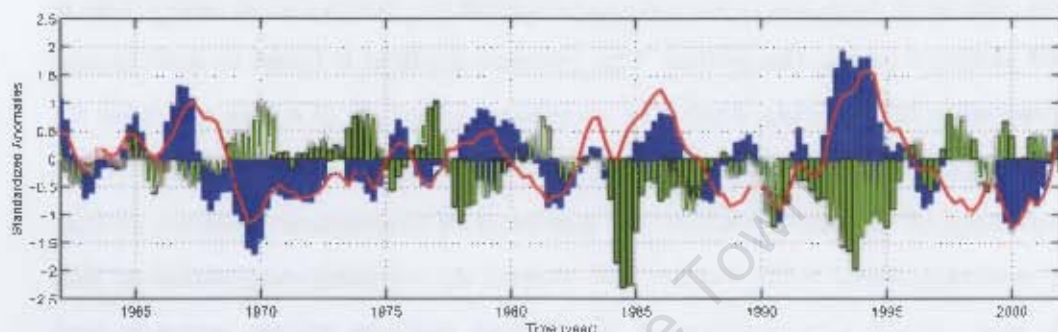


Figure 3.36: Mean SST (red line) and 850 mb zonal wind (green bars) expansion coefficients for the first leading SVD mode in the tropical Indian Ocean sector, together with Indian Ocean Dipole Index Mode (DMI) in blue bars. The DMI has been smoothed using a 3 months-running-mean window for consistency.

3.2.3 Summary

The use of multivariate techniques over the broad Indian Ocean basin helped identify leading EOF modes of variability within the individual SST and SLP fields. The main features characteristic of these individual modes are found again to some extent in the SVD analysis between both fields and which we extended to 850 mb winds and latent heat flux over South Indian regions.

The first mode of variability can be described by the recent warming within the Indian Ocean basin which coincides with enhanced surface pressures particularly pronounced to the northeast of the basin. This mode is also dominated by inter-annual variability (at about 3-4 years) and substantial links with ENSO are found

when warming in the Pacific leads by 4 months.

The second mode of variability is characterized by a dipole structure in SSTs consisting of warm/cold SSTs to the southwest/northeast of the basin, thus resembling the subtropical Indian dipole mode (SID) of *Behera and Yamagata* (2001). The development of positive events seems to be preceded by enhanced/reduced surface pressures to the southwest/northeast of the basin respectively, which could correspond to the establishment and migration of the west coast trough over western coastal regions of Australia in austral summer while the South Indian anticyclone is shifted eastward during this period. This pressure gradient is found to drive strong southeasterlies off Australia where the anomalous advection of colder and drier air could result in enhanced evaporation and upper ocean mixing thus warming SSTs to the northeast of the basin, while to the southwest of Madagascar, reduced midlatitude westerlies would bring warmer and moister air, reducing evaporation as well as the equatorward Ekman transport of cold high latitudes waters, acting to cool SSTs there (*Behera and Yamagata*, 2001). At the end of austral summer, when the Australian landmass cools and high pressures establish over the continent, the subtropical high migrates northwestwards, inferring shifts in the prevailing easterlies and westerlies that could act to terminate the event (*Behera and Yamagata*, 2001). A wavelet analysis helps reveal both interannual (mainly at 3 and 7 years) and interdecadal (at about 13 years) variability associated with this mode.

Due to the existence of a zonal mode in the tropical Indian Ocean sector (*Saji et al.*, 1999; *Webster et al.*, 1999), SVD analyses were applied over a domain ranging from 20°N to 20°S and from 20°E to 120°E. A primary mode consists of SST anomalies of opposite signs off the east coast of Africa and west of Sumatra, that bears resemblance with the Indian Ocean tropical dipole, despite some differences on the extent of the eastern negative loadings. An east-west surface pressure gradient across the basin is found to coincide with modifications in the easterly wind regime in the central basin. The use of an index between the tropical Indian Ocean and tropical southeastern regions as in *Saji et al.* (1999), confirms a strong relationship between the intensity of the SST dipole and the strength in zonal wind

anomalies at equatorial latitudes. This mode appears to be strongly phase locked on the seasonal cycle having its peak in austral spring (*Webster et al.*, 1999; *Saji et al.*, 1999; *Basquero-Bernal et al.*, 2002; *Fauchereau*, 2004). A CWT analysis indicates that interannual variability (around 3.5 years) dominates this mode and that potential connections exist between this mode of variability and ENSO.

Primary modes of variability within the ocean-atmosphere system in the South Atlantic and Indian Ocean regions have been identified in this part. They will be used as a basis in the following to identify connections between water vapour transport variability over southern Africa and variations within the neighbouring ocean-atmosphere systems in each basin.

Chapter 4

Contribution of the oceans in water vapour input and rainfall variability over southern Africa

In the following section, zonal water vapour transport over southern Africa in summer is investigated. Calculations of moisture fluxes and convergence helped to identify key areas for deep convection and main components entering into play in the zonal advection of moisture from the South Atlantic and Indian Oceans onto the subcontinent (see Chapter 2). Multivariate analyses are then used on zonal moisture fluxes along the west (Atlantic) and east (Indian) coasts of southern Africa to examine how their variability can modulate local rainfall during this season. Further connections are emphasized with primary modes of ocean-atmosphere variability within the South Atlantic and Indian Ocean extracted in the previous chapter in order to identify these regions of importance over the basin in direct vicinity of the coast considered for each modes of zonal moisture fluxes. In this respect tropical modes of ocean-atmosphere variability in each basin region as well as cross influences, i.e. from the Indian/South Atlantic Ocean for west/east coast moisture fluxes variability, have not been considered in this part.

4.1 Role of the South Atlantic Ocean

Considering the humidity input from the South Atlantic towards the subcontinent as mainly zonal, at least over tropical latitudes, moisture fluxes were averaged zonally for grid points along the west coast of southern Africa (ranging from 2.5°N to 37.5°S in latitude and from 7.5°E to 15°E in longitude) over the 1979-2000 period. The variability within zonal water vapour transport is then examined using Empirical Orthogonal Functions (EOF) for this wall-like domain along the west coast. As higher modes seem degenerate, only the first two modes of variability have been retained and they are described in more detail in this section.

The influence of the South Atlantic on southern African climate has been evidenced in previous studies (*Hirst and Hastenrath*, 1983; *Walker*, 1990; *Mason*, 1995; *Jury*, 1996; *Reason and Godfred-Spenning*, 1998; *Camberlin and Poccard*, 2001; *Todd and Washington*, 2004; *Rouault et al.*, 2003; *Reason and Jagadheesha*, 2005a; *Reason et al.*, 2006). The transition season between early and late summer rainfall is of great importance to southern African climate. Using lead/lag correlations between the expansion coefficients of each mode and rainfall data, it is the period for which the most significant relationships are obtained, which was confirmed by the temporal coherence found with the corresponding signal in land hydrology data (NDVI and soil moisture). Figure 4.1 presents January-February vector moisture fluxes maps with contours of mean humidity convergence at 850 mb, 700 mb and 500 mb. The mean January-February climatological structure for zonal moisture flux along the west coast of southern Africa is shown in Figure 4.2. It is worth noting that the transient term represents less than 10% of the total zonal moisture flux, thus the stationary component will be well suited to represent changes in moisture fluxes due to large scale circulation. As described in Chapter 2, a westerly flux is found in summer from the equator to 15°S at surface levels. Overlying this westerly flux, the southern extension of the African Easterly Jet (AEJ) at midtropospheric levels is located just above 700 mb in January-February.

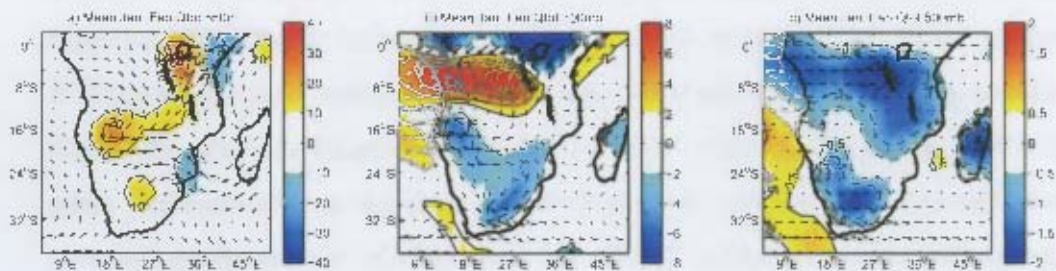


Figure 4.1: Mean January-February surface moisture fluxes (streamlines in $g.kg^{-1}.m.s^{-1}$ with arrows scaled at 1 unit/degree of latitude) over the 1979-2000 period, together with contours of moisture convergence (in $g.kg^{-1}.s^{-1}$) at 850 mb (a), 700 mb (b) and 500 mb (c). Positive values contour areas of moisture convergence while negative values refer to moisture divergence at given levels.

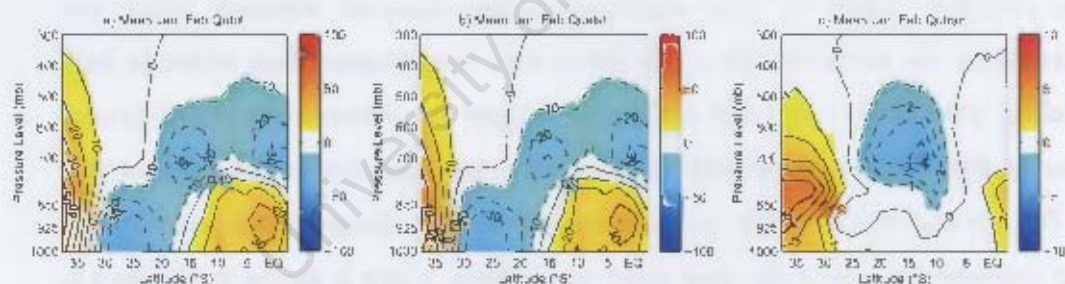


Figure 4.2: Mean vertical structure (a) of zonal moisture fluxes along the west Southern African coast (in $g.kg^{-1}.m.s^{-1}$) for January-February over the 1979-2000 period, together with its stationary (b) and transient (c) components. Positive values correspond to westerly fluxes while negative values refer to easterly fluxes.

To the south (between 17°S and 32°S) an easterly flux is driven by the South Atlantic anticyclone, connecting to the southern AEJ in particular during summer. Finally, south of 32°S, a westerly moisture flux occupies most of the air column.

Using the EOF method, leading modes of variability for zonal winds, specific humidity and zonal moisture fluxes were extracted at each tropospheric levels, from 1000 to 300 mb, over the region ranging from 2.5°N to 37.5°S and from 5°E to 50°E. The stability of a time series being essential in a multivariate analysis approach, the stationarity of these primary modes by levels is studied using a non-parametric statistical method, the Mann-Whitney-Pettitt test (see Chapter 2) on their expansion coefficients. Such method has been used successfully in earlier studies to examine stationarity of climatic series (*Vandiepenbeeck, 1996*).

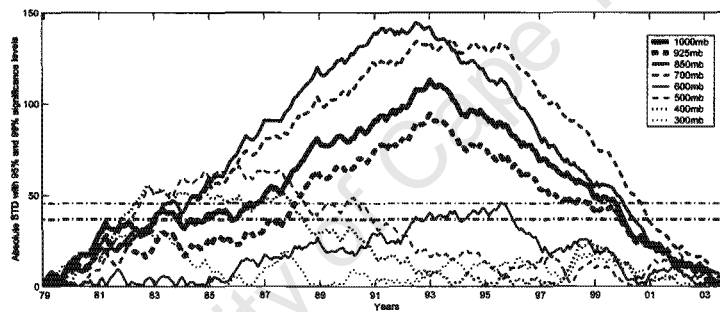


Figure 4.3: Absolute standard deviations from Pettitt test on first EOF leading modes expansion coefficients of zonal moisture fluxes at tropospheric levels (from 1000 mb to 300 mb) computed from NCEP R2 dataset for the domain [2.5°N-37.5°S;05°E-50°E] over the 1979-2003 period, together with 95% and 99% significance levels (bold and light dashdotted lines). Significant inflection points on the absolute standard deviation curves correspond to abrupt shifts in the time-series.

Figure 4.3 shows major abrupt shifts for the computed zonal moisture fluxes:

- (a) in 1993 at lower tropospheric levels, also identified in zonal wind (not shown),
- (b) in 1995 at 700 and 600 mb, also in zonal wind and specific humidity (not shown),
- (c) in 1986 at 500 and 400 mb, also in specific humidity (not shown).

Further investigation is needed to understand the origins for these instabilities within the time-series. However, it is worth noting that major breaks in stationarity identified in earlier version of NCEP re-analyses have been shown to be related to the inclusion/exclusion of observation data in the re-analysis procedures (*Poccard, 2000; Camberlin and Poccard, 2001*). To overcome these ruptures in stationarity, the high frequency filtered zonal moisture fluxes were extracted using a high-pass Butterworth filter at 96 months. When submitted to a similar Pettitt-test (not shown), the filtered data do not exhibit anymore statistically significant ruptures, except for the 400 mb level showing the remaining marginal break. Because most of transport features considered in the following are located below 400 mb, this post-processing procedure appears quite satisfying in regards of the objectives of this study. Consequently, this study will focus in the following on time scales of variability below 8 years.

EOF techniques were then applied to the high-filtered (HF) zonal moisture fluxes along the west coast of southern Africa using the covariance matrix. This approach has the advantage in the present context to respect the relative importance of lower tropospheric levels in moisture transport compared to higher altitudes. The first two leading EOF modes were retained in accordance with the scree test (not shown) and their associated spatial patterns are shown in Figure 4.4 for all seasons.

The first mode, which will be referred to as the midlatitude South Atlantic mode, is typical of variations in intensity and of the latitudinal migration of the circulation linked with the midlatitude westerlies and South Atlantic anticyclone. It explains about 24.8% of the total variance. Positive loadings are particularly strong from the surface up to 600 mb and are associated with patterns of opposite sign between 10°S and 20°S. Given the filtering applied, the wavelet spectra for its corresponding expansion coefficient (not shown) reveals variability mostly intra-annual as well as high-frequency seasonal signals (at 10 and 4-6 months), but also at interannual timescales (about 5-6 years).

The second mode, which will be referred to as the equatorial Atlantic westerly mode, characterizes the westerly moisture flux in the tropics, from the Equator to about 15°S. It represents 10.9% of the total variance explained. It has a simultaneous loading at the surface between 20°S and 30°S. As for the previous mode, time-series analysis using the CWT (not shown) exhibits mostly intra-annual and interannual variability for this mode (just below 1 year and at about 4-5 years respectively).

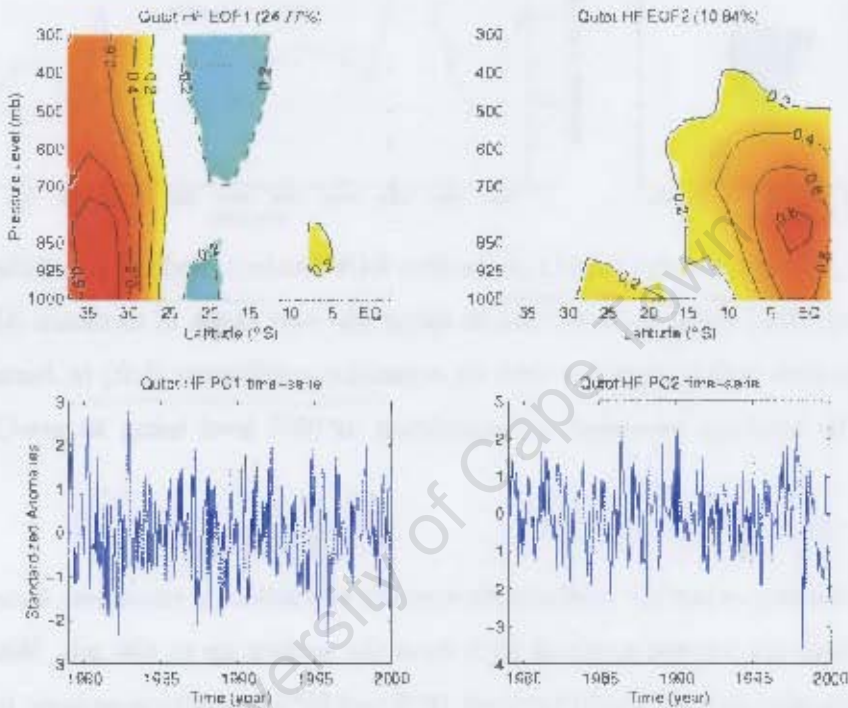


Figure 4.4: Spatial patterns (top) of the first two EOF leading modes of variability in high frequency (HF) zonal moisture fluxes along the west coast of southern Africa over the 1979-2000 period, together with their respective expansion coefficients (bottom) for all seasons.

In the following, this study will focus on the transition season between early and late summer rainfall (January-February) and the projection of these modes during this period will be discussed.

4.1.1 The midlatitude South Atlantic mode

Figure 4.5a shows the spatial pattern of the first EOF leading mode in high-filtered (HF) zonal moisture fluxes along the west coast of southern Africa during the January-February period.

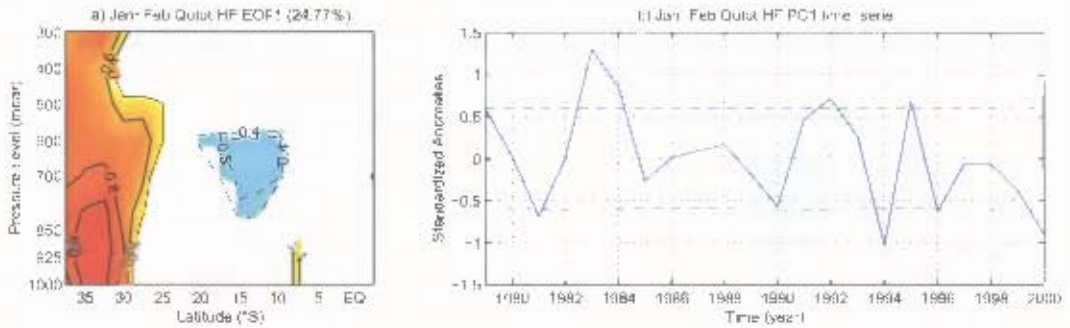


Figure 4.5: Spatial patterns (right) of the first EOF leading mode of variability in high frequency (HF) zonal moisture fluxes along the west coast of southern Africa over the 1979-2000 period, together with its expansion coefficients (left) in January-February. The loadings presented are significant at 95% level using Monte-Carlo simulations.

In January-February, when the midlatitude westerly circulation is sustained, strongest positive loadings are located south of 25°S from the surface up to 600 mb. Weaker loadings of opposite sign are found between 10°S and 20°S at mid-tropospheric levels (between 700 mb and 500 mb) where prevails the southern extension of the AEJ in summer. Most pronounced in austral summer, it is replaced by a low level easterly flux in winter (see Chapter 2) and this could be an explanation for the absence of loadings over the southern AEJ location in Figure 4.4. Finally, positive/negative events of the midlatitude South Atlantic mode in January-February correspond with enhanced/reduced midlatitude westerly circulation accompanied by the strengthening/weakening of the moisture fluxes linked with the southern dependance of the AEJ.

In order to investigate the potential impact of the midlatitude South Atlantic mode on January-February rainfall over the subcontinent, direct rainfall estimates from the CRU TS 2.0 dataset are used and have been filtered at 8 years for consistency.

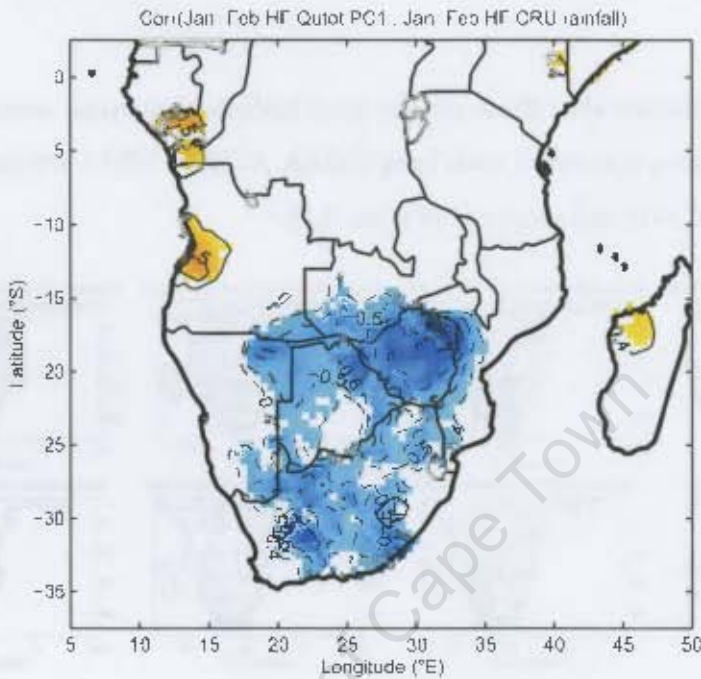


Figure 4.6: Heterogeneous correlations between HF zonal moisture fluxes PC1 and CRU rainfall in January-February over the 1979-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

Heterogeneous correlations between the January-February midlatitude South Atlantic mode expansion coefficient (JF MFPC1 in the following) and CRU rainfall are presented in Figure 4.6. High negative loadings for synchronous correlation are found to the south of the uplands surrounding the Congo basin, with maximums over Zimbabwe and Botswana stretching southward west of the east coast escarpment to South Africa and westwards to Namibia. In addition, significant negative maximums are found over the Eastern Cape and the Northern Cape provinces of South Africa. In contrast, the Kalahari regions show lower loadings, which might reflect uncertainties within CRU data locally. Positive loadings prevail over western

coastal regions of tropical southern Africa, typically along the coast stretching from south Gabon to south Angola. Rainfall composites (not shown) for positive and negative years of the JF MFPC1 time-series (Figure 4.5b) confirm the symmetry of the relation with significant anomalies over these areas of maximum correlation scores.

To further emphasize the effects on the local hydrological cycle, heterogeneous correlations have been computed with both NOAA AVHRR NDVI v.2 dataset and NOAA NCEP CPC GM soil moisture (Figure 4.7).

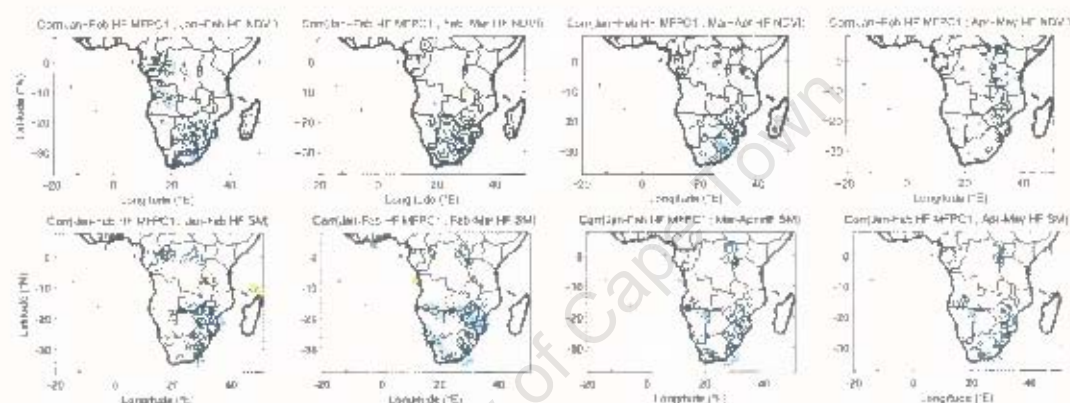


Figure 4.7: Heterogeneous correlations between HF January-February zonal moisture fluxes PC1 and NDVI (top) as well as soil moisture (bottom) in January-February (far left), February-March (left), March-April (right) and April-May (far right) over the 1982-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

Maximum negative relationships are found from January-February to February-March in both NDVI and soil moisture. In NDVI the signal is maximum to the south of the Congo basin, across Zimbabwe and Botswana stretching southward to South African regions. In soil moisture maximum loadings are located to the southeast of the subcontinent, particularly pronounced over northeast South Africa, southern Mozambique and Zimbabwe, while weaker but still significant loadings are found over south Angola and south Namibia. Differences over central regions of

South Africa (i.e. Karoo) between NDVI and soil moisture might reflect discrepancies in soil moisture data that could result from uncertainties in rainfall data used for computation over these regions. Composites for positive/negative events (not shown) from the JF MFPC1 time-series also confirm the symmetry of the relation with corresponding significant anomalies. This suggests that impacts on rainfall regimes to the south of the Congo basin result with about 2 months lag in changes within the vegetation cover for these regions. Such results are in agreement with *Farrar et al.* (1994) who found that globally the correlation between NDVI and rainfall is highest when considering precipitated amounts from the concurrent and previous 2 months.

4.1.1.1 Connections with ocean-atmosphere variability in the South Atlantic basin

In the previous chapter, dominant modes of ocean-atmosphere variability within the South Atlantic Ocean basin from 1962 to 2002 have been presented. Since 1979-2000 is the period of interest in the following, leading modes have been re-computed using SVD analysis on ECMWF SLP re-analyses data and Hadley SSTs from 1979 to 2000, in order to investigate potential relationships in variability between zonal moisture fluxes along the west coast of southern Africa and the South Atlantic ocean-atmosphere system over this period. The two first leading modes are presented in Figure 4.8. The respective patterns of variability between SST and SLP fields from 1979 to 2000 resemble greatly the one isolated in Chapter 3 for the 1962-2002 period. The SST expansion coefficients issued from both SVD analyses appear to be significantly well correlated with scores of about 0.99 and 0.86 for the first and second mode respectively. Such findings give confidence in the fact that the modes found over the 1979-2000 period are consistent with the ones obtained previously for 1962-2000. Typically the first mode shows warming/cooling within the tropical Atlantic resulting in a shift of the region of maximum SST gradient. Such changes are associated with modulations in intensity of the South Atlantic anticyclone due to reduced/enhanced surface pressures in the tropical South Atlantic (i.e. to the

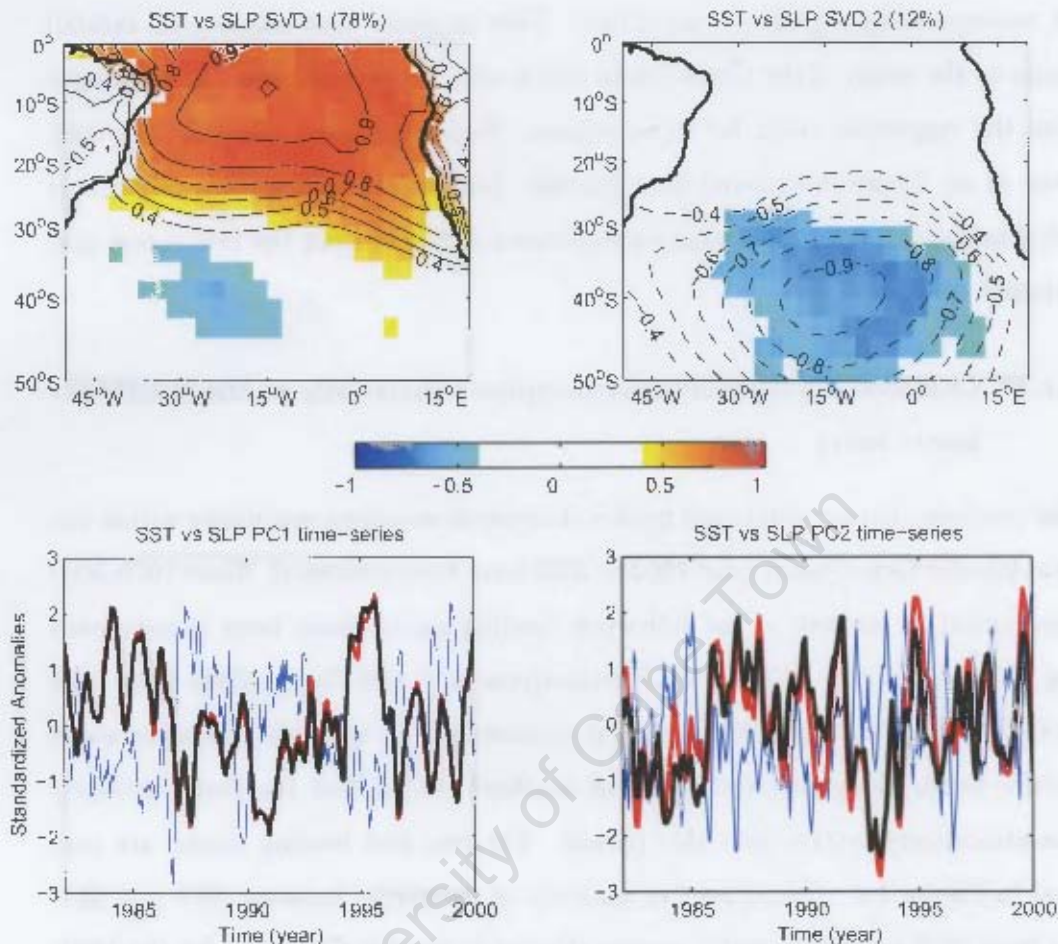


Figure 4.8: On top, SVD modes 1 (left) and 2 (right) between Hadley SSTs (color) and ECMWF SLPs (contours) for the 1979-2000 period and the expansion coefficient for SLPs (blue) and SSTs (thick red) in the bottom panel. The corresponding truncated SST expansion coefficients obtained from SVD analysis over the 1962-2002 period (see Chapter 3) are plotted in thick black.

north of its climatological position). The second mode characterizes the variability in the South Atlantic midlatitudes, linked with the weakening/strengthening of the South Atlantic high and its north/south displacement, due to modulation of surface pressures there (i.e. in the midlatitudes, to the south of its climatological position).

The potential relationships between these previous modes of variability within the ocean-atmosphere system and the January-February (JF) midlatitude South Atlantic mode in zonal moisture fluxes is assessed using lead/lag correlations up to a year (from JF-1 to JF+1) between JF MFPC1 and bimestrial SVD 1 and 2 SST/SLP expansion coefficients. The results presented in Figure 4.9 show significant relationships between positive/negative phases of the midlatitude South Atlantic mode in January-February (JF 0) and,

- (a) cooling/warming of SSTs in the tropical Atlantic (SST SVD 1) at the end of the preceding boreal summer (from JA-1 to ON-1, maximum in AS-1),
- (b) enhanced/reduced SLPs in the tropical Atlantic (SLP SVD 1) where the subtropical high is strengthened/weakened at the end of the preceding boreal summer and persisting until austral summer (from JA-1 to DJ 0, maximum in AS-1),
- (c) the meridional adjustments (northward/southward) and variations in intensity (weaker/stronger) of the South Atlantic anticyclone (SLP SVD 2) at the beginning of austral summer (from ON-1 to DJ 0, maximum in ND-1) just before the events,
- (d) the cooling/warming in the midlatitude South Atlantic (SST SVD 2) during the austral summer of the events (from DJ 0 to FM 0) and maximum in JF 0 when the events develop.

To validate the relationships found in December-January, corresponding composites are shown in Figure 4.10. Due to the period of analysis chosen (1979-2000), the 1978-79 December-January event could not be incorporated in the positive sample. The composites confirm that preceding positive/negative events of the midlatitude South Atlantic mode (in January-February), reduced/enhanced pressures are found within the South Atlantic anticyclone, which is weakened/strengthened and extending less/more south of its climatological position. A north-south dipole-like pattern

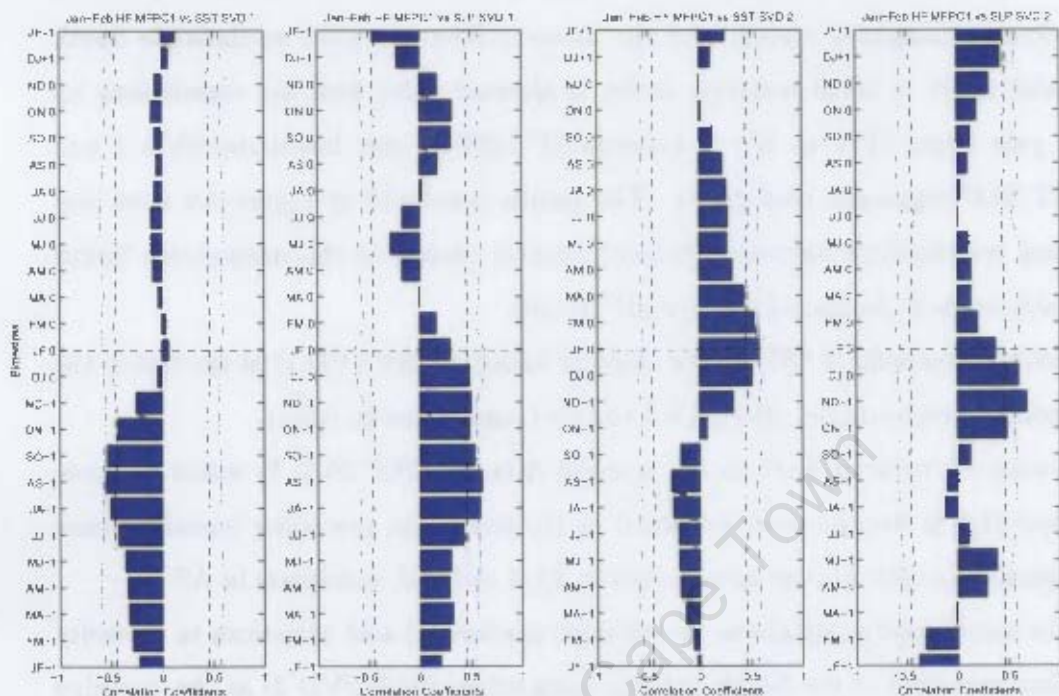


Figure 4.9: Lagged correlations between January-February midlatitude South Atlantic mode expansion coefficient for high frequency zonal moisture fluxes along the west southern African coast and bi-mestrial SVD 1/2 SST and SLP expansion coefficients (left/right) over the 1979-2000 period. 95% and 99% significance levels using Monte-Carlo simulations are indicated respectively in dashed black/dashdot blue lines. Year 0 corresponds to the calendar year of the event.

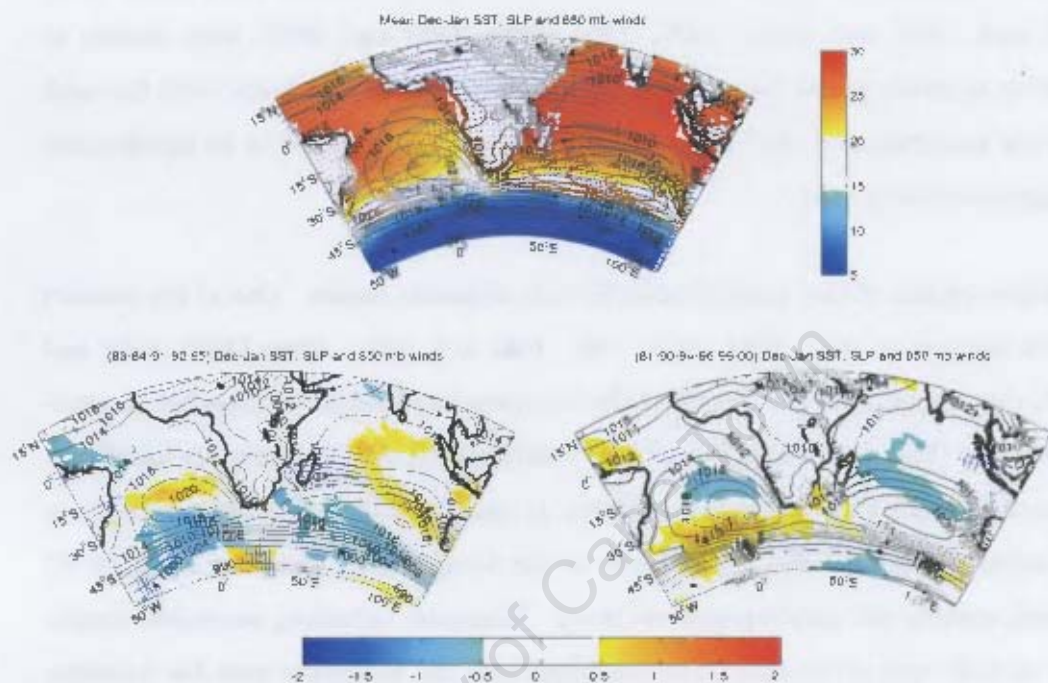


Figure 4.10: Mean 1979-2000 climatology (top), (1983, 1984, 1991, 1992 and 1995) composite (bottom left) and (1981, 1990, 1994, 1996, 1999 and 2000) composite (bottom right) for December-January (DJ) SLPs (contours), SSTs (color) and 850 mb winds (stream lines) significant at 90% level significance of Student t-test.

in SST anomalies further illustrates variability in the midlatitude South Atlantic just before the events.

4.1.1.2 Dynamics associated with the midlatitude South Atlantic mode

From the January-February midlatitude South Atlantic mode expansion coefficient time-series (Figure 4.5b), extreme years are extracted using a threshold at 0.6 of normalized anomalies deviation from the mean. (1979, 1983, 1984, 1991, 1992 and 1995) and (1981, 1990, 1994, 1996, 1999 and 2000) were chosen as positive/negative events respectively, for which the circulation linked with the midlatitude westerlies and the South Atlantic anticyclone appeared to be significantly strengthened/weakened.

Positive phase of the midlatitude South Atlantic mode Out of the positive events isolated in 1979, 1983, 1984, 1991, 1992 and 1995, three (1983, 1992 and 1995) correspond with low phases of the Southern Oscillation. Composites in vertical velocity (Figure 4.11 and Figure 4.12 middle panel) show statistically significant anomalous uplift widely over the continent at equatorial latitudes, while subsidence anomalies prevail broadly to the south of the Congo basin (south of about 20°S) at both surface and midtropospheric levels. Maximum subsiding anomalies significant at 95% level of Student t-test are found over the Botswana heat-low location, but also in the tropical Atlantic north of the Equator. Vertical velocity patterns in Figure 4.11 indicate also more marginal anomalous uplift and downlift respectively north and south of the Mozambique channel. It could suggest a potential eastward shift of the ascending Walker type circulation over southeast Africa and Indian Ocean regions for these years.

These results are confirmed by vertical maps of latitudinally averaged zonal moisture fluxes (Figure 4.12 middle panel). A deficit in zonal water vapour transport is noticeable to the south of the Congo basin (10°S-20°S) in the dry composites showing low-level easterly anomalies. In addition, significant easterly anomalies at midtropospheric levels (from 700 mb to 500 mb), maximum between 5°E and 30°E,

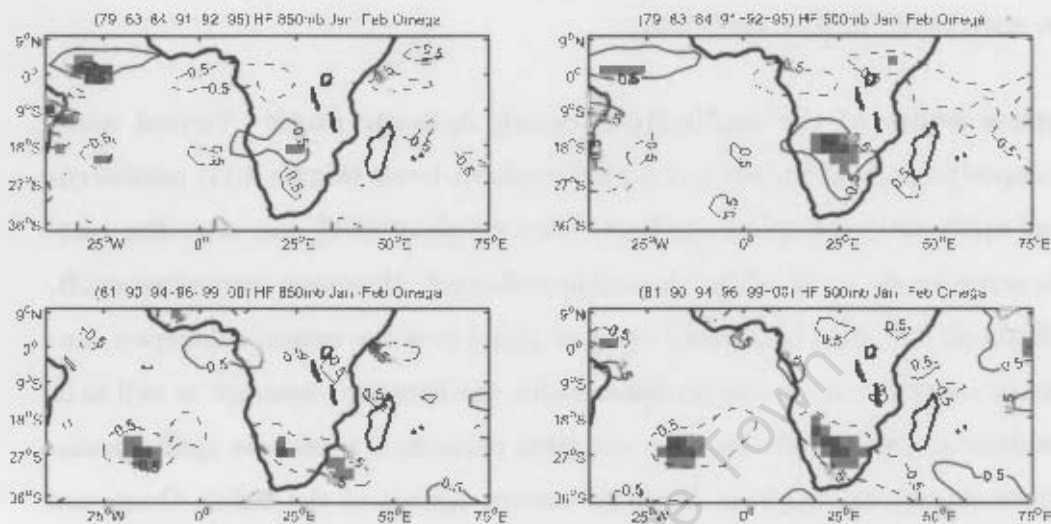


Figure 4.11: (1979, 1983, 1984, 1991, 1992, and 1995) composite (top) and (1981, 1990, 1994, 1996, 1999, and 2000) composite (bottom) for NCEP R2 HF vertical velocity at both 850mb (left) and 500mb levels (right) in January-February over the 1979-2000 period with dark/light shaded areas representing respectively 95% and 90% confidence level of Student t-test.

suggest a sustained zonal component of the southern extension of the African Easterly Jet (AEJ), consistent with the vertical loadings found for the midlatitude South Atlantic mode (Figure 4.5a). It could be instrumental in bringing more moisture to western tropical regions, typically along the Angolan coast, at the disadvantage of central regions south of the Congo basin. Further south (20°S - 35°S), significant westerly anomalies are found, as far as 30°E , across the southern tip of the subcontinent. This indicates a northward shift of the midlatitude westerly circulation during these years, which would act to drive moisture eastward out of continental areas, and further inhibit convection.

Negative phase of the midlatitude South Atlantic mode Vertical velocity composites at both surface and midtropospheric levels (Figure 4.11) exhibit enhanced uplift over subtropical southern Africa (south of 15°S) and extending eastwards across to the south of the Mozambique channel. Maximum anomalous uplift, significant at 95% level of Student t-test, is found over the central subtropics, emphasizing enhanced convection processes within the Botswana heat-low as well as in the midlatitude South Atlantic. The elongated patterns of maximum uplift anomalies from subtropical southern Africa to western regions of the Indian Ocean are typical of sustained convection activities within the South Indian Convergence Zone (SICZ) or tropical temperate troughs (TTTs) location, known as the most important rain producing system over the region (*Preston-Whyte and Tyson, 1988; D'Abreton, 1992; Cook et al., 2004*). Such results are confirmed by vertical maps (Figure 4.12 bottom panel) showing enhanced uplift locally and extending further south to the Botswana heat-low location. Simultaneously, subsiding anomalies are found over the Gabonese coast and across to the east of the subcontinent, which could further indicate alterations of the meridional circulation.

Figure 4.12 bottom panel shows from 10°S to 20°S , broad westerly anomalies within the air-column up to 400 mb level. At surface it would suggest that more moisture is advected over the subcontinent from the coast. Maximum westerly anomalies, significant at 95% significance level of Student t-test, at midtropospheric levels (from

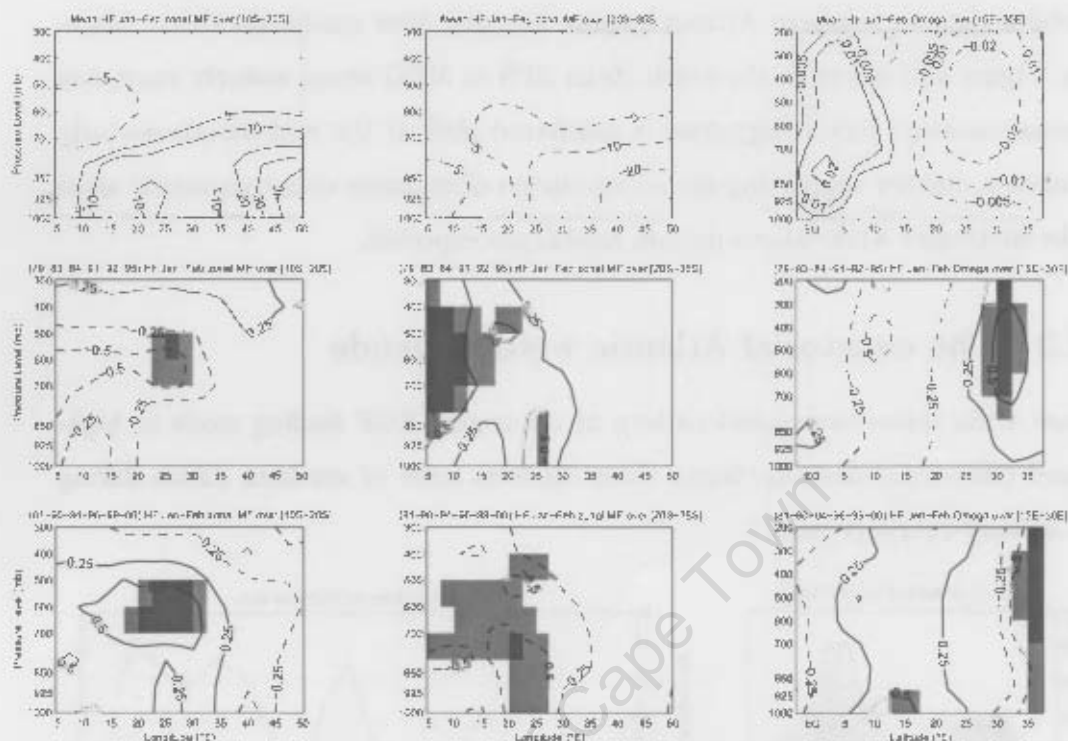


Figure 4.12: Mean (top), {1979, 1983, 1984, 1991, 1992 and 1995} composite (middle) and {1981, 1990, 1994, 1996, 1999 and 2000} composite (bottom) for HF zonal moisture fluxes averaged over [15°S-25°S] (left), and [25°S-35°S] (middle), with positive values corresponding to westerly fluxes while negative values refer to easterly orientated fluxes, together with NCEP R2 HF vertical velocity (right) averaged over [15°E-30°E] in January-February over the 1979-2000 period with dark/light shaded areas representing respectively 95% and 90% confidence level of Student t-test.

700 mb to 500 mb) show a weakened zonal component of the southern extension of the AEJ. It would act to accumulate the excess of moisture locally, while more water vapour penetrates deep inland at subtropical latitudes of the subcontinent. As a consequence, more humidity is available for convection within the Botswana heat-low location. Such a situation would also create a deficit in moisture over western coastal regions in southern African tropics, bringing drier conditions there. Moreover, Figure 4.12 shows to the south (from 20°S to 35°S) broad easterly anomalies in zonal moisture fluxes suggesting a southward shift of the midlatitude westerly circulation, further supporting the accumulation of moisture over continental areas in the subtropics where above-normal rainfall are expected.

4.1.2 The equatorial Atlantic westerly mode

Figure 4.13a shows the spatial pattern of the second EOF leading mode in high-filtered (HF) zonal moisture fluxes along the west coast of southern Africa during the January-February period.

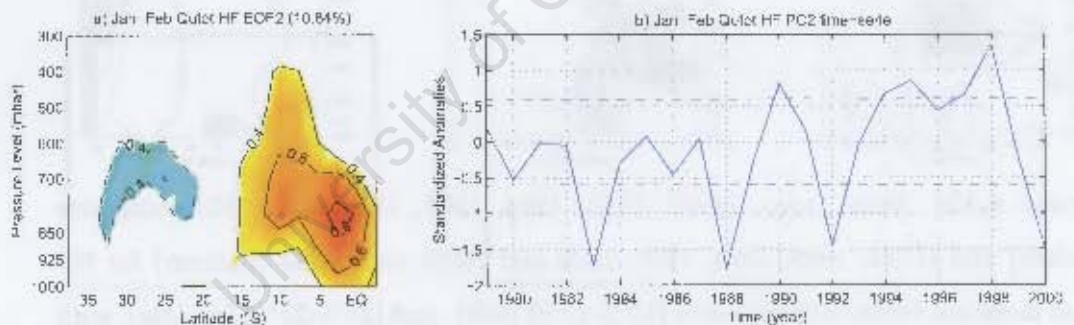


Figure 4.13: Spatial patterns (right) of the second EOF leading mode of variability in high frequency (HF) zonal moisture fluxes along the west coast of southern Africa together with its expansion coefficients (left) in January-February over the 1979-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

In January-February, when the westerly flux of moisture in the tropics is most pronounced, maximum positive loadings are found from the Equator to about 15°S , characteristic of this westerly moisture flux. Weaker negative loadings are located between 20°S and 30°S , at about 700 mb level. This negative pattern is absent from Figure 4.4 and this might be due to the variability within the southern extension of the AEJ located just to the north: most developed in summer it connects with the surface easterly flux to the south, where these negative loadings are found. Positive/negative events of this mode in January-February correspond with enhanced/reduced equatorial westerly flux from the tropical Atlantic.

Figure 4.14 shows heterogeneous correlations between the January-February equatorial Atlantic westerly mode expansion coefficient (JE MFPC2 in the following) and CRU rainfall.

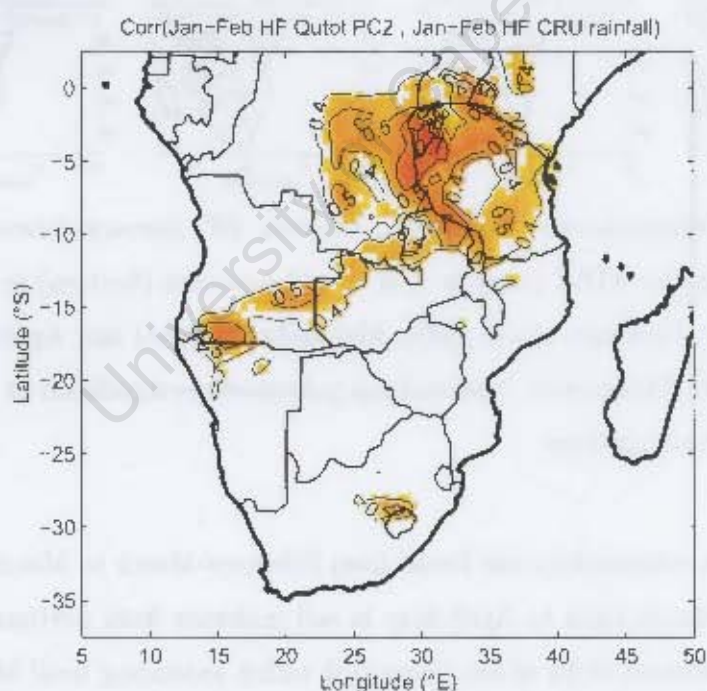


Figure 4.14: Heterogeneous correlations between HF zonal moisture fluxes PC2 and CRU rainfall in January-February over the 1979-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

High loadings for synchronous correlation are found all around the Congo basin, with maximums from east/southeast DRC and northwest Tanzania stretching to the south along the Rift valley and westwards to south Angola. Rainfall composites (not shown) for positive and negative years of the January-February moisture flux PC2 time-series (Figure 4.13b) confirm the symmetry of the relation with significant anomalies over these areas of maximum correlation scores.

To further emphasize the effects on the local hydrological cycle, heterogeneous correlations have been computed with both NDVI and soil moisture (Figure 4.15).

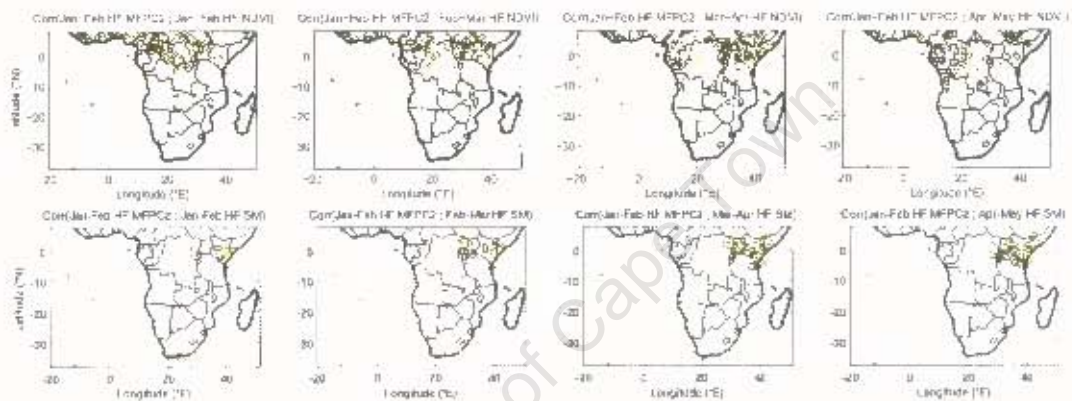


Figure 4.15: Heterogeneous correlations between HF January-February zonal moisture flux PC2 and NDVI (top) as well as soil moisture (bottom) in January-February (far left), February-March (left), March-April (right) and April-May (far right) over the 1982-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

Maximum positive relationships are found from February-March to March-April in NDVI and from March-April to April-May in soil moisture from northeast Congo regions along the eastern slope of the Great Rift valley, extending until May across Uganda and Kenya. Composites for positive/negative events (not shown) from the JF MFPC2 time-series also confirm the symmetry of the relation with corresponding significant anomalies. This suggests that impacts on rainfall regimes to the east of the Congo basin result with 2-3 months lag in changes within the vegetation cover

for regions to the northeast of the Great Rift valley, probably fuelled by runoff from higher altitude. Such results are again in agreement with *Farrar et al.* (1994) who found for semi-arid regions that NDVI is controlled by soil moisture in the concurrent month and that the correlation between NDVI and rainfall is highest when considering precipitated amounts from the concurrent and previous 2 months.

4.1.2.1 Connections with ocean-atmosphere variability in the South Atlantic basin

As for the first mode, links between the equatorial Atlantic westerly mode in zonal moisture fluxes and the ocean-atmosphere system within the South Atlantic basin are assessed using lead/lag correlations up to a year (from JF-1 to JF+1) between JF MFPC2 and bimestrial SVD 1 and 2 SST/SLP expansion coefficients. Figure 4.16 shows significant relationships between positive/negative phases of the equatorial Atlantic westerly mode in January-February (JF 0) and,

- (a) reduced/enhanced SLPs in the tropical Atlantic (SLP SVD 1) during early austral summer the year of the events (maximum in SO-1),
- (b) the meridional shift (southward/northward) of the South Atlantic anticyclone which is simultaneously strengthened/weakened (SLP SVD 2) the preceding boreal summer (not significant but maximum in MJ-1) and in austral summer (significant and maximum in ND-1), just before the development of the events in JF 0,
- (c) warming/cooling in the South Atlantic midlatitudes (SST SVD 2) the following boreal summer (not significant but maximum in JA 0/AS 0).

For events associated with this mode, significant links are only identified with the atmospheric component of the SVD analyses suggesting that there is no evidence for a strong linkage between this mode and ocean-atmosphere variability within the South Atlantic basin. November-December composites are presented in Figure 4.17 and confirm that preceding positive/negative events of this mode (in January-February), enhanced/reduced surface pressures are found within the South Atlantic anticyclone which is strengthened/weakened and extending more/less to the south of its climatological position.

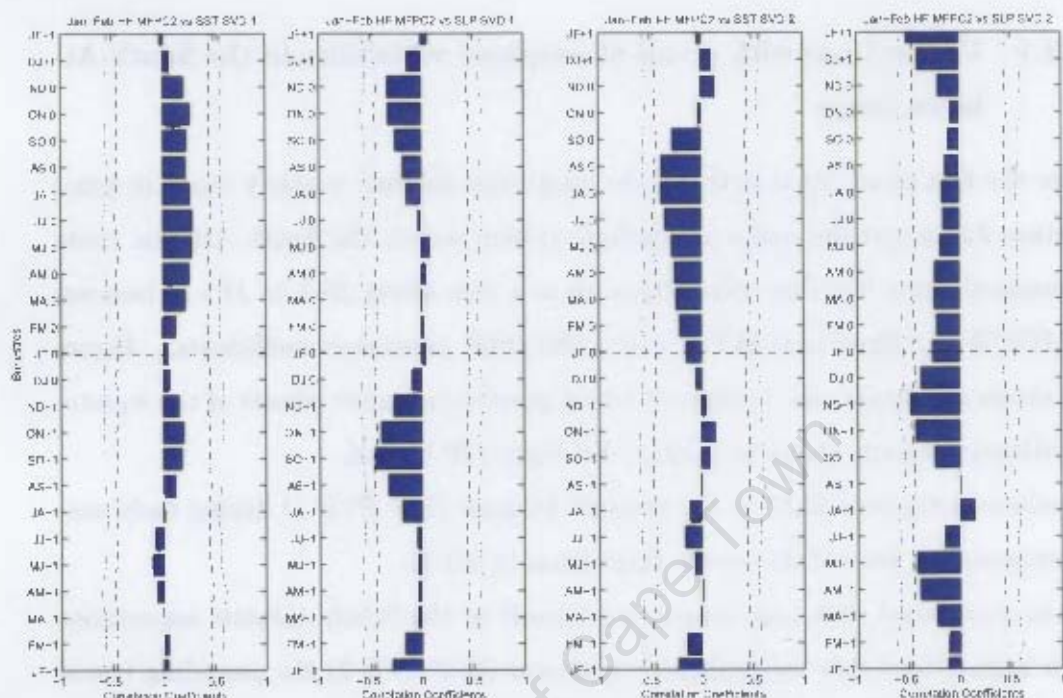


Figure 4.16: Lagged correlations between January-February equatorial Atlantic westerly mode expansion coefficient for high frequency zonal moisture fluxes along the west southern African coast and bi-mestrial SVD 1/2 SST and SLP expansion coefficients (left/right) over the 1979-2000 period. 95% and 99% significance levels using Monte-Carlo simulations are indicated respectively in dashed black/dash-dot blue lines. Year 0 corresponds to the calendar year of the event.

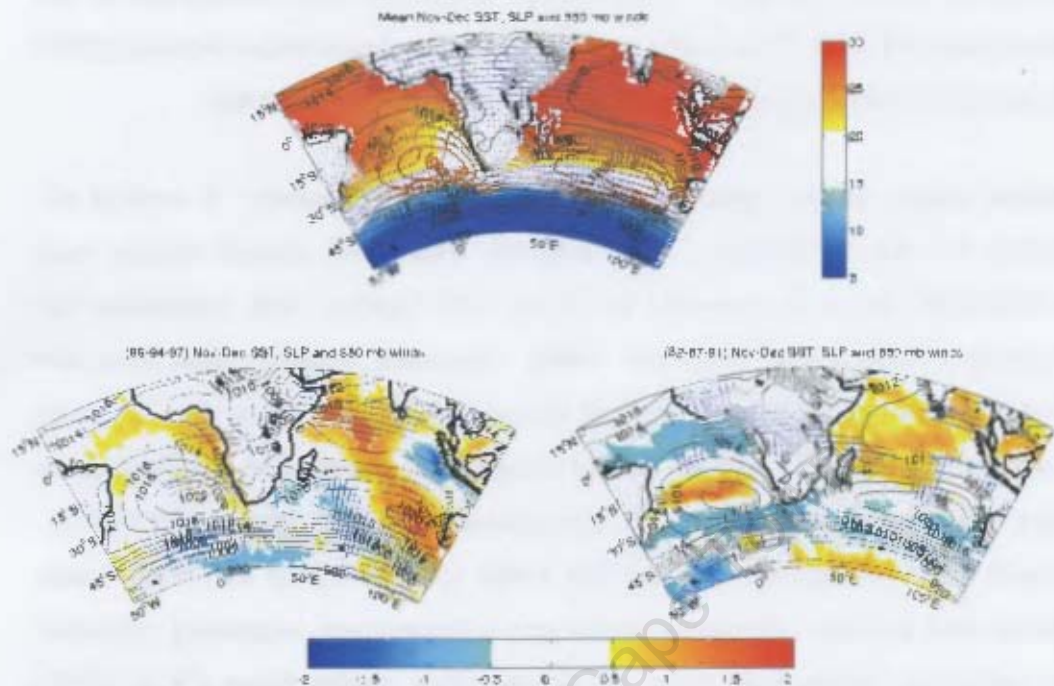


Figure 4.17: Mean 1979-2000 climatology (top), (1989, 1994 and 1997) composite (bottom left) and (1982, 1987 and 1991) composite (bottom right) for November-December (ND-1) SLPs (contours), SSTs (color) and 850 mb winds (stream lines) significant at 90% level significance of Student t-test.

4.1.2.2 Dynamics associated with the equatorial Atlantic westerly mode

From the January-February equatorial Atlantic westerly mode expansion coefficient time-series (Figure 4.13b), extreme years are extracted using a threshold at 0.6 of normalized anomalies deviation from the mean. (1990, 1995 and 1998) and (1983, 1988, and 1992) were chosen as positive/negative events respectively, for which the expression of the equatorial Atlantic westerly mode in zonal fluxes appeared to be significantly enhanced/reduced. Interestingly, all but one year corresponds to the mature phase of ENSO. This could explain the absence of correlation between ENSO and rainfall in central southern Africa (*Camberlin and Poccard, 2001*).

Positive phase of the equatorial Atlantic westerly mode A vertical climatology for January-February zonal moisture fluxes and vertical velocity over the 1979-2000 period is presented in Figure 4.19 together with composites for the positive events (1990, 1995, and 1998). Enhanced westerly fluxes from the Atlantic at both equatorial and tropical latitudes support the idea that humidity accumulates anomalously over the central Congo basin and western Angola at both surface and midtropospheric levels. In the equatorial latitudes (from 2.5°N to 5°S), increased zonal convergence over the Rift valley (30°E-35°E) up to 700 mb levels contrasts with sustained divergence in the upper troposphere, suggesting enhanced deep convection mechanisms there. In the southern tropics (from 5°S to 15°S), a similar situation is found but displaced westwards with convection mechanisms enhanced locally but more restricted to the lower troposphere.

In Figure 4.18, vertical velocity composites at both surface and midtropospheric levels exhibit enhanced uplift north of 15°S, statistically significant over the tropical Atlantic, the Gabonese coast and extending eastwards across the subcontinent. Maximum anomalous uplift is found over the central Congo basin, emphasizing sustained convection processes there. Simultaneously, pronounced subsiding anomalies extend from the south Namibian coast to southeastern regions of South Africa, further indicating alterations in the meridional circulation.

The January-February climatology presented in Figure 4.1 and 4.2 clearly shows

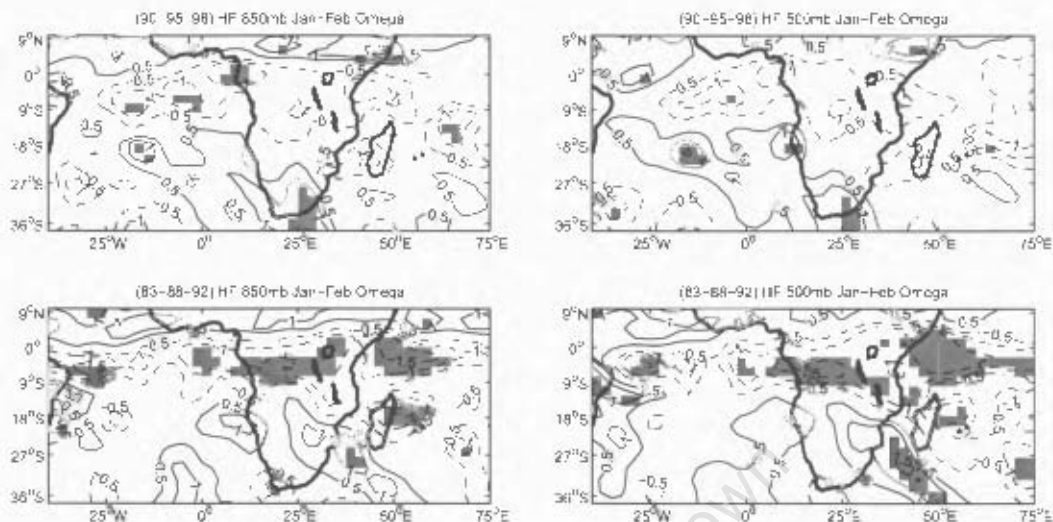


Figure 4.18: (1990, 1995, and 1998) composite (top) and (1983, 1988, and 1992) composite (bottom) for HF omega at both 850 mb (left) and 500 mb levels (right) in January-February over the 1979-2000 period with shaded areas representing 90% (light grey) and 95% (dark grey) confidence level of Student t-test. Negative (positive) values correspond to ascending (descending) motion.

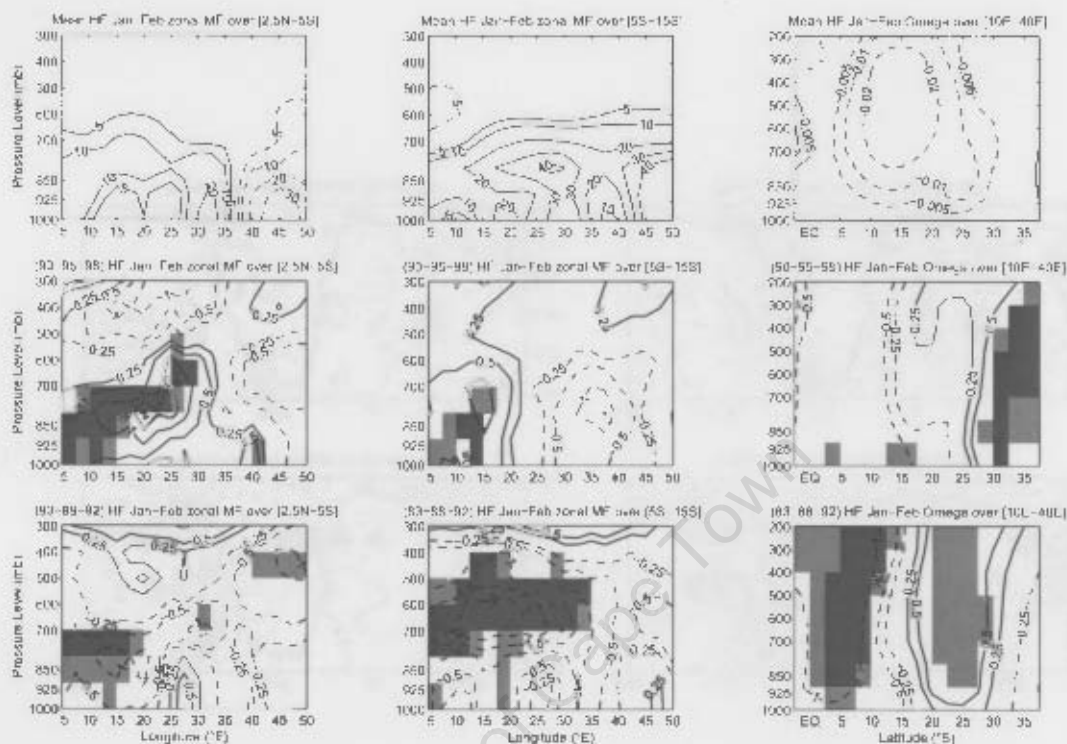


Figure 4.19: Mean (top), (1990, 1995 and 1998) composite (middle) and (1983, 1988, and 1992) composite (bottom) for January-February HF zonal moisture fluxes (in $g.kg^{-1}.m.s^{-1}$) averaged over $[2.5^{\circ}N-5^{\circ}S]$ (left), and $[5^{\circ}S-15^{\circ}S]$ (middle), with positive values corresponding to westerly fluxes while negative values refer to easterly orientated fluxes. On the right, mean (top) and similar composites for HF omega in January-February averaged over $[10^{\circ}E-40^{\circ}E]$. Negative (positive) values for omega correspond to ascending (descending) motion. Data are averaged over the 1979-2000 period and shaded areas represent 90% (light grey) and 95% (dark grey) confidence level of Student t-test.

the presence of the southern AEJ at mid-tropospheric levels (500-700 mb) feeding moisture over tropical areas of the subcontinent. In this context of changes in the meridional circulation, transport of available moisture could be increased at midtropospheric levels from the Congo basin southwards via the southern extension of the AEJ.

Negative phase of the equatorial Atlantic westerly mode The negative events isolated in 1983, 1988 and 1992 from the equatorial Atlantic westerly mode expansion coefficient in January-February (Figure 4.13b), also correspond with some of the low phases of the Southern Oscillation.

Vertical maps of latitudinally averaged zonal moisture fluxes (Figure 4.19 bottom panel) exhibit in the dry composites a pronounced deficit in zonal water vapour advection from the Atlantic all over the Congo basin. In addition, low-level easterly anomalies and upper westerly anomalies in both equatorial and tropical sections also show modifications in deep convection.

Composites in vertical velocity (Figure 4.18) witness of statistically significant anomalous uplift north of about 10°S and subsidence to the south of the Congo basin at both surface and midtropospheric levels. Vertical velocity patterns in Figure 4.18 and 4.19 also indicate anomalous uplift and downlift respectively to north and south of the Mozambique channel, significant at 95% level of Student t-test. Together these results further suggest a potential eastward shift of deep convection for these years as shown by elongated patterns to the southeast of the subcontinent characteristic of the SICZ. As a consequence, anomalous uplift over central southern Africa would further contribute to deprive these regions from moisture which is advected southwards toward southwest Indian Ocean regions, rather than over the subcontinent where drier conditions thus prevail.

4.1.3 Summary

An EOF analysis on high-filtered zonal moisture fluxes along the west coast of southern Africa helped identify two main modes of variability whose are listed as follow,

The midlatitude South Atlantic mode The first mode of zonal moisture fluxes (contributing to 25% of the variance explained) linked with the midlatitude westerly circulation and the South Atlantic anticyclone is found to directly impact, during mid-summer months (January-February), rainfall amounts over subtropical areas of southern Africa (south of 15°S), with also opposite influences over western coastal regions at tropical latitudes (roughly from 5°S to 15°S). Beside the fact that they potentially modulate moisture input over the subcontinent, enhanced/reduced westerly moisture fluxes south of 32°S are resulting in below/above normal rainfall locally, which suggest that they could act to inhibit/support convection over southern Africa subtropics. The impacts on the hydrological cycle extend well over these regions of the subcontinent, with substantial signals developing until February-March in both vegetation cover and soil moisture.

This mode shows variability at both intra-annual and interannual time-scales and high frequency background seasonal signal. Significant relationships are found with ocean-atmosphere variability within the South Atlantic Ocean basin: particularly in December-January preceding positive/negative events of this mode (in January-February), the South Atlantic anticyclone is found to be weakened/strengthened and extending less/more south of its climatological position, while north-south dipole-like SST anomalies further illustrate variations in the midlatitude South Atlantic. Such findings are consistent with earlier studies which have evidenced that changes in the midlatitude South Atlantic SSTs would imply meridional shifts of the midlatitude westerly circulation associated with rainfall anomalies over southern Africa (*Mason, 1995*).

During the positive phase of the midlatitude South Atlantic mode (1979, 1983, 1984, 1991, 1992 and 1995) when the South Atlantic anticyclone is weakened and

extends less to the south of its climatological position, a reduction in available moisture occurs south of the Congo basin together with weakened convection processes in subtropical latitudes of the subcontinent. These changes are associated with potential modifications of the Walker type circulation, suggesting an eastward shift of its ascending branch over the western Indian Ocean and southeastern African regions. In addition, the southern extension of the AEJ, which zonal component is strengthened during these events, could play a role in reducing/enhancing moisture availability respectively over central regions south of 15°S and along the west coast between 5°S and 15°S . The midlatitude westerly circulation shifted northwards, supports anomalous eastward moisture transport from continental areas to southwest Indian Ocean regions. Such changes would explain the overall reduction in moisture availability and convection over subtropical southern Africa leading to below-normal rainfall locally while western coastal regions of tropical southern Africa are subject to wetter conditions. In the events sample, three out of six events are ENSO years and previous studies show similar modifications in the three dimensional circulation during low phases of the Southern Oscillation linked with below-normal rainfall over the subcontinent (*Preston-Whyte and Tyson, 1988; Reason, 1998; Reason et al., 2000; Richard et al., 2000*). It further confirms that drier conditions during El Niño events, correspond with a weakened South Atlantic anticyclone and a northward shift of storm tracks (*Lindesay et al., 1986*).

On the other hand, 1981, 1990, 1994, 1996, 1999 and 2000 were years for which the South Atlantic anticyclone was strengthened and extended further south of its climatological position. Our results show that more humidity is accumulated and advected inland, to the detriment of western coastal regions in the tropics, by a strengthened westerly moisture advection at surface, while the zonal component of the southern extension of the AEJ is weakened at midtropospheric levels. As a consequence, more moisture from the South Atlantic is available over the Botswana heat-low location in summer. The concomitant southward shift of the midlatitude westerly circulation further supports the idea that moisture accumulates over continental areas of the subtropics. Convective processes favoured within the SICZ/TTTs

location result in above-normal rainfall over subtropical regions of southern Africa, while drier conditions prevail over western coastal areas of tropical southern Africa.

The equatorial Atlantic westerly mode The equatorial Atlantic westerly mode (contributing to 11% of the variance explained) is correlated in January-February with rainfall amounts over the uplands all around the Congo basin, i.e. from the east DRC and northwest Tanzania down to south Angola. This agrees with the findings of *Mapande and Reason* (2005) suggesting links between enhanced westerly moisture input from the Congo basin and above normal rainfall to the east. In addition, these results further emphasize a potential decoupling between northern and southern hydrological regimes at basin scale, confirming the study from *Laraque et al.* (2001). The impacts on the hydrological cycle further extend to tropical regions to the east of the Great Rift valley, i.e. across Uganda through Kenya, with substantial signals developing until March-April and April-May respectively in vegetation cover and soil moisture.

This mode exhibit mainly intra-annual and interannual variability. Interestingly, prior to January-February, substantial relationships are found between the equatorial Atlantic westerly mode in zonal moisture flux along the west coast of southern Africa and the South Atlantic ocean-atmosphere system through atmospheric variability exclusively. This suggests no evidence of particular linkage between this mode of variability in zonal moisture fluxes and ocean-atmospheric variability in the South Atlantic basin. In November-December preceding positive/negative events of this mode (in January-February), these results suggest that the South Atlantic anticyclone is strengthened/weakened and extending more/less to the south of its climatological position.

For years when the equatorial westerly flux of moisture was particularly strengthened (1990, 1995 and 1998), the results show an anomalous advection of moisture over the Congo basin creating a situation where deep convection processes are enhanced. Excess water vapour is channelled from these regions around the Congo basin to the east and southeast at surface, while the southern extension of the AEJ

could be the mechanism for transporting more moisture southwards at mid-tropospheric levels. Anomalies in vertical velocities further support an enhanced water vapour transfer to the south, resulting in more moisture available for local convection and enhanced rainfall over the uplands surrounding the Congo basin. During the years when the equatorial westerly moisture transport was reduced (1983, 1988 and 1992), the dynamical processes involved seem less coherent than for the first mode discussed earlier for instance. Nevertheless, a deficit in moisture advection over the Congo basin did occur. These changes are accompanied by modifications in deep convection processes, suggesting a shift eastwards to the southwest Indian Ocean and southeastern parts of southern Africa. This reduction in convection over the subcontinent led to below average rainfall. This finding agrees with *D'Abreton and Tyson* (1995) who showed that during dry summers, transport from the tropical Atlantic is replaced by an enhanced moisture source over the Indian Ocean. Moreover, in the events sample all are corresponding with ENSO years and thus these findings also agree with previous studies stating similar modifications in the three dimensional circulation during low phases of the Southern Oscillation linked with below-normal rainfall over the subcontinent (*Preston-Whyte and Tyson*, 1988; *Reason*, 1998; *Reason et al.*, 2000; *Richard et al.*, 2000).

4.2 Role of the South Indian Ocean

As for the previous section, the humidity input from the Indian Ocean towards the subcontinent has been considered as mainly zonal, and moisture fluxes were averaged zonally for grid points along the east coast of southern Africa (ranging from 2.5°N to 37.5°S and from 20°E to 120°E). The variability within zonal water vapour transport is similarly examined using EOFs for this wall-like domain along the east coast this time. As higher modes seem degenerate, only the first three leading modes of variability have been retained and they are described in more detail in this section.

Connections between South Indian Ocean SSTs and southern African climate have been evidenced in earlier studies (*Mason, 1990, 1995; Reason and Godfred-Spenning, 1998; Reason and Mulenga, 1999; Landmann and Mason, 1999; Saji et al., 1999; Reason, 2001a; Behera and Yamagata, 2001; Rouault et al., 2002; Hansingo and Reason, 2006; Washington and Preston, 2006*). The early summer season has not received so much attention despite its importance for southern African rainfall. Using lead/lag correlations between the expansion coefficients of each mode and rainfall data, it is the period for which the impacts of zonal moisture fluxes variability along the east coast of southern Africa on local rainfall regimes are found to be the most pronounced, which was confirmed by the temporal coherence found with the corresponding signal in land hydrology data. In this respect, November-December vector moisture fluxes maps with contours of mean humidity convergence at 850 mb, 700 mb and 500 mb are presented in Figure 4.20. The mean November-December climatological structure for zonal moisture fluxes along the east coast of southern Africa is shown in Figure 4.21. As for the west coast, the transient term represents less than 10% of the total zonal moisture flux, and the stationary component will be well suited to represent changes in moisture fluxes due to large scale circulation. During early summer, an easterly moisture flux linked to the northern Indian trades is found between the Equator and 15°S in latitudes and below 500 mb level in altitudes. To the south (between 18°S and

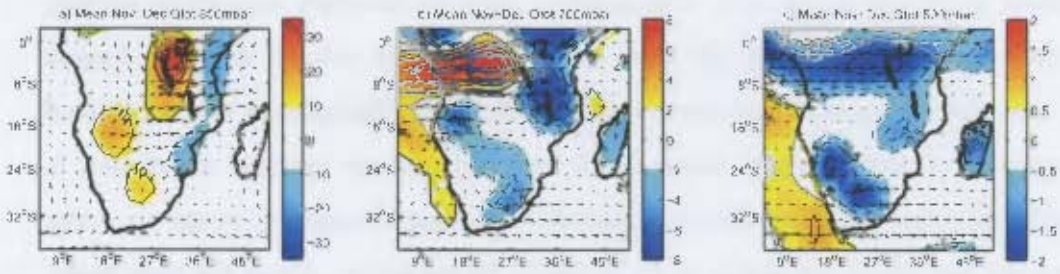


Figure 4.20: Mean November-December surface moisture fluxes (streamlines in $g.kg^{-1}.m.s^{-1}$ with arrows scaled at 1 unit/degree of latitude) over the 1979-2000 period, together with contours of moisture convergence (in $g.kg^{-1}.s^{-1}$) at 850 mb (a), 700 mb (b) and 500 mb (c). Positive values contour areas of moisture convergence while negative values refer to moisture divergence at given levels.

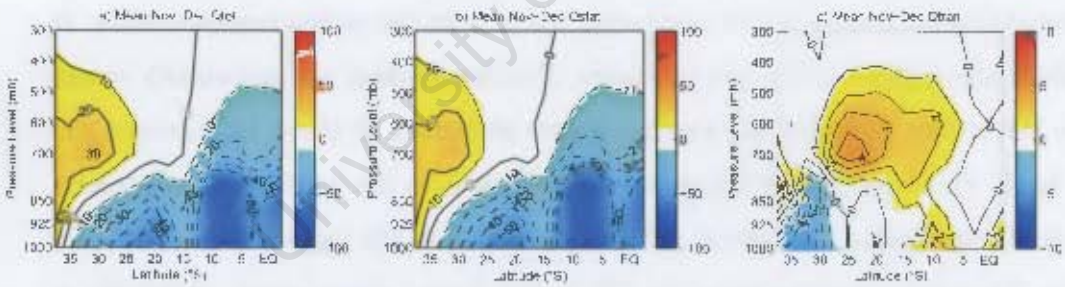


Figure 4.21: Mean vertical structure (a) of zonal moisture fluxes along the east Southern African coast (in $g.kg^{-1}.m.s^{-1}$) for November-December over the 1979-2000 period, together with its stationary (b) and transient (c) components. Positive values correspond to westerly fluxes while negative values refer to easterly fluxes.

35°S) another easterly flux is driven by the southern Indian trades at surface (below 850 mb). South of 35°S, a midtropospheric westerly moisture flux has its main core centered at 700 mb in altitude. Compared to January-February, convective activity seems better developed over the east Congo basin while reduced over the Angola low and the subtropical heat-low locations. North of Madagascar the meridional component of water vapour transport is reduced and no cyclonic circulation is established over the Mozambique channel. Stronger divergence seems to occur along the east coast, extending as far as south Mozambique. Along the west coast, the westerly flux in the tropics is reduced in intensity, thus advecting less moisture over the subcontinent interior.

As previously, the data have been filtered at 8 years and EOF techniques were applied to the high-filtered (HF) zonal moisture fluxes along the east coast of southern Africa using the covariance matrix. The first three leading EOF modes shown for all seasons in Figure 4.22, have been retained in accordance to the scree test (not shown) and this truncated basis contributes almost half of the total variance explained.

The first mode is typical of the variations within the northern Indian trades accompanied by changes in the circulation linked with the midlatitude westerlies. It explains about 20.2% of the total variance. Positive loadings are particularly strong from the surface up to 500 mb showing a core maximum at about 15°S around 850 mb levels while loadings of opposite sign are found in the subtropics particularly below 700 mb levels. This mode will be referred to as the tropical Indian easterly mode. Time-series analysis using the CWT (not shown) shows mostly variability at interannual time-scales (at 1, 3 and 5 years) with high frequency background signal.

The second mode characterizes the variability in the circulation linked with the midlatitude wind regimes and the South Indian anticyclone. It represents 13.7% of the total variance explained. It has simultaneous positive/negative loadings mainly below 700 mb levels, respectively south of 30°S and between 15°S and 10°S. In the following, this mode will be referred to as the midlatitude South Indian mode. Some

similitudes are found with the patterns associated with the first mode described above. Still, the magnitude of the loadings is a difference that has to be taken into account: the use of rotated EOFs could help to draw a more distinct picture but this is beyond the scope of this work. This mode exhibits most variability (not shown) at interannual time-scales (at 1, 2 and 5-6 years) with a marked seasonal signal.

The third mode reflects modulations in moisture fluxes at equatorial latitudes accompanied by opposite changes within the northern Indian trades latitudes to the south, which could illustrate the variability of monsoon fluxes in summer and winter. It represents almost 10% of the total variance explained. It has simultaneous positive/negative loadings mainly below 500 mb levels, respectively from the Equator to about 15°S and to the south between 15°S and 20°S. This mode will be referred to as the equatorial Indian easterly mode. Time-series analysis using the CWT (not shown) exhibits mostly variability at interannual time-scales (at 1, 5 and 8 years).

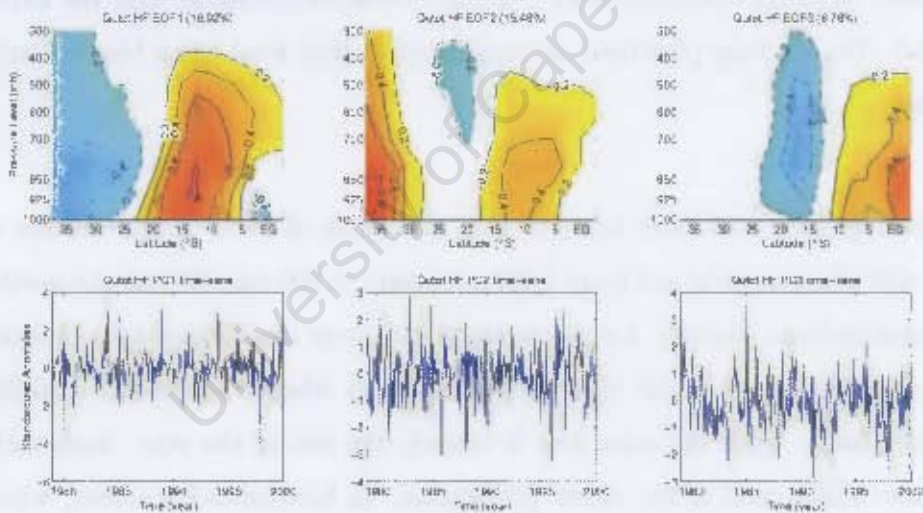


Figure 4.22: Spatial patterns (top) of the first two EOF leading modes of variability in high frequency (HF) zonal moisture fluxes along the east coast of southern Africa over the 1979-2000 period, together with their respective expansion coefficients (bottom) for all seasons.

In the following, the early summer season will be the period of interest and the corresponding projections of these modes during this period will be discussed.

4.2.1 The tropical Indian easterly mode

Figure 4.23a shows the spatial pattern of the first EOF leading mode in high-filtered (HF) zonal moisture fluxes along the east coast of southern Africa in November-December.

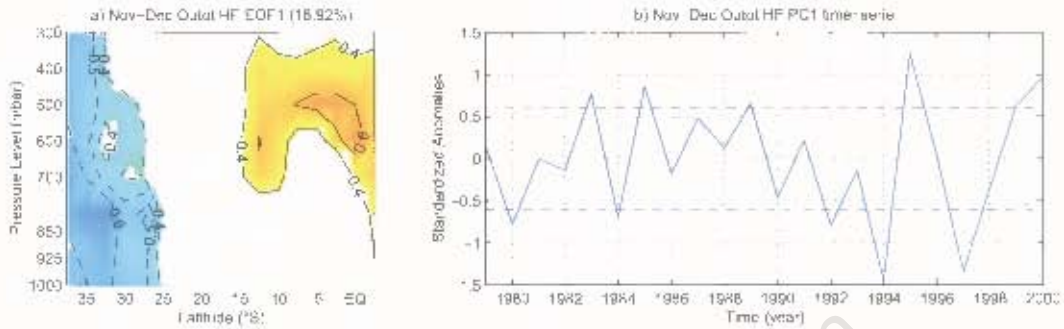


Figure 4.23: Spatial patterns (right) of the first EOF leading mode of variability in high frequency (HF) zonal moisture fluxes along the east coast of southern Africa together with its expansion coefficients (left) in November-December over the 1979-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

Marked loadings are found both north of 15°S and south of 25°S , but the former is shifted in altitude compared to Figure 4.22, maximum at 500 mb. Within the northern trades latitudes in summer, a cyclonic circulation over the Mozambique channel brings a westerly flux at about 15°S (see Chapter 2) where Figure 4.22 exhibits maximum loadings, while the main flow is easterly the rest of the year. Such variability is not represented in the above projections. In November-December, when the northeast monsoon prevails, this mode describes reduced/enhanced midtropospheric easterly water vapour transport at tropical latitudes (corresponding with weakened/strengthened northeast monsoon regimes) together with a weakening/strengthening of the midlatitude westerly circulation.

Heterogeneous correlations between the November-December tropical Indian easterly mode expansion coefficient (ND MFPC1 in the following) and CRU rainfall are presented in Figure 4.24.

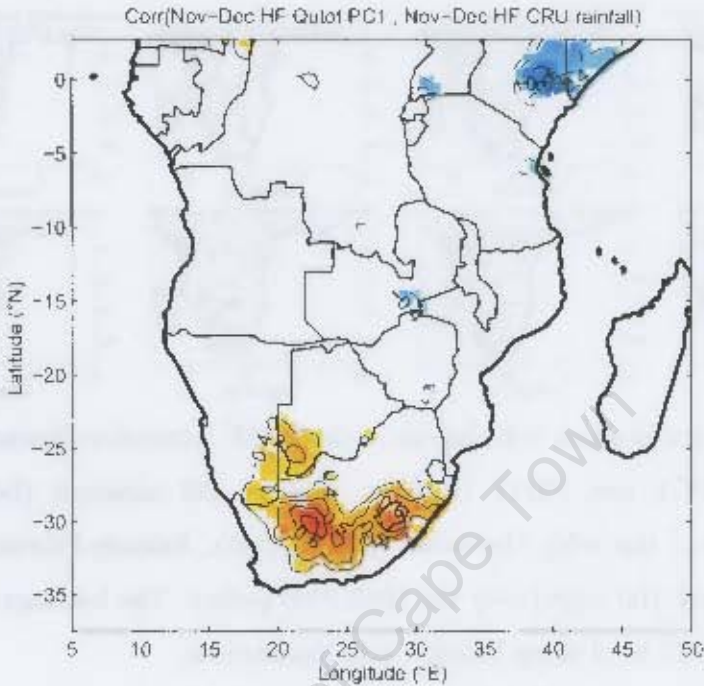


Figure 4.24: Heterogeneous correlations between HF zonal moisture fluxes PC1 and CRU rainfall in November-December over the 1979-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

High positive loadings for synchronous correlation are found over subtropical regions of the subcontinent, with maximums extending from southeast Namibia across to southeast regions of South Africa. Patterns of opposite signs are found simultaneously over northeast southern Africa, mainly over eastern Kenya/southern Somalia. Rainfall composites (not shown) for positive and negative years of the ND MFPC1 time-series (Figure 4.23b) confirm the symmetry of the relation with significant anomalies over these areas of maximum correlation scores.

To further emphasize the effects on the local hydrological cycle, heterogeneous correlations have been computed with both NDVI and soil moisture (Figure 4.25).

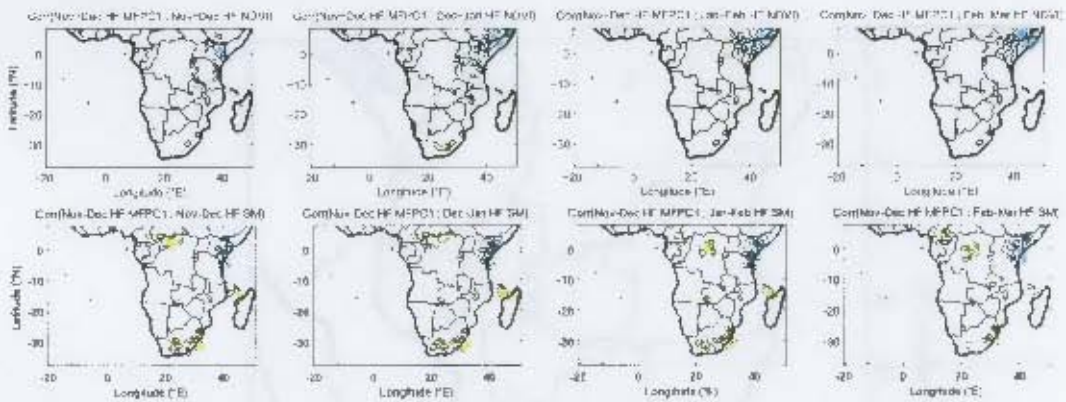


Figure 4.25: Heterogeneous correlations between IIF November-December zonal moisture fluxes PC1 and NDVI (top) as well as soil moisture (bottom) in November-December (far left), December-January (left), January-February (right) and February-March (far right) over the 1982-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

Globally, highest correlation scores in land hydrology data match the patterns found in rainfall. Regarding subtropical regions of southern Africa, maximum positive relationships are found in December-January for NDVI and from November-December to January-February in soil moisture, mainly over central and southeastern regions of South Africa. Composites for positive/negative events from the ND MFPC1 time-series (not shown) confirm the symmetry of the relation with corresponding significant anomalies. The limited impacts in time on vegetation cover seems to reflect the different mechanisms entering into play during early and late summer season. On the other hand, the temporal inertia of the signal found in soil moisture typically shows the importance of early summer rainfall on land conditions for the rest of the season over semi-arid regions of South Africa (i.e. the Karoo).

For northeastern regions of southern Africa, maximum negative relationships are found from December-January to February-March in both NDVI and soil mois-

ture with highest loadings over Kenya and south Somalia. Composites for positive/negative events (not shown) also confirm the symmetry of the relation in both fields. This suggests that impacts on rainfall regimes to the east of the Great Rift escarpment have a 2-3 months lag in changes within land hydrology, most pronounced over Kenya.

As evidenced previously for semi-arid regions to the northeast and southeast of southern Africa, NDVI appears to be controlled by soil moisture in the concurrent month and the correlation between NDVI and rainfall is found to be highest at 2 months lag, still in agreement with *Farrar et al.* (1994).

4.2.1.1 Connections with ocean-atmosphere variability in the Indian Ocean basin

Dominant modes of variability have been emphasized in the South Indian Ocean region (see Chapter 3). In order to investigate potential relationships between zonal moisture fluxes variability along the east coast of southern Africa and the ocean-atmosphere system in the Indian Ocean basin for the 1979-2000 period, leading modes of variability between SSTs and SLPs have been re-computed using SVD techniques on ECMWF SLP re-analyses data and Hadley SSTs for this period over the South Indian Ocean region. The two first leading modes are presented in Figure 4.26. The respective patterns of variability between SST and SLP fields from 1979 to 2000 resemble greatly the ones isolated for the 1962-2002 period. The SST expansion coefficients issued from both 20 years and 50 years SVD analyses appear to be significantly well correlated with scores of about 0.98 and 0.78 for the first and second mode respectively. Such findings give confidence in the fact that the modes found over the 1979-2000 period are consistent with the ones obtained previously for 1962-2002. Typically the first mode shows the recent warming trend over the broad South Indian Ocean regions, with corresponding enhanced/reduced surface pressures particularly to the northeastern parts of the basin. Substantial links with ENSO have been evidenced for this mode (see Chapter 3). The second mode represents the variability associated with southwest/northeast SST/SLP gradients

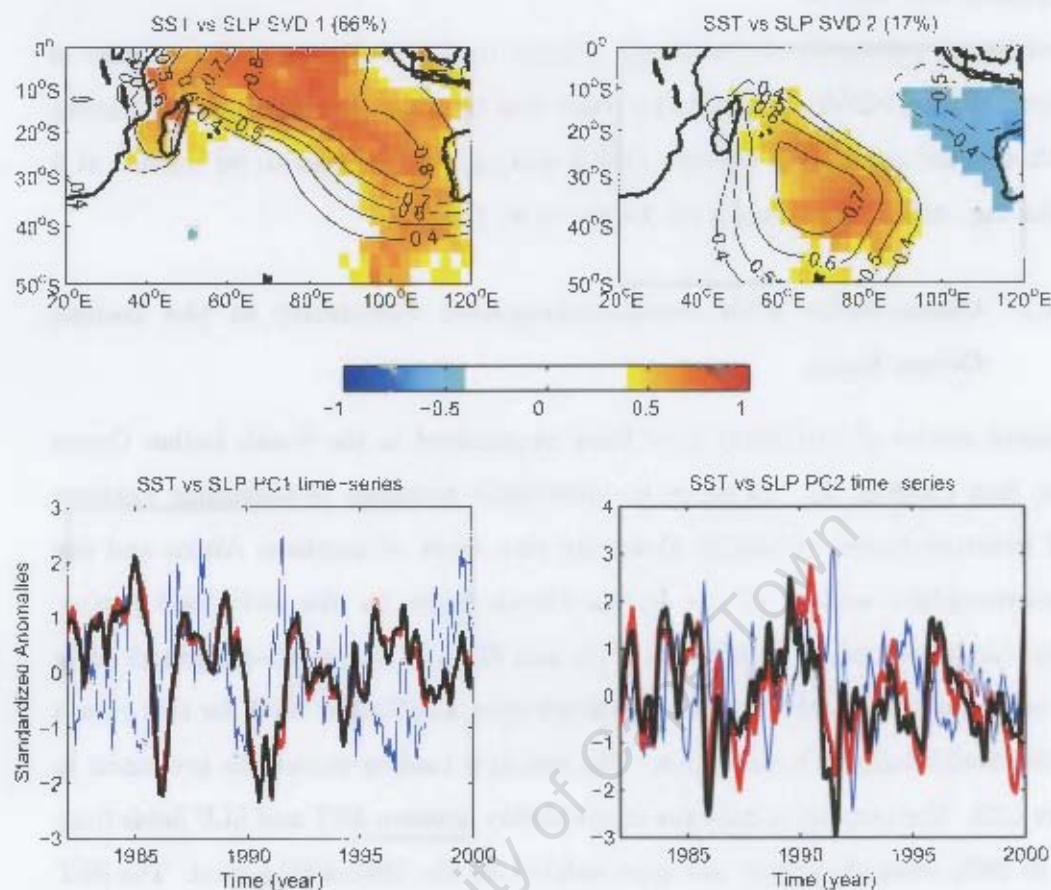


Figure 4.26: On top, SVD modes 1 (left) and 2 (right) between Hadley SSTs (color) and ECMWF SLPs (contours) for the 1979-2000 period and the corresponding expansion coefficient for SLPs (blue) and SSTs (thick red) in the bottom panel. The corresponding truncated SST expansion coefficients obtained from the SVD analysis over the 1962-2002 period (see Chapter 3) is also plotted in thick black.

in the South Indian Ocean regions. It resembles the subtropical Indian dipole (SID) of *Behara and Yamagata (2001)*.

Potential relationships between these previous modes of variability within the ocean-atmosphere system and the November-December (ND) tropical Indian easterly mode in moisture fluxes are assessed using lead/lag correlations up to a year (from ND-1 to ND + 1) between ND MFPC1 and bimestrial SVD 1 and 2 SST/SLP expansion coefficients. The results presented in Figure 4.27 show significant relationships between positive/negative phases of the tropical Indian easterly mode in November-December (ND 0) and,

(a) the establishment of a negative/positive SID-type gradient in surface pressures (SLP SVD 2) across Indian Ocean regions (opposite anomalies in SLPs to the northeast/southwest of the basin) during the preceding austral summer (maximum in DJF -1/JF 0). An opposite situation (positive/negative SID-type SLP gradient) is found in early austral summer the year of the events (from SO 0 to ND 0, maximum in ON 0), just before their development in ND 0.

(b) warming/cooling of SSTs to the northeast of the basin (SST SVD 1) during the preceding boreal summer prior to the events (not significant but maximum in JA 0), while it is the opposite after the events (not significant but maximum in MJ + 1),

(c) reduced/enhanced surface pressures to the northeast of the basin (SLP SVD 1) at the end of the preceding austral summer (maximum in MA 0), at the end of the boreal summer preceding the events (maximum in AS 0/SO 0) and during the austral summer of the event (not significant but maximum in JF + 1).

The main links are identified exclusively with atmospheric variability in the South Indian Ocean, suggesting no evidence for strong relationships between this mode and ocean-atmosphere variability in the Indian Ocean basin. October-November composites are presented in Figure 4.28 and confirm that, preceding positive/negative events of the tropical Indian mode in November-December, enhanced pressures are found to the southwest/northeast of the South Indian anticyclone centre where it is strengthened.

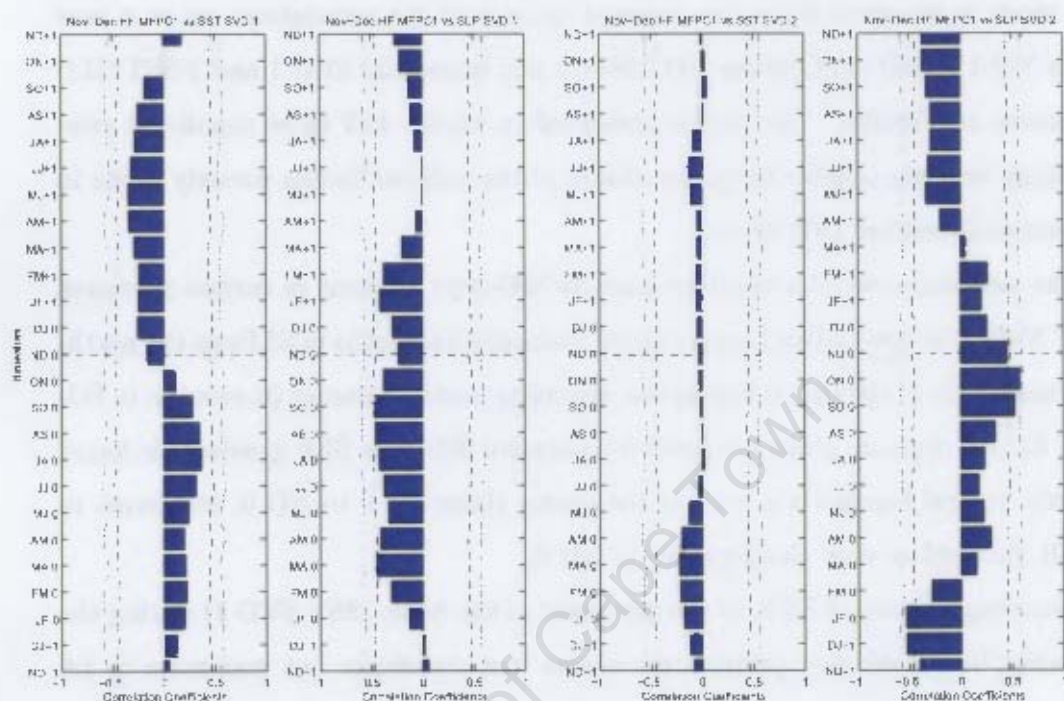


Figure 4.27: Lagged correlations between November-December tropical Indian mode expansion coefficient (ND MFPC1) for high frequency zonal moisture fluxes along the east southern African coast and bi-mestrial SVD 1/2 SST and SLP expansion coefficients (left/right) over the 1979-2000 period. 95% and 99% significance levels using Monte-Carlo simulations are indicated respectively in dashed black/dashdot blue lines. Year 0 corresponds with the calendar year of the event.

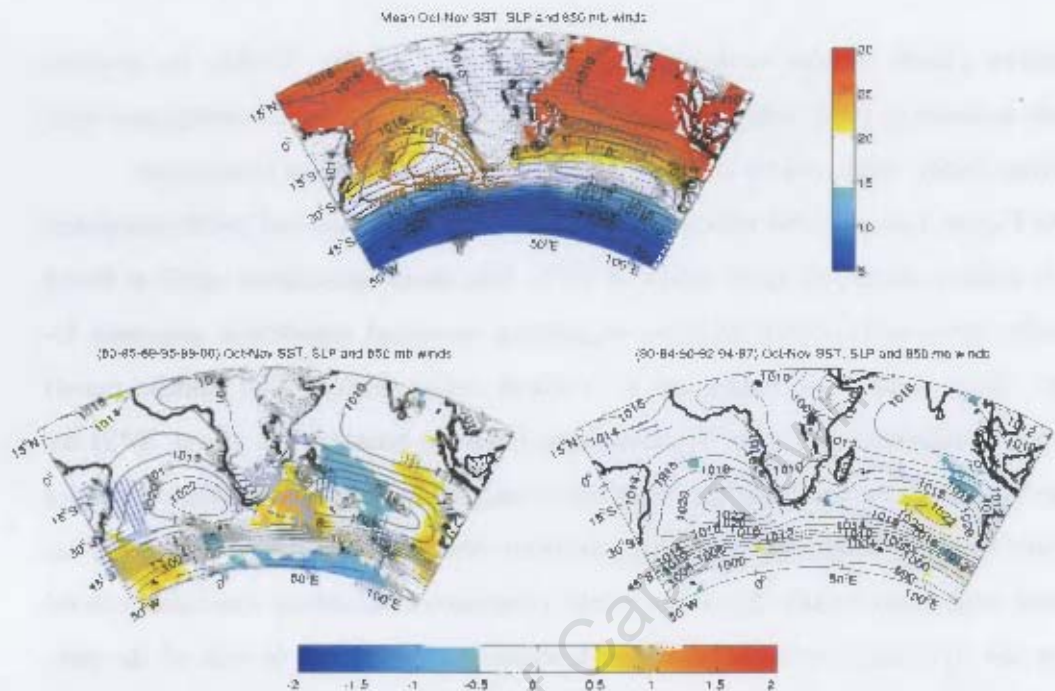


Figure 4.28: Mean 1979-2000 climatology (top), (1983, 1985, 1989, 1995, 1999 and 2000) composite (bottom left) and (1980, 1984, 1990, 1992, 1994 and 1997) composite (bottom right) for October-November (ON 0) SLPs (contours), SSTs (color) and 850 mb winds (stream lines) significant at 90% level significance of Student t-test.

4.2.1.2 Dynamics associated with the tropical Indian easterly mode

From the November-December tropical Indian easterly mode expansion coefficient time-series (Figure 4.23b) are extracted extreme years using a threshold at 0.6 of normalized anomalies deviation from the mean. (1983, 1985, 1989, 1995, 1999 and 2000) and (1980, 1984, 1990, 1992, 1994 and 1997) were chosen as positive/negative events respectively. Three out of six negative events correspond to El Niño years.

Positive phase of the tropical Indian easterly mode Within the positive events isolated in 1983, 1985, 1989, 1995, 1999 and 2000, one year corresponds with La Niña (2000) while others are independent from the Southern Oscillation.

In Figure 4.29, vertical velocity composites at both surface and midtropospheric levels exhibit enhanced uplift south of 10°S . Maximum anomalous uplift is found broadly across subtropical latitudes suggesting sustained convection processes locally. Such results are confirmed by vertical maps (Figure 4.30 middle panel) showing enhanced uplift over the Botswana heat-low location (at about 30°S) for these years. It is worth noting that the elongated patterns over the subtropics (Figure 4.29) are found over the mean location where TTTs develop indicating enhanced convection locally. Simultaneously, pronounced subsiding anomalies extend along the equatorial latitudes from the Gabonese coast across to east of the sub-continent over central Indian Ocean regions where convective activity is weakened, further indicating alterations in the meridional circulation. The patterns for omega in Figure 4.30 middle panel show enhanced subsidence at tropical latitudes, and further suggest reduced convection to the east of the Congo basin.

A vertical climatology for November-December zonal moisture fluxes over the 1979-2000 period is presented in Figure 4.30 together with composites for the same anomalous years. North of 20°S , significant westerly anomalies at surface (east of 35°E) and at midtropospheric levels (maximum from 10°E to 35°E but more marked north of 10°S).

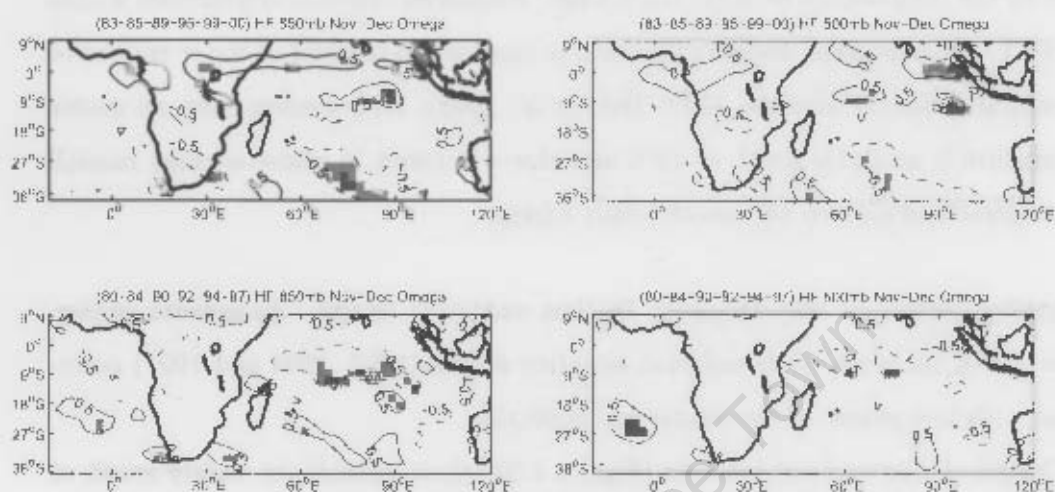


Figure 4.29: (1983, 1985, 1995, 1999 and 2000) composite (top) and (1980, 1984, 1992, 1994 and 1997) composite (bottom) for NCEP R2 HF vertical velocity at both 850mb (left) and 500mb levels (right) in November-December over the 1979-2000 period with shaded areas representing 95% confidence level of Student t-test.

This suggests reduced water vapour transport westward resulting in less moisture available at these latitudes along the east coast of the subcontinent, in particular east of the Great Rift valley, where convection is reduced and thus below-normal rainfall are expected. Over the subtropics (between 25°S and 35°S), broad easterly anomalies particularly significant east of 25°E are characteristic of reduced westerly wind regimes in the midlatitudes: it would result in more moisture accumulating over these regions of southern Africa, as it is not carried eastward off the continental areas by the midlatitude westerly circulation. Enhanced convection processes within the TTT/SICZ location would further act to sustain the transfer of water vapour to the midlatitudes (D'Abreton, 1992; Todd et al., 2002). As a consequence, an excess of moisture is available south of 25°S and this will result in above-average rainfall over central and eastern regions of South Africa.

Negative phase of the tropical Indian easterly mode As noticed earlier, three out of six years of the selected negative events (1992, 1994 and 1997) correspond with low phases of the Southern Oscillation.

Composites in vertical velocity (Figure 4.29) show subsidence widely south of about 15°S , and in particular over the southern tip of the subcontinent where convection processes could be weakened. On the other hand, anomalous uplift is significant within tropical regions of the central Indian Ocean suggesting enhanced convection there. Subsiding anomalies also seem to prevail north of the Equator, and vertical patterns for omega in Figure 4.30 bottom panel could indicate modifications in the meridional circulation. Vertical velocity patterns in Figure 4.29 also exhibit anomalous uplift and downlift respectively to north and south of the Mozambique channel. This could further suggest a potential eastward shift of the ascending Walker type circulation for these years, and thus reduced meridional energy transfer from subtropical latitudes to midlatitudes over continental areas.

Vertical maps of latitudinally averaged zonal moisture fluxes (Figure 4.30 bottom panel) show north of 20°S , easterly anomalies at surface and midtropospheric levels, (maximum between 30°E and 50°E north of 10°S and east of 35°E south of

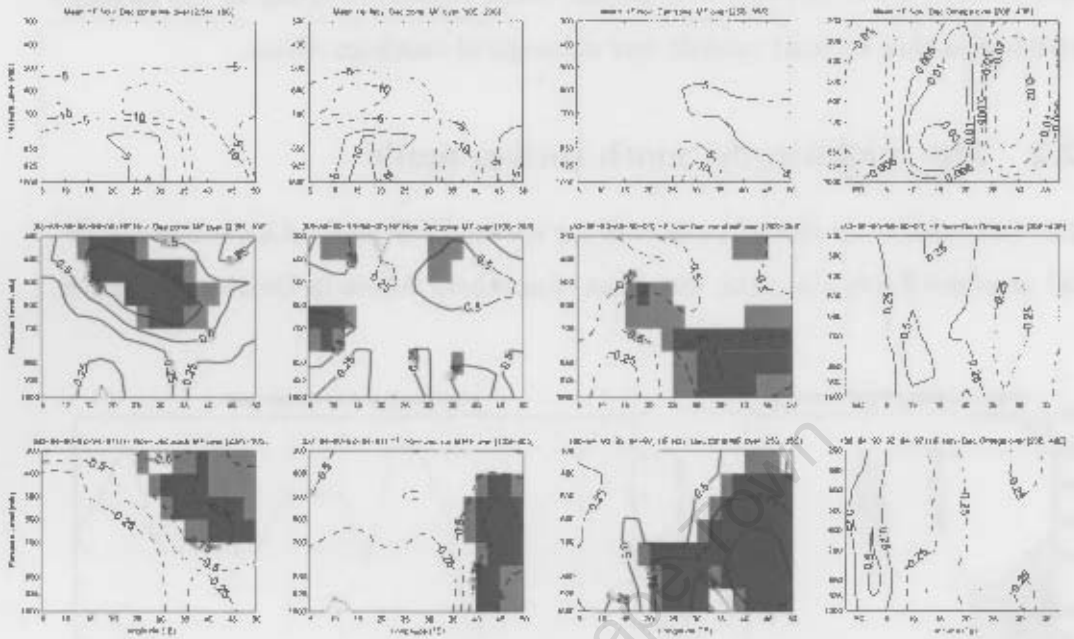


Figure 4.30: Mean (top), (1983, 1985, 1989, 1995, 1999 and 2000) composite (middle) and (1980, 1984, 1990, 1992, 1994 and 1997) composite (bottom) for HF moisture convergence averaged over $[2.5^{\circ}\text{N}-10^{\circ}\text{S}]$ (left), $[10^{\circ}\text{S}-20^{\circ}\text{S}]$ (middle left) and $[25^{\circ}\text{S}-35^{\circ}\text{S}]$ (middle right), with positive/negative values corresponding to moisture convergence/divergence, together with NCEP R2 HF vertical velocity (right) averaged over $[20^{\circ}\text{E}-40^{\circ}\text{E}]$ in November-December over the 1979-2000 period with shaded areas representing 95% confidence level of Student t-test.

10°S), indicating that more humidity is advected westwards from the Indian Ocean at equatorial/tropical latitudes. Enhanced zonal convergence east of the Great Rift valley would lead to above-average rainfall there. South of 25°S, westerly anomalies prevail over the whole vertical domain and suggest enhanced westerly wind regimes locally. Maximums are found east of 15°E, suggesting that most of the water vapour locally is advected to the east over Indian Ocean regions resulting in less moisture available and thus reduced rainfall over subtropical southern Africa.

4.2.2 The midlatitude South Indian mode

Figure 4.31a shows the spatial pattern of the second EOF mode of high-filtered (HF) zonal moisture fluxes along the east coast of southern Africa in October-November.

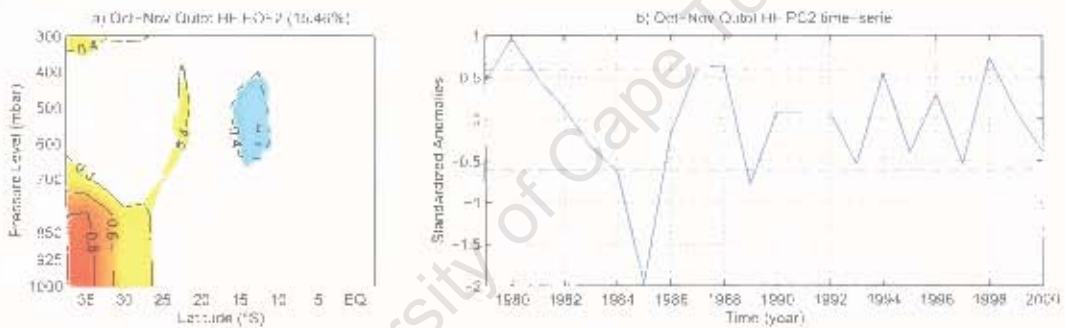


Figure 4.31: Spatial patterns (right) of the second EOF leading mode of variability in high frequency (HF) zonal moisture fluxes along the east coast of southern Africa together with its expansion coefficients (left) in October-November over the 1979-2000 period. The loadings are significant at 95% level using Monte-Carlo simulations.

This mode shows maximum loadings south of 25°S particularly pronounced below 700 mb within the westerly midlatitude circulation and weaker opposite loadings at 500 mb between 10°S and 15°S. The absence of surface loadings at tropical latitudes compared with Figure 4.22 could be linked with the variability in zonal moisture fluxes associated at these latitudes with the monsoon regimes, which

is not reflected in the October-November projection. Positive/negative events of this mode will thus correspond with variations in the circulation linked with the midlatitude westerlies and the South Indian anticyclone.

Heterogeneous correlations between the October-November midlatitude South Indian mode expansion coefficient (ON MFPC2 in the following) and CRU rainfall are presented in Figure 4.32.

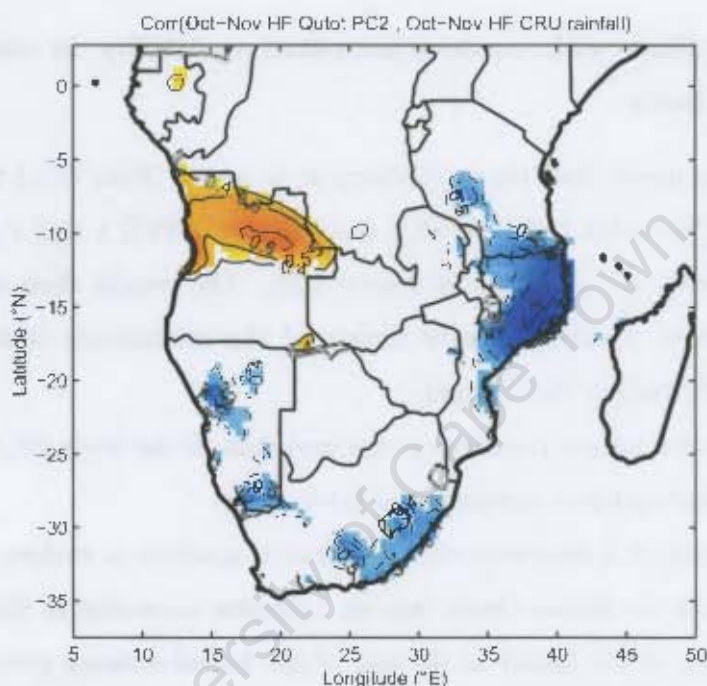


Figure 4.32: Heterogeneous correlations between HF zonal moisture fluxes PC2 and CRU rainfall in October-November over the 1979-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

High loadings for synchronous correlation are found over tropical regions of the sub-continent, showing anticorrelation patterns between western and eastern tropics. Negative maximums are located mainly over northern Mozambique extending to southern/central Tanzania. Significant negative loadings are also found over south-east South Africa and the Namibian interior. Positive loadings are located over northern Angola. Rainfall composites (not shown) for positive/negative years of the

ON MFPC2 time-series (Figure 4.31a) confirm the symmetry of the relation with significant anomalies over these areas of maximum correlation scores.

No significant correlation has been identified between ON MFPC2 and land hydrology data, i.e. both NDVI and soil moisture. It is worth noting that most significant loadings are located in areas which are above the typical rainfall threshold characteristic of the valid use of NDVI, and this could be an explanation for the lack of relationship found with land vegetation for this mode.

4.2.2.1 Connections with ocean-atmosphere variability in the Indian Ocean basin

As for the previous mode, lead/lag correlations up to a year (from ON-1 to ON+1) between October-November (ON) MFPC2 and bimestrial SVD 1 and 2 SST/SLP expansion coefficients are presented in Figure 4.33. The results show significant relationships between positive/negative phases of the midlatitude South Indian mode in October-November (ON 0) and,

- (a) enhanced/reduced surface pressures to the northeast of the basin (SLP SVD 1) the preceding austral summer (maximum in DJ-1/JF 0),
- (b) the establishment of a negative/positive SID-type gradient in surface pressures (SLP SVD 2) across the Indian Ocean regions (opposite anomalies in SLPs to the northeast/southwest of the basin) at the end of the boreal summer preceding the events (maximum in JA 0/AS 0),
- (c) the establishment of a positive/negative SID-type SST gradient (SST SVD 2) across the Indian Ocean (opposite anomalies in SSTs to the northeast/southwest of the basin) at the end of the boreal summer preceding the events (maximum in AS 0/SO 0), just before their development in ON 0. A similar situation seems to develop the following boreal summer (not significant but maximum in SO + 1).

In order to validate some of the relationships, September-October composites are presented in Figure 4.34. The composites confirm that preceding positive/negative events of the midlatitude South Indian mode in October-November, reduced/enhanced surface pressures are found to the southwest of the South Indian anticyclone which

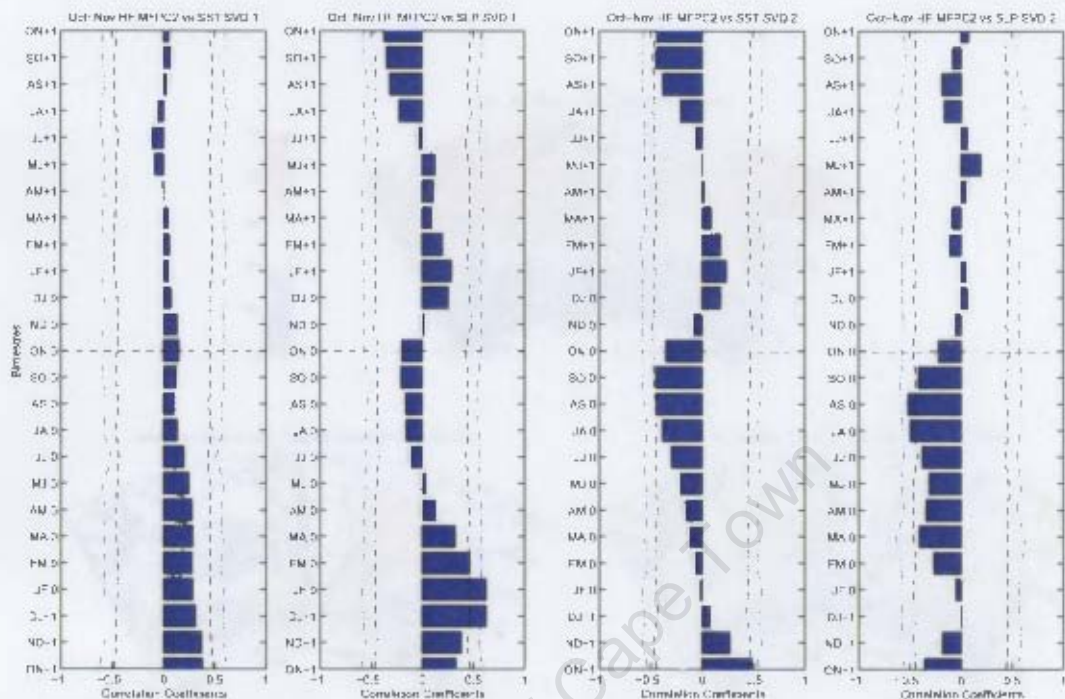


Figure 4.33: Lagged correlations between October-November midlatitude South Indian mode expansion coefficient for high frequency zonal moisture fluxes along the east southern African coast and bi-mestrial SVD 1/2 SST and SLP expansion coefficients (left/right) over the 1979-2000 period. 95% and 99% significance levels using Monte-Carlo simulations are indicated respectively in dashed black/dash-dot blue lines. Year 0 starts with the event and finishes 12 months later.

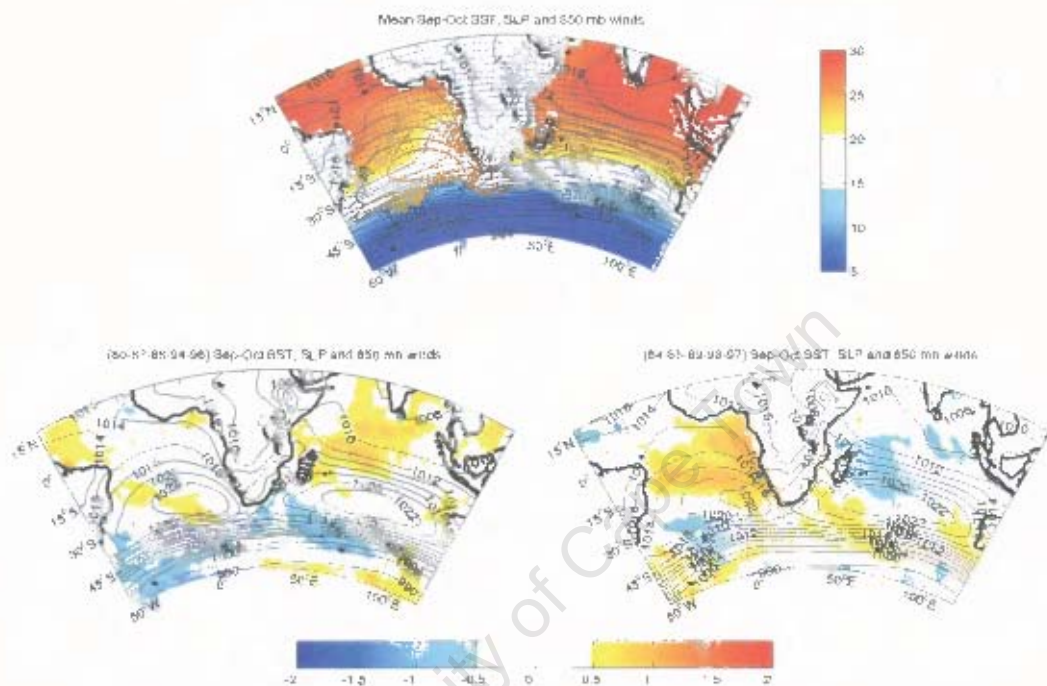


Figure 4.34: Mean 1979-2000 climatology (top), (1980, 1987, 1988, 1994 and 1998) composite (bottom left) and (1984, 1985, 1989, 1993 and 1997) composite (bottom right) for September-October (SO-1) SLPs (contours), SSTs (color) and 850 mb winds (stream lines) significant at 90% level significance of Student t-test.

is ridging less/more towards the subcontinent, while opposite modulations to the northeast witness of enhanced/reduced SLPs there. Furthermore, Figure 4.34 shows a meridional dipole-like pattern in SSTs over southwest Indian Ocean regions during this period. Warm/cold anomalies to the north/south of Madagascar are characteristic of positive events linked with this mode while an opposite situation prevails for negative events. Such structure have been evidenced in earlier studies to impact southern African rainfall (*Washington and Preston, 2006*). In addition, substantial positive SST anomalies are found over the eastern South Atlantic from the tropics to the Benguela region for negative events.

4.2.2.2 Dynamics associated with the midlatitude South Indian mode

From the October-November midlatitude South Indian mode expansion coefficient time-series (Figure 4.31b), extreme years are extracted using a threshold at 0.6 of normalized anomalies deviation from the mean. (1980, 1987, 1988, 1994 and 1998) and (1984, 1985, 1989, 1993 and 1997) were chosen as positive/negative events respectively, for which the expression of the midlatitude South West Indian mode in zonal fluxes appeared to be significantly enhanced/reduced.

Positive phase of the midlatitude South Indian mode Within the positive events isolated above, two years correspond with El Niño (1987 and 1994) and two others with La Niña (1988 and 1998).

In Figure 4.35, vertical velocity composites at surface and midtropospheric levels exhibit anomalous uplift/subsidence to the north/south of the Mozambique channel respectively, with loadings extending over eastern coastal areas, i.e. over north and south coast of Mozambique. Below 25°S, anomalous uplift is found over the South Atlantic and the southern tip of the subcontinent, at surface and midtropospheric levels, with maximums anomalies located between 10°W and 10°E for oceanic areas and over the southeasternmost regions of the subcontinent. Figure 4.36 middle panel shows surface subsidence anomalies in the tropics (from 5°S to 20°S), suggesting reduced convection over tropical southern Africa.

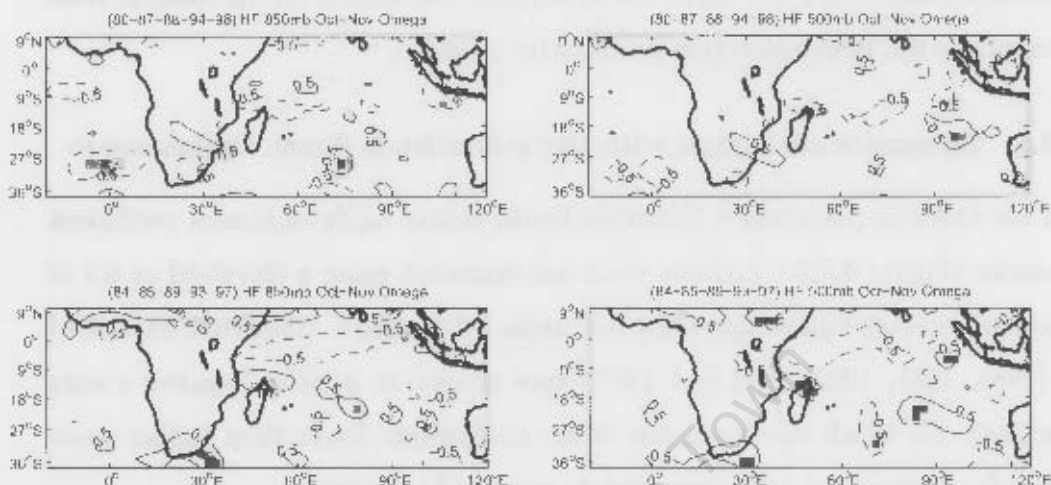


Figure 4.35: (1980, 1987, 1998, 1994 and 1998) composite (top) and (1984, 1985, 1989, 1993 and 1997) composite (bottom) for NCEP R2 HF vertical velocity at both 850mb (left) and 500mb levels (right) in October-November over the 1979-2000 period with shaded areas representing 95% confidence level of Student t-test.

A vertical climatology for October–November zonal moisture fluxes over the 1979–2000 period is presented in Figure 4.36 together with composites for the same anomalous years. In the tropics (from the Equator down to 15°S), rather weak surface westerly/easterly anomalies are found respectively at 30°E and 15°E which could suggest reduced advection of water vapour over central areas of the subcontinent from the east over the Great Rift valley and from the west (i.e. tropical Atlantic regions) towards the subcontinent interior respectively. This marginal enhancement of the westerly flux from the tropical Atlantic could result in more moisture over coastal regions of northern Angola leading to above-normal rainfall locally, while reduced advection to the east of the Congo basin together with weakened convection processes over the surrounding regions would act to reduced rainfall to the north of Mozambique. Between 15°S and 25°S, easterly anomalies are found between 35°E and 40°E and west of 15°E. The latter would result in enhanced moisture transfer from coastal regions of Namibia towards the South Atlantic, and thus drier conditions might prevail at the coast during such events. South of 25°S broad westerly anomalies dominate east of 20°E, which support the idea that the midlatitude westerly circulation is strengthened. Typically this would drive water vapour eastward from continental areas to South Indian Ocean regions, resulting in a deficit of moisture available for convection over continental areas. Together with subsiding anomalies south of 20°S in omega vertical profiles, this would lead to below-normal rainfall over southeastern regions of South Africa.

Negative phase of the midlatitude South Indian mode From the selected negative events, only one (1997) corresponds with the low phase of the Southern Oscillation.

Composites in vertical velocity (Figure 4.35) show statistically significant subsiding anomalies north of the Equator, north of Madagascar and over southernmost regions of the subcontinent. Simultaneous uplift anomalies are found at both middle and lower tropospheric levels north of 15°S, extending broadly from the east of the Congo basin to eastern regions of the tropical Indian Ocean. This would suggest

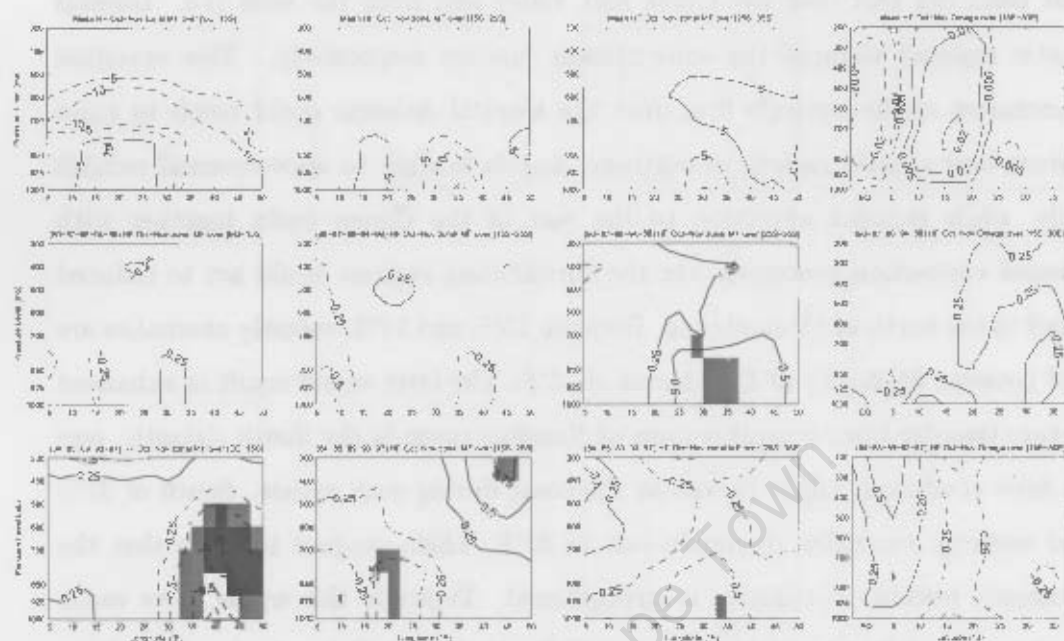


Figure 4.36: Mean (top), (1980, 1987, 1998, 1994 and 1998) composite (middle) and (1984, 1985, 1989, 1993 and 1997) composite (bottom) for IIF moisture convergence averaged over $[2.5^{\circ}\text{N}-10^{\circ}\text{S}]$ (left), $[10^{\circ}\text{S}-20^{\circ}\text{S}]$ (middle left) and $[25^{\circ}\text{S}-35^{\circ}\text{S}]$ (middle right), with positive/negative values corresponding to moisture convergence/divergence, together with NCEP R2 IIF vertical velocity (right) averaged over $[15^{\circ}\text{E}-30^{\circ}\text{E}]$ in October-November over the 1979-2000 period with shaded areas representing 95% confidence level of Student t-test.

enhanced convection processes over these regions, together with an excess of available water vapour at least over oceanic areas. In addition, vertical patterns in Figure 4.36 bottom panel show modification in the meridional circulation resulting in uplift anomalies south of 10°S , further suggesting enhanced convective activity. Vertical maps of latitudinally averaged zonal moisture fluxes (Figure 4.36 bottom panel) show north of 15°S , significant easterly anomalies extending from the surface to midtropospheric levels (500 mb) east of 25°E , characteristic of enhanced advection of water vapour from the tropical Indian Ocean regions. Together with enhanced convection to the east of the Congo basin, it would result in above-rainfall locally, in particular over south Tanzania and north Mozambique. Weak and more marginal westerly anomalies prevailing at surface west of 15°E could create a deficit of moisture at tropical latitudes of the west coast as more humidity would be advected eastwards, potentially leading to drier conditions over northern Angola. Between 15°S and 25°S , significant easterly anomalies particularly pronounced at surface from 15°E to 30°E suggest that less humidity is advected offshore from the Namibian coast where above-normal rainfall are expected. Upper level westerly anomalies east of 35°E prevail to the east of the subtropical jet which is strengthened. It is known that the strongest upper tropospheric divergence occurs when a jet maximum is on the eastern side of a trough where a cyclone then develops, coinciding with the present situation. This pattern definitely corresponds with enhanced cyclogenesis within the subtropical heat-low, driven by a strengthened subtropical jet as described in *Preston-Whyte and Tyson (1988)*, agreeing with uplift anomalies south of 25°S in omega vertical profiles. South of 25°S , broad easterly anomalies indicate a weakened midlatitude westerly circulation, allowing moisture to accumulate over continental areas. Such changes would result in above-normal rainfall over southeastern regions of South Africa.

4.2.3 The equatorial Indian easterly mode

Figure 4.37a shows the spatial pattern of the third EOF leading mode in high-filtered (HF) zonal moisture fluxes along the east coast of southern Africa during the September-October period.

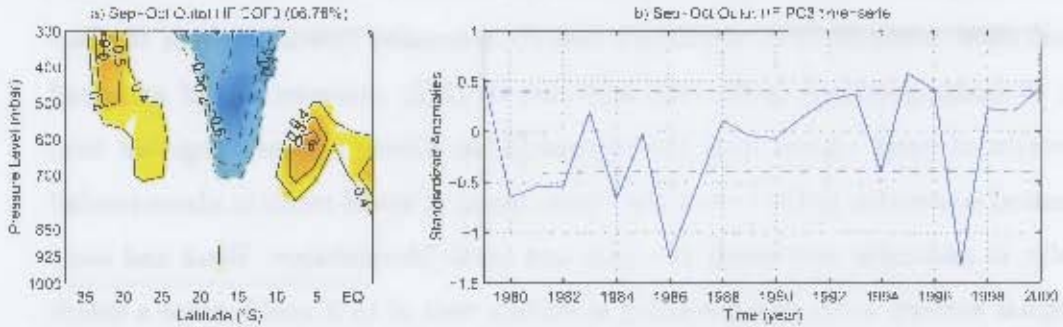


Figure 4.37: Spatial patterns (right) of the third EOF leading mode of variability in high frequency (HF) zonal moisture fluxes along the east coast of southern Africa together with its expansion coefficients (left) in September-October over the 1979-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

The differences in pattern between Figure 4.22 and Figure 4.37a could be due again to the fact that, the loadings at equatorial/tropical latitudes in Figure 4.22 are related to the variability in monsoon regimes between summer and winter, which is not depicted in September-October, when the zonal component of moisture flux at the Equator is not so pronounced. As a result, maximum positive loadings are found during this period at equatorial/tropical latitudes and between 30°S and 35°S, at 700-600 mb and above 400 mb respectively, while maximum negative loadings of the same magnitude are located between 15°S and 20°S at 600 mb. The alternating positive/negative patterns further suggest modifications of the meridional circulation. Considering that most of the water vapour is transported below 500 mb, positive/negative phase of this mode will particularly reflect a reduced/enhanced easterly moisture transfer at midtropospheric levels over equatorial/tropical latitudes linked with alterations in the meridional circulation.

As for the other modes, heterogeneous correlations between the September-October equatorial Indian easterly mode expansion coefficient (SO MFPC3 in the following) and CRU rainfall are presented in Figure 4.38.

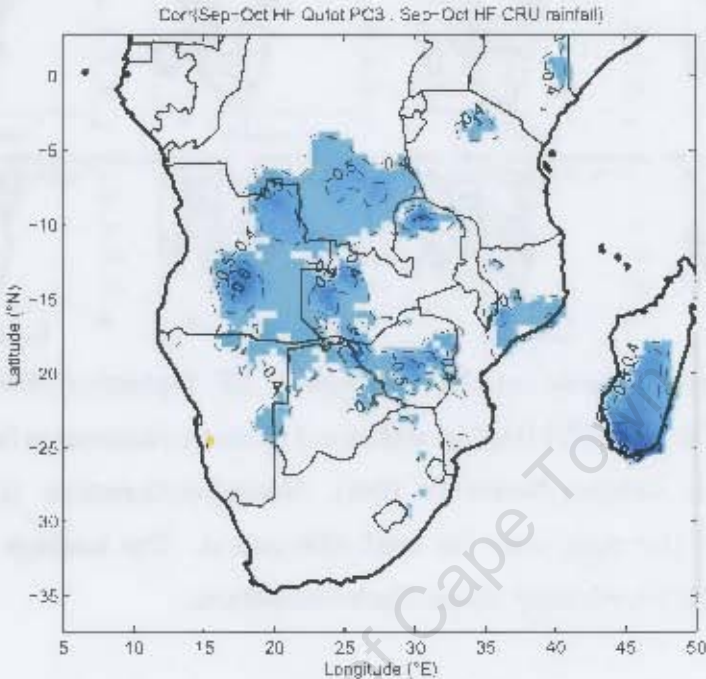


Figure 4.38: Heterogeneous correlations between HF zonal moisture fluxes PC3 and CRU rainfall in September-October over the 1979-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

High negative loadings for synchronous correlation are located mainly over central southern African tropics, from south DRC extending west to east Angola and south across southern Zambia to Zimbabwe, north Botswana and northeast regions of Namibia. Significant negative loadings are also found to the south of Madagascar, and to a lesser extent over eastern parts of the subcontinent, just north of the Zambezi mouth in Mozambique, and over both north Tanzania and northeast Kenya. Rainfall composites (not shown) for positive/negative years of the SO MFPC3 time-series (Figure 4.37b) confirm the symmetry of the relation with significant anomalies over these areas of maximum correlation scores.

To further emphasize the effects on the local hydrological cycle, heterogeneous correlations have been computed with both NDVI and soil moisture (Figure 4.39).

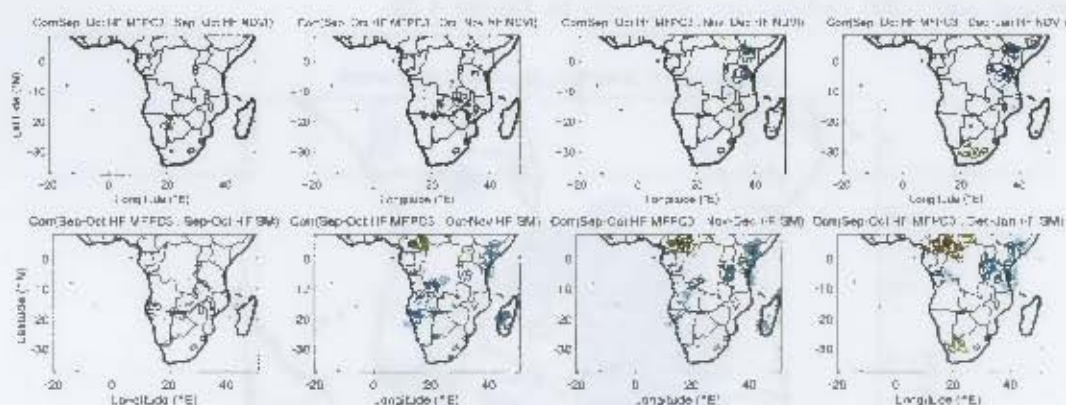


Figure 4.39: Heterogeneous correlations between HF September-October zonal moisture fluxes PC3 and NDVI (top) as well as soil moisture (bottom) in September-October (far left), October-November (left), November-December (right) and December-January (far right) over the 1982-2000 period. The loadings presented are significant at 95% level using Monte-Carlo simulations.

Because the regions where maximum correlation in rainfall are located in areas where rainfall amounts are over the threshold for land hydrology data validity (see Chapter 2), no similar signals are exhibited in vegetation cover as well as in soil moisture. However, negative relationships are found from November-December to December-January in both NDVI and soil moisture, maximum over Tanzania and Kenya mainly. Such findings still indicate that the correlation between NDVI and rainfall is highest when considering precipitated amounts from the concurrent and previous 2 months (Farrar *et al.*, 1994). In soil moisture maximum negative loadings even extend to the north over bordering areas between Kenya, Somalia and Ethiopia. Nevertheless, from October to December, negative loadings in soil moisture over the Bie plateau stretching to the southwest and northeast seem to illustrate the influence of this mode on the local hydrology, potentially through runoff from the plateau to neighbouring areas of lower altitudes. Composites for positive/negative events (not shown) from SO MFPC3 time-series also confirm the symmetry of the relation.

4.2.3.1 Connections with ocean-atmosphere variability in the Indian Ocean basin

As for the other modes, lead/lag correlations up to a year (from SO-1 to SO+1) between SO MFPC3 and bimestrial SVD 1 and 2 SST/SLP expansion coefficients are presented in Figure 4.40. It shows significant relationships between positive/negative phases of the equatorial Indian easterly mode in September-October (SO 0) and,

- (a) the establishment of a negative/positive SID-type gradient in surface pressures (SLP SVD 2) across Indian Ocean regions (opposite anomalies in SLPs to the northeast/southwest of the basin) the previous austral summer (from ND-1 to JF 0, maximum DJ-1). An opposite situation (positive/negative SID-type SLP gradient) is found in austral summer (from SO 0 to FM+1, maximum in ON 0), just after the events in SO 0,
- (b) warming/cooling SSTs to the northeast of the basin (SST SVD 1) the preceding austral summer (from JF 0 to MA 0, maximum in JF 0/FM 0). An opposite situation is found at the end of the events until the end of the following boreal summer (from JF +1 to AS+1, maximum in JA +1),
- (c) reduced/enhanced surface pressures to the northeast of the basin (SLP SVD 1) the preceding boreal summer and persisting during the events until early austral summer (from MJ 0 to SO 0, maximum in JJ 0/JA 0),
- (d) the establishment of a negative/positive SID-type SST gradient (SST SVD 2) across the Indian Ocean (opposite anomalies in SSTs to the northeast/southwest of the basin) at the end of the preceding austral summer and persisting until the development of the events in SO 0 (from FM 0 to SO 0, maximum in MA 0 and JA 0/AS 0).

To validate some of the relationships in August-September, composites are presented in Figure 4.41 and show that preceding positive/negative events (in September-October), reduced/enhanced surface pressures are found to the northeast of the basin while a SID-type SST gradient is less visible. Nevertheless, substantial positive SST anomalies are found over the Benguela region for negative events.

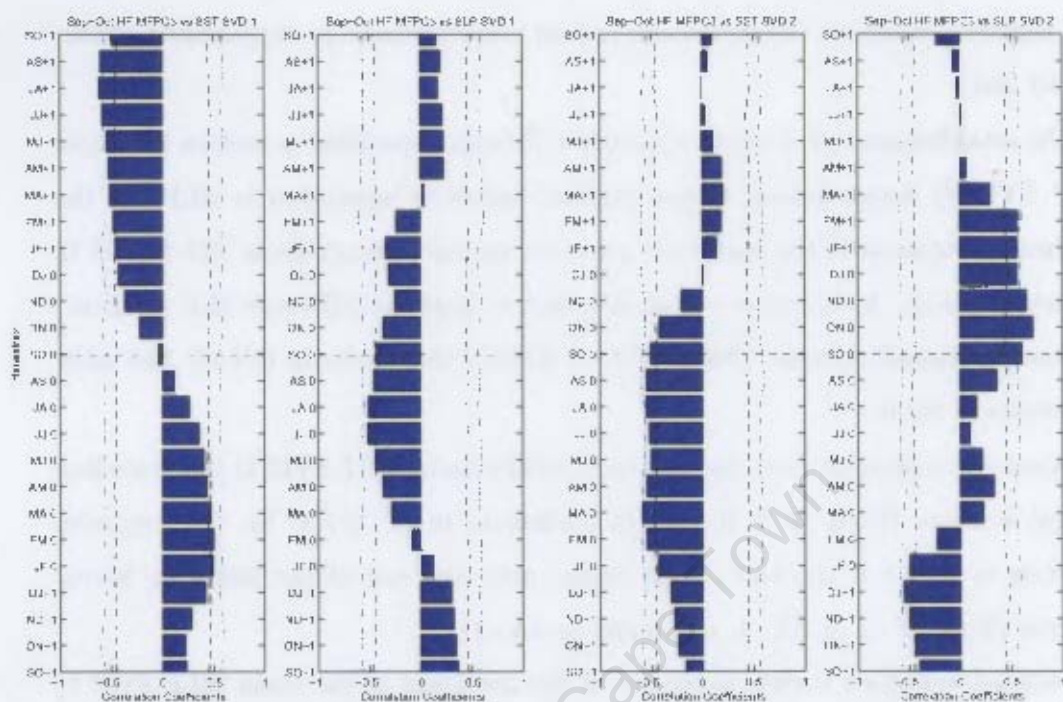


Figure 4.40: Lagged correlations between September-October equatorial Indian easterly mode expansion coefficient for high frequency zonal moisture fluxes along the east southern African coast and bi-mestrial SVD 1/2 SST and SLP expansion coefficients (left/right) over the 1979-2000 period. 95% and 99% significance levels using Monte-Carlo simulations are indicated respectively in dashed black/dash-dot blue lines. Year 0 corresponds to the calendar year of the event.

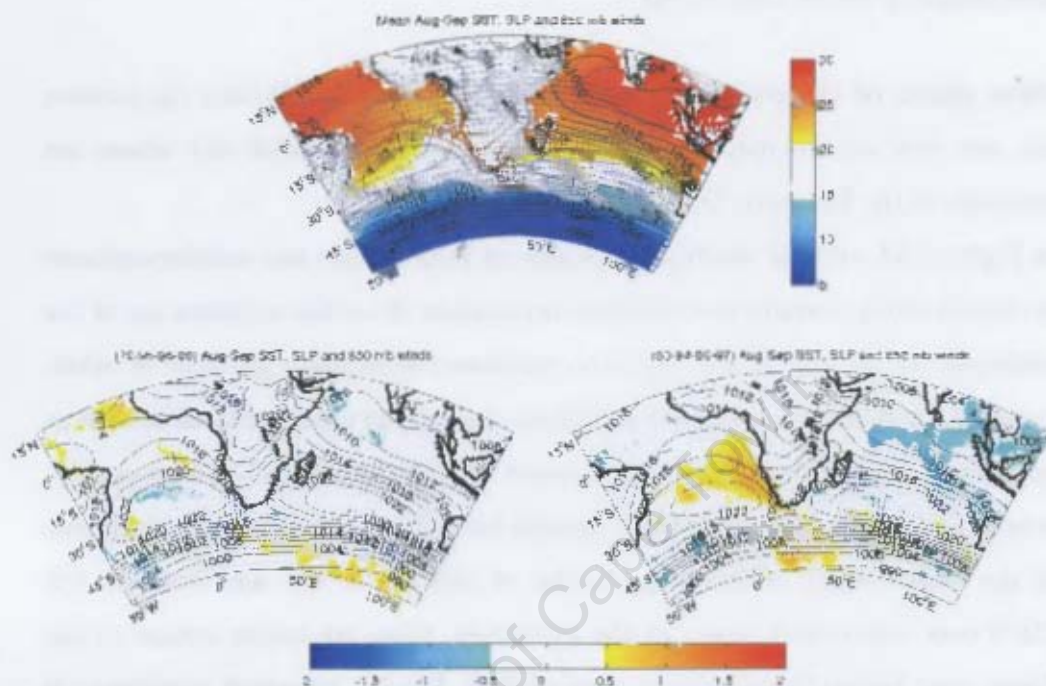


Figure 4.41: Mean 1979-2000 climatology (top), (1979, 1995, 1996 and 2000) composite (bottom left) and (1980, 1984, 1986 and 1997) composite (bottom right) for August-September (AS 0) SLPs (contours), SSTs (color) and 850 mb winds (stream lines) significant at 90% level significance of Student t-test.

4.2.3.2 Dynamics associated with the equatorial Indian easterly mode

From the October-September equatorial Indian easterly mode expansion coefficient time-series (Figure 4.37b) extreme years are extracted. The anomalies being less strong within the time-series, a threshold at 0.4 of normalized anomalies deviation from the mean have been used to define significant extreme years this time. (1979, 1995, 1996, and 2000) and (1980, 1984, 1986, and 1997) were chosen as positive/negative events respectively.

Positive phase of the equatorial Indian easterly mode Within the positive events, one year corresponds to La Niña conditions (2000) while the others are independent of the Southern Oscillation.

In Figure 4.42, vertical velocity composites at both surface and midtropospheric levels exhibit along a southwest-northeast orientation (from the southern tip of the subcontinent to the east of the maritime continent) alternative patterns of subsidence/uplift anomalies. Statistically significant anomalous uplift patterns stretch at surface from central Indian Ocean regions south of 10°S to the north of Madagascar. More marginally, enhanced subsidence prevails broadly at both equatorial latitudes along the west African coast (in particular at 500 mb levels) and between 5°S and 25°S over continental areas. In the subtropics, these anomalies extend to the southwest over Indian Ocean regions west of 60°E . Finally, enhanced subsidence is found over the maritime continent and the west Australian coast. These patterns suggest changes in the tri-dimensional circulation resulting in reduced convection over central southern Africa. Such results are confirmed by omega vertical profiles (Figure 4.43 middle panel) showing enhanced subsidence/uplift respectively between 10°S and 25°S and north of 10°S where statistically significant.

A vertical climatology for September-October zonal moisture fluxes over the 1979-2000 period is presented in Figure 4.43 together with composites for the same anomalous years. North of 10°S , broad westerly anomalies prevail at midtropospheric levels, particularly pronounced between 25°E and 40°E at about 600 mb, indicating that less moisture is advected westwards over the subcontinent interior.

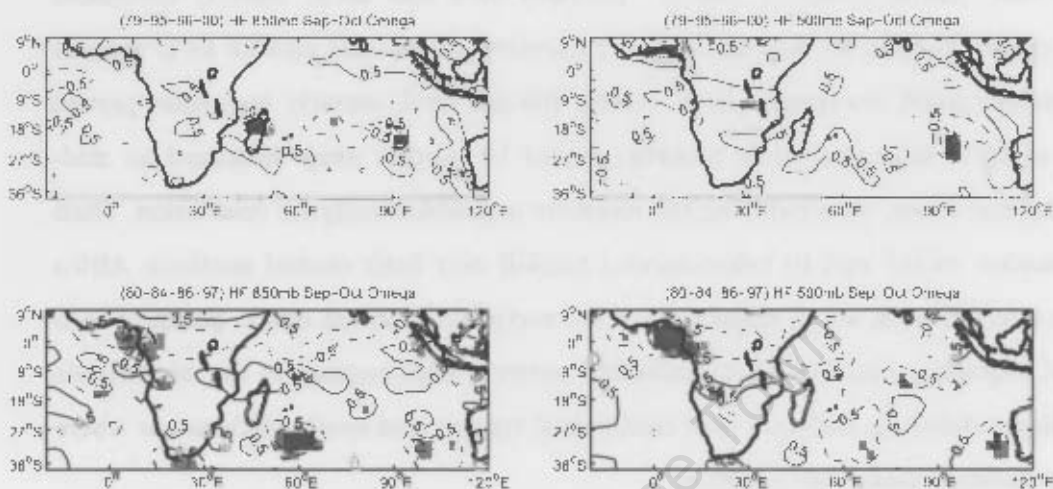


Figure 4.42: (1979, 1995, 1996 and 2000) composite (top) and (1980, 1984, 1986 and 1997) composite (bottom) for NCEP R2 HF vertical velocity at both 850mb (left) and 500mb levels (right) in September-October over the 1979-2000 period with shaded areas representing 95% confidence level of Student t-test.

Together with subsiding anomalies on the eastern parts of the continent at equatorial latitudes, this would result in below-normal rainfall over north Tanzania/Kenya. At surface, easterly anomalies west of 20°E also show reduced water vapour transport from the Atlantic, and this would result in a deficit of moisture over tropical southern Africa interior, where convection is reduced. Upper westerly together with surface easterly anomalies typically illustrate changes in the Walker type circulation, suggesting as shown in omega profiles, the shift of its subsiding limb over central southern Africa. Between 10°S and 20°S , easterly anomalies and enhanced westerly flux at surface respectively west and east of 15°E suggest anomalous zonal divergence there. Above 700 mb level, easterly anomalies prevail east of 25°E indicating that moisture would be carried away westward at mid-tropospheric levels, thus reducing the moisture available locally for convection. Such a situation would lead to below-normal rainfall over both central southern Africa and eastern coastal areas, mainly along the north Mozambican coast. South of 25°S broad westerly anomalies reflect enhanced westerly wind regimes in the midlatitude creating a deficit in moisture over continental regions and south Madagascar where below-normal rainfall are expected.

Negative phase of the equatorial Indian easterly mode Within the negative sample considered, one event corresponds with El Niño conditions (1997) while the rest of the events are independent of the Southern Oscillation.

Composites in vertical velocity (Figure 4.42) show uplift anomalies at both middle and lower tropospheric levels from equatorial latitudes to about 10°S , statistically significant over the Guinea Gulf, while subsidence is found to the south of the Congo basin. This could indicate enhanced/reduced convection processes over these regions of the subcontinent respectively. Alternating patterns of subsidence/uplift anomalies also support changes in the meridional circulation. Vertical profiles for omega (Figure 4.43 bottom panel) typically show anomalous uplift between 5°S and 20°S further illustrating enhanced convection processes over central regions of southern Africa, while reduced/anomalous uplift is to be found north of 5°S and

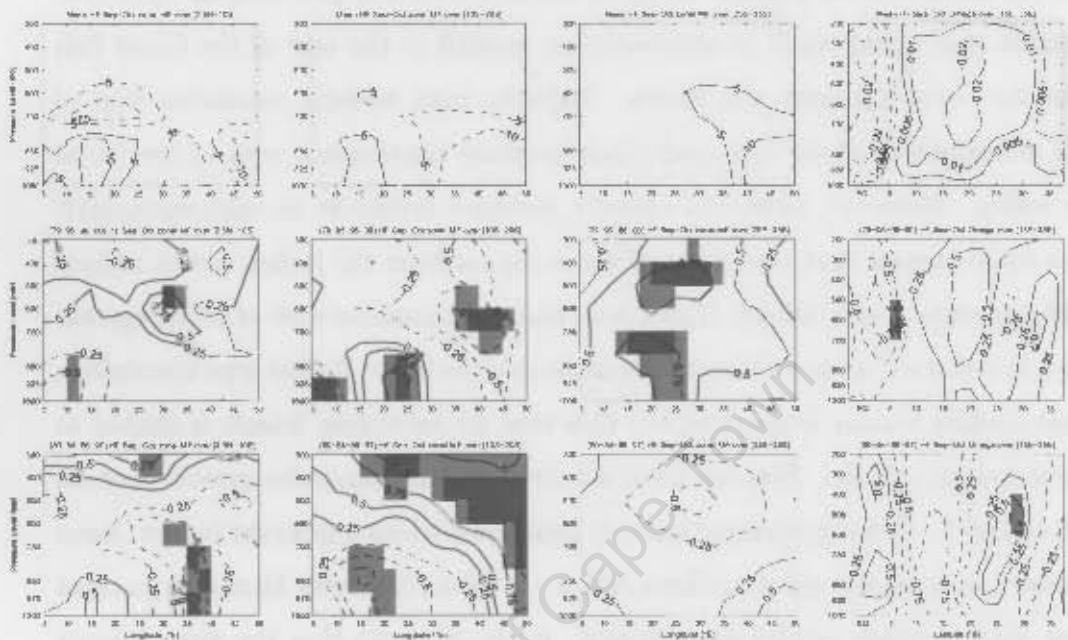


Figure 4.43: Mean (top), (1979, 1995, 1996, and 2000) composite (middle) and (1980, 1984, 1986, and 1997) composite (bottom) for IIF moisture convergence averaged over $[2.5^{\circ}\text{N}-10^{\circ}\text{S}]$ (left), $[10^{\circ}\text{S}-20^{\circ}\text{S}]$ (middle left) and $[25^{\circ}\text{S}-35^{\circ}\text{S}]$ (middle right), with positive/negative values corresponding to moisture convergence/divergence, together with NCEP R2 IIF vertical velocity (right) averaged over $[15^{\circ}\text{E}-35^{\circ}\text{E}]$ in September-October over the 1979-2000 period with shaded areas representing 95% confidence level of Student t-test.

south of 30°S/between 20°S and 30°S respectively. Such changes further suggest alterations in the meridional circulation.

Vertical maps of latitudinally averaged zonal moisture fluxes (Figure 4.43 bottom panel) show north of 10°S, broad easterly anomalies east of 25°E up to 500 mb levels which are the manifestation of an enhanced equatorial easterly moisture flux from the Indian Ocean regions. With enhanced convection processes in equatorial latitudes, this would result in above-average rainfall to the east of the Great Rift valley, i.e. over Tanzania and Kenya. Together with westerly anomalies west of 25°E at surface, it shows enhanced zonal moisture convergence west of the Great Rift valley. Moreover, enhanced easterly moisture transport at midtropospheric levels could suggest that the excess of water vapour from the Indian Ocean regions could penetrate deeper inland. Upper level easterly anomalies west of 10°E together with lower westerly anomalies could illustrate changes in the Walker type circulation, omega profiles further supporting the idea that its ascending branch is shifted to central southern Africa. Between 10°S and 20°S, easterly anomalies prevail between 10°E and 40°E, showing enhanced easterly moisture advection from the Indian Ocean to eastern regions of tropical southern Africa, i.e. along the north Mozambican coast where above-average rainfall are expected. It also suggests that the water vapour excedent from equatorial latitudes is channelled at surface inland to the southern tropics, west of the Great Rift valley. A reduced westerly flux from the tropical Atlantic is noticed at these latitudes. Westerly anomalies dominate from middle to upper tropospheric levels typically showing weakened easterly transport at these altitudes which would favour the accumulation of moisture to the south of the Congo basin. Together with enhanced convection processes as found in omega maps, above-normal rainfall are expected over central regions of southern African tropics. On the other hand, westerly anomalies at surface east of 40°E support the idea that more moisture is advected to the south of Madagascar, which might also lead to enhanced precipitated volumes there. South of 25°S, more marginal easterly anomalies at midtropospheric levels could show a reduced westerly midlatitude circulation at these altitudes.

4.2.4 Summary

An EOF analysis on high-pass zonal moisture fluxes along the east coast of southern Africa helped identify three main modes of variability. They can be listed as follow,

The tropical Indian easterly mode Variations in intensity of the Indian monsoon easterly flux at tropical latitudes and simultaneous changes in the midlatitude westerly circulation during early summer months, i.e. November-December (contributing to 19% of the variance explained) seem to directly modulate rainfall amounts over eastern parts of the subcontinent -to the east of the Great Rift escarpment- and over central as well as southeastern regions of South Africa. The patterns in rainfall are less widespread than these obtained for the modes along the west coast for instance, still impacts on land hydrology are noticeable mainly over Kenya with signals in vegetation cover and soil moisture persisting until February-March. Influences on southern parts of the subcontinent are less marked but still changes within the vegetation canopy are to be noticed over central/southeastern regions of South Africa.

This mode exhibits variability at interannual time-scales with high frequency background signal. There is no particular evidence of strong linkage between ocean-atmosphere variability within the Indian Ocean regions and this mode of variability in zonal moisture fluxes. Nevertheless, in October-November preceding positive/negative events of this mode (in November-December), enhanced pressures are found to the southwest/northeast of the South Indian anticyclone centre where it is strengthened.

During these years corresponding with the positive phase of the tropical Indian easterly mode (1983, 1985, 1989, 1999 and 2000), the tropical easterly flux of moisture from the Indian Ocean was reduced and the midlatitude westerly circulation weakened. It seems that convective activity is reduced over central Indian Ocean regions and less humidity as well as reduced convection are found over equatorial regions east of the Great Rift valley (typically central Kenya), where below-normal rainfall are expected. In addition, weakened westerly wind regimes in the midlati-

tudes result in more moisture available over continental areas south of 25°S where convection and meridional transport of energy are enhanced within the SICZ, thus resulting in above-normal rainfall over central and eastern regions of South Africa. This is consistent with previous studies (*Preston-Whyte and Tyson, 1988*) stating that during southern African wet spell periods, enhanced meridional energy fluxes result in the poleward shift of storm tracks.

For years corresponding with the negative phase of the tropical Indian easterly mode (1980, 1984, 1990, 1992, 1994 and 1997), the tropical easterly flux of moisture from the Indian Ocean was sustained while the midlatitude westerly circulation was strengthened. With enhanced convective activity over central Indian Ocean regions, more moisture appears to be advected at equatorial latitudes east of the Great Rift valley, where zonal moisture convergence is sustained and above-normal rainfall are expected. Strengthened westerlies in the midlatitudes create a situation where water vapour is strongly advected from continental areas eastwards to Indian Ocean regions resulting in a deficit of available moisture over the subcontinent south of 25°S. Substantial modifications in the Walker circulation further suggest an eastward shift of its ascending branch over the western Indian Ocean and southeastern African regions. Such changes would result in reduced convection and meridional energy flux from the subtropics to the midlatitudes, and thus, below-normal rainfall are expected over central and eastern regions of South Africa. These results agree with the work of *Preston-Whyte and Tyson (1988)* who found that during dry spell years, the meridional energy transfer is diminished and the westerly storm tracks move equatorward. Considering that three out of the six years chosen are El Niño events, it is worth noting that these results correspond with previous findings showing similar modifications in the three dimensional circulation during low phases of the Southern Oscillation linked with below-normal rainfall over the subcontinent (*Preston-Whyte and Tyson, 1988; Reason, 1998; Reason et al., 2000; Richard et al., 2000*).

The midlatitude South Indian mode Modulations of the midlatitude westerly circulation in the southwest Indian Ocean regions (contributing to more than 15% of the variance explained) during early summer months, i.e. October-November, are found to influence rainfall amounts over southeast regions of South Africa and southern African tropics with anticorrelation patterns between north Angola and northern Mozambique/southern Tanzania. Here again, the patterns in rainfall are less widespread than for theses obtained for the modes along the west coast, and no significant impacts on land hydrology are identified, which might be due to the limitations within land hydrology data for these tropical regions of the subcontinent.

Most variability associated with this mode is found at interannual time-scales with a marked background seasonal signal. Relationships are identified with ocean-atmosphere variability within the Indian Ocean regions: particularly in September-October preceding positive/negative events of this mode (in October-November), reduced/enhanced surface pressures are found to the southwest of the South Indian anticyclone which is ridging less/more towards the subcontinent, while opposite modulations to the northeast indicate enhanced/reduced SLPs there. A meridional dipole-like structure in the southwest Indian Ocean is characteristic of SST anomalies during this period, consistent with earlier studies (*Washington and Preston, 2006*) reporting that such prevailing conditions could drive an easterly flux of moisture to the north of Madagascar leading to substantial modulations in southern African rainfall, particularly over eastern tropical regions where maximum relationships are found in the present analysis.

1980, 1987, 1988, 1994 and 1998 were years for which the midlatitude westerly circulation was strengthened. The dynamical processes involved for such events are show a less coherent structure than for other modes for instance. Nevertheless it seems that less humidity is found over southeastern regions of South Africa, thus affecting local convection and leading to below-average rainfall there. On the other hand, less moisture appears to be transfered from the tropical Indian Ocean regions to the east of the Congo basin, typically east of 30°E, which would bring drier conditions mainly over south Tanzania/north Mozambique. Interestingly,

reduced advection of water vapour inland from the tropical Atlantic leads to more moisture available for convection over northern Angola where above-normal rainfall are expected. To the southwest, enhanced moisture transfer from coastal regions of Namibia towards the South Atlantic, would lead to reduced rainfall.

During those years when the midlatitude westerly circulation was weakened (1984, 1985, 1989, 1993 and 1997), more moisture seems to accumulate over the mid-latitude continental areas, where convection is enhanced through sustained upper divergence driven by a strengthened subtropical jet. This will act to sustain the meridional transfer of energy to the midlatitudes, thus leading to increased rainfall over southeast regions of South Africa. A reduced offshore advection of water vapour from the Namibian coast is found to result in above-normal rainfall over both Namibia and northwest South Africa. Enhanced moisture transfer from the tropical Indian Ocean regions towards the east of the Congo basin together with enhanced convection locally, lead to above-normal over south Tanzania and north Mozambique, agreeing with the findings of *Washington and Preston* (2006). On the other hand, the westerly flux from the tropical Atlantic towards the interior appears to be sustained over coastal areas to the west of the subcontinent, and this would create a deficit of moisture locally bringing drier conditions over northern Angola.

The equatorial Indian easterly mode Variations in intensity of the equatorial easterly flux from the Indian Ocean accompanied by changes in the northern trades latitudes and in the midlatitudes (contributing to 7% of the variance explained), reflecting alterations in the meridional circulation, seem to modulate early in the summer season (September-October) rainfall amounts over central southern African tropics, all along the south Congo basin, from south DRC and eastern Angola, to southern Zambia and Zimbabwe. An influence is also noticeable to the south of Madagascar and over eastern parts of the subcontinent, in particular over north Mozambique, north Tanzania and north Kenya. Despite the fact that the patterns obtained in rainfall are well widespread, nevertheless they appear to be less strong than those obtained for the modes along the west coast for instance.

Substantial impacts on land hydrology are most pronounced over northeastern regions of the subcontinent (i.e. Tanzania and Kenya) where changes persist until December-January suggesting the importance of early summer rainfall for the rest of the seasonal cycle locally. Modifications in soil moisture are also found around the Bie plateau supporting the idea that lower-elevated neighboring areas are potentially affected by runoff from the plateau. The absence of signal in NDVI probably reflects the limitations within the data for these regions.

This mode is characterized by variability at interannual time-scales essentially. Relationships are identified with ocean-atmosphere variability within the Indian Ocean regions: particularly in August-September preceding positive/negative events of this mode (in September-October), reduced/enhanced surface pressures are found to the northeast of the Indian Ocean basin.

1979, 1995, 1996 and 2000 were years corresponding with the positive phase of the equatorial Indian easterly mode when the meridional circulation was modified leading to reduced moisture input from the Indian Ocean at equatorial/tropical latitudes. Simultaneous changes in the Walker type circulation would bring its subsiding limb over central African regions. As a result, less humidity as well as reduced convection are found over central regions of southern African tropics and equatorial/tropical coastal areas to the east (i.e. over Tanzania and Kenya as well as along the north Mozambican coast), where below-normal rainfall are expected. South of 25°S, the westerly midlatitude circulation seems strengthened, carrying most of the moisture out of continental areas, including south Madagascar where rainfall amounts are reduced.

During those years corresponding with the negative phase of the equatorial Indian easterly mode (1980, 1984, 1986 and 1997), the meridional circulation was modified leading to enhanced moisture input from the Indian Ocean at equatorial/tropical latitudes. Convection is found to be enhanced at equatorial/tropical latitudes thus resulting in increased rainfall over tropical regions along the east coast of southern Africa. The excess of moisture seems to be channeled at surface levels (and also potentially at midtropospheric levels, where the easterly flux from the Indian Ocean

regions might be strong enough to penetrate deeper inland to the southern tropics), and zonal moisture convergence is favoured to the west of the Great Rift valley. Simultaneous changes in the Walker type circulation bring its ascending limb over the central southern Africa further favouring convection there. In conclusion, above-average rainfall are expected over eastern regions of southern African tropics (i.e. from north Mozambique, Tanzania to Kenya) but also deeper inland across to central regions south of the Congo basin. Anomalous westerly advection of moisture at surface over southern parts of Madagascar could also be responsible for increased precipitated volumes locally.

Key players of zonal water vapour transport over southern Africa from both the South Atlantic and Indian Oceans have been identified during summer. The use of multivariate analyses helped to isolate and find how they modulate substantially rainfall regimes over distinct regions of the subcontinent. Significant changes in precipitated amounts are found to further persist within the hydrological cycle.

Potential links are identified between primary modes of variability in zonal moisture fluxes along the east/west coasts of southern Africa and both ocean-atmosphere systems surrounding the subcontinent. Strongest relationships illustrate of the influences of modulations in intensity and in position of the subtropical anticyclones regarding moisture input from both oceanic basins. Potential links with anomalies the surface oceans are noticed for some of the modes.

The atmospheric mechanisms involved reveal the importance of tropical and subtropical convective systems and how zonal water vapour transport at surface and midtropospheric levels act to alter their intensity and persistence.

Chapter 5

Summary findings and conclusions

The aim of this thesis has been to work toward an identification of the key mechanisms in moisture input from both the South Atlantic and Indian Oceans, their variability, and how they modulate processes related to convection and ultimately rainfall regimes over both tropical and subtropical regions of southern Africa.

Moisture transport is essential to rainfall over the subcontinent and, particularly in summer, the zonal circulation is the most important. As fundamental hypotheses for this work, it has been assumed that zonal moisture transport from the oceans can condition convective activity and thus rainfall regimes in southern Africa. Furthermore, zonal water vapour exchange at the coasts has been considered as a potential link between summer rainfall over the subcontinent and ocean-atmosphere variability in the neighbouring oceanic basins. Given the substantial importance of meridional moisture transport, this represents one of the limitations of the chosen approach.

In testing these hypotheses, the annual cycle has been investigated, as well as variations at interannual time scales and connections with large scale ocean-atmosphere variability. Several findings helped to draw a consistent picture in terms of prevailing continental convection processes and main features of moisture input from the oceanic basins over southern Africa.

They are listed as follows:

(a) The calculations of moisture fluxes and divergence from NCEP R2 re-analyses helped identify three key areas for convection, located to the east of the Congo basin along the Great Rift valley (at about 4°S and 30°E), within the Angola low location over the Bie plateau (at about 17°S and 18°E) and to the south of Botswana (at 27°S and 25°E) corresponding with the subtropical heat-low location.

(b) Averaging zonal moisture fluxes along the west and east coasts of southern Africa also helped emphasize major fluxes at the land-ocean interfaces which could potentially modulate water vapour exchange between the neighbouring oceans and the subcontinent.

Along the west coast in summer, from the Equator to 15°S , a westerly monsoon-like flux is present at surface levels. Overlying this westerly flux is found the southern extension of the African Easterly Jet (AEJ) at midtropospheric levels. To the south (between 17°S and 32°S), an easterly flux prevails, driven by the South Atlantic anticyclone. South of 32°S , a westerly flux associated with the midlatitude circulation is occupying most of the air column.

Along the east coast in summer, an easterly flux corresponding to the northeast monsoon prevails at surface from the Equator to about 10°S . Between 10°S and 20°S a westerly flux linked with the cyclonic circulation over the Mozambique channel is found between 850 mb and 700 mb in summer. To the south (between 20°S and 35°S) an easterly flux is characteristic of the southwest Indian trades. Further south, a westerly flux associated with the midlatitude circulation is present all along the air column, but it is most pronounced at midtropospheric levels.

Investigating ocean-atmosphere variability within both the South Atlantic and Indian Ocean basins, primary modes were isolated over the 1962-2002 period:

(a) For the South Atlantic region, a first mode of variability consists of the warming/cooling in the tropical Atlantic resulting in a shift of maximum temperature gradient, consistent with the findings of *Wallace et al.* (1990), *Kushnir* (1994), *Venegas et al.* (1997) and *Colberg* (2006); *Colberg and Reason* (2006). It is accompanied by a simultaneous weakening/strengthening of the South Atlantic anticyclone modulated by reduced/enhanced surface pressures in the tropical Atlantic. Northerly/southerly anomalies in the central basin could be the response of the wind field to the induced pressure gradient. Reduced/enhanced trades in the western equatorial Atlantic could be responsible for SST anomalies there, but the absence of relationship with winds in the southeastern tropics where warm/cold SSTs also develop rather suggests non-local influences, such as the propagation of equatorial Kelvin waves as proposed in earlier studies (*Carton and Huang*, 1994; *Florenchie et al.*, 2003, 2004; *Colberg*, 2006; *Colberg and Reason*, 2006). The induced changes by the SST anomalies in the atmospheric circulation could in turn strengthen the existing SST anomaly by air-sea heat exchanges as noticed in *Venegas et al.* (1997). This mode shows variability at interdecadal (14 years) and interannual (4-5 years) time-scales. Variability in the southeast Atlantic have been shown to impact rainfall regimes over the subcontinent (*Rouault et al.*, 2003) suggesting that this mode could be of importance to southern African climate.

A second mode reflects cooling/warming in the midlatitude South Atlantic SSTs, accompanied by the weakening/strengthening of the South Atlantic anticyclone due to reduced/enhanced surface pressures in the midlatitudes. Such changes prior to positive/negative events would result in the anomalous southwestward/northeastward advection of warmer and moister/colder and drier air, thus reducing/sustaining latent heat flux in the southwest basin. As the events develop, a dampening of the original latent heat flux anomalies could be due to stronger/weaker-than-normal northerlies in the southwest midlatitude South Atlantic, thus cooling/warming

SSTs until the events peak. Interdecadal as well as interannual variations are found to prevail in this mode of variability, and substantial links are found with ENSO. Connections have been evidenced between SST variability in the midlatitude South Atlantic, and climate anomalies over the subcontinent (*Mason, 1995*), and thus this mode could have substantial implications regarding southern African summer rainfall variability.

In the tropical sector, a first mode resembling the Atlantic Niño or equatorial mode (*Merle, 1980; Zebiak, 1993; Carton and Huang, 1994; Ruiz-Barradas et al., 2000*) illustrates warming/cooling of SSTs in the eastern part of the basin accompanied by a potential relaxation/intensification of the equatorial trades locally, and reduced/enhanced surface pressures to the west of the ITCZ. Such mechanisms could correspond with the thermocline/SSTs/winds feedback accompanied by the latitudinal shift of the ITCZ, as proposed in earlier studies (*Carton and Huang, 1994; Ruiz-Barradas et al., 2000*). At a few months lag, reduced/enhanced northeasterlies in the western parts of the tropical South Atlantic could lead to reduced/enhanced evaporation and thus act to sustain the original SST anomalies. Both interdecadal (14 years) and interannual (2.5 years) variability are found to dominate this mode.

A second mode of variability consists of a warming/cooling in the northern/southern tropical Atlantic accompanied by an enhanced pressure gradient within the thermal equator. It resembles the interhemispheric or dipole mode (*Hastenrath and Greischar, 1993; Chang et al., 1997; Ruiz-Barradas et al., 2000*). The pressure gradient is found to drive a cross-equatorial flow, which could impact the meridional position of the ITCZ (*Hastenrath and Greischar, 1993*). As a result, winds are enhanced in the hemisphere of negative SST anomalies where further cooling of the surface ocean is sustained through enhanced evaporation, while reduced winds in the hemisphere of warm SSTs would lead to further warming locally. Such findings agree with earlier studies which have evidenced similar mechanisms (*Ruiz-Barradas et al., 2000; Kushnir et al., 2002; Chang et al., 2006*) referred to as the Wind-Evaporation-SST (WES) feedback (*Xie and Philander, 1994*). Interdecadal variability (12 years) is

characteristic of this mode. Recent works have emphasized links between anomalous meridional SST gradient in the equatorial Atlantic, and precipitations, particularly over the northern coast of the Guinea Gulf (*Chang et al.*, 2006), which suggest that this mode could be of importance in modulating local climate.

(b) For Indian Ocean regions, the first mode of variability can be interpreted as the recent warming noticed in the basin, associated with modulations in surface pressures particularly pronounced over northeastern parts. In addition this mode is also dominated by interannual variability (3-4 years) and shows strong teleconnections with ENSO when the warming in the Pacific leads by 4 months.

The second mode consists of a dipole in SSTs with warm/cold SSTs to the southwest/northeast of the basin, resembling the subtropical Indian Ocean dipole (SID) mode of *Behera and Yamagata* (2001). The development of positive events seem to be preceded by enhanced/reduced surface pressures to the southwest/northeast of the basin respectively. This could correspond to the establishment and migration of the west coast trough over western coastal regions of Australia in austral summer, while the South Indian anticyclone is shifted eastward during this period. The resulting pressure gradient could drive strong southeasterlies off Australia, leading to the anomalous advection of colder and drier air. This would enhance evaporation as well as upper ocean mixing and thus cools SSTs locally. Reduced midlatitude westerlies to the southwest of Madagascar could bring anomalous warmer and moister air reducing evaporation, while the equatorward Ekman transport of cold high latitudes waters is weakened, thus acting to warm SSTs there, in agreement with *Behera and Yamagata* (2001). At the end of austral summer when high pressures establish over Australia, the subtropical high migrates northwestwards and the prevailing easterlies and westerlies are shifted which could act to terminate the event (*Behera and Yamagata*, 2001). Both interannual (mainly at 3 and 7 years) and interdecadal (around 13 years) variability are found to dominate this mode. Correlations between SID events and rainfall anomalies over the subcontinent (*Behera and Yamagata*, 2001) suggest that this mode could have influences on

southern African climate.

In the tropical sector, a leading mode consisting in SSTs of positive/negative loadings off the East African coast and to the southwest of Sumatra respectively, resembles the Indian dipole mode (IOD) or Indian Ocean zonal (IOZ) mode (*Saji et al.*, 1999; *Webster et al.*, 1999). As stated by *Saji et al.* (1999), a strong relationship is also found between the intensity of the SST dipole and the strength in the easterlies prevailing in the equatorial latitudes over the central basin. This mode appears to be strongly phase locked on the seasonal cycle and is reported to have its peak in austral spring (*Webster et al.*, 1999; *Saji et al.*, 1999; *Basquero-Bernal et al.*, 2002; *Fauchereau*, 2004). Interannual variability (at about 3.5 years) characterizes this mode of variability and potential connections are found with ENSO. In its extreme years the IOD mode has been found to be connected with rainfall anomalies particularly over eastern Africa (*Saji et al.*, 1999).

The variability in water vapour transport from the South Atlantic and Indian Oceans onto the subcontinent was then investigated by extracting primary EOF modes of zonal moisture fluxes along the west and east coasts of southern Africa. Relationships between the hydrological cycle and variability in the ocean-atmosphere system of each neighbouring basins were further examined.

Concerning moisture input from the South Atlantic;

(a) The modulation in intensity as well as the latitudinal displacement of the South Atlantic anticyclone are found to contribute the most (about 25% of variance explained) to variability in moisture input from the South Atlantic onto the subcontinent at subtropical latitudes. It is generally accompanied in December-January by variations in the midlatitude South Atlantic SSTs with a north-south dipole-like structure. This mode is found to modulate, in January-February, rainfall amounts south of the Congo basin, together with opposite influences over western coastal regions in the tropics. At subtropical latitudes, impacts on the hydrological cycle are persisting until February-March. This mode is characterized mainly by intra-annual

as well as interannual variability with high frequency seasonal signals.

Its positive/negative phase corresponds to a reduction/increase of available moisture south of 15°S (in January-February). The strengthening/weakening of the zonal component of the southern extension of the AEJ could also modulate the meridional transfer of moisture south of 15°S to the advantage/detriment of Angolan coastal regions, where the advection of less/more moisture inland at surface further leads to above/below rainfall. In its positive phase, strongly related to ENSO, an eastward shift of the ascending branch of the Walker type circulation over the western Indian Ocean and southeastern African regions acts to suppress convection over subtropical regions and thus leads to drier conditions. In its negative phase, convective processes favoured within the SICZ/TTTs region bring above-normal rainfall over southern African subtropics. These results are consistent with the work of *Mason* (1995) who found that variations in midlatitude South Atlantic SSTs through a meridional shift in the midlatitude westerly circulation would lead to rainfall anomalies in southern Africa.

(b) The variations in intensity of the westerly flow that penetrates from the tropical Atlantic, contributing to about 11% of variance explained, are found to be of primary importance regarding moisture availability over tropical regions of southern Africa. This mode appears to modulate rainfall over the uplands surrounding the Congo basin in January-February, i.e. from the east DRC and northwest Tanzania down to south Angola, with impacts on the hydrological cycle persisting until April-May to the east of the Great Rift valley. It is preceded in November-December by a strengthening/weakening of the South Atlantic anticyclone, but there is no evidence of linkage between ocean-atmosphere variability in the South Atlantic and this mode of variability in zonal moisture fluxes. This mode seems dominated mostly by intra-annual as well as interannual variability.

Enhanced/reduced advection of moisture over the Congo basin is accompanied by favoured/inhibited deep convection processes. In the positive phase of this mode, the excess water vapour is channelled from the Congo basin to the east

and southeast at surface, while the southern extension of the AEJ could play a role in transporting more moisture southwards at midtropospheric levels, leading to above-average rainfall. During its negative phase, related to ENSO, a shift eastwards of the ascending branch of the Walker circulation is found to reduce convection and thus rainfall.

Regarding the contribution in moisture input over southern Africa from the Indian Ocean regions;

(a) The modulations of the Indian monsoon easterly flux at tropical latitudes, accompanied by changes in the midlatitude westerly circulation in November-December (contributing to about 19% of variance explained), are found to alter rainfall amounts to the east of the Great Rift escarpment (where impacts on land hydrology persist until February-March) and over central/southeast regions of South Africa. Positive/negative events of this mode of variability are generally preceded by enhanced pressures in October-November, to the southwest/northeast of the South Indian anticyclone centre which is strengthened. Nevertheless, there is no evidence for strong relationship between ocean-atmosphere variability in the Indian Ocean basin and this mode of variability in zonal moisture fluxes. This first mode exhibits mainly variability at interannual time-scales with high frequency seasonal signal.

In the positive/negative phases of this mode, less/more humidity is found at tropical latitudes where convection is reduced/favoured (enhanced zonal moisture convergence could play a role during negative events) leading to above/below-normal rainfall east of the Great Rift valley. To the south, a reduced/sustained westerly circulation is found to enhance/reduce convection within the Botswana heat-low. This acts to sustain/weaken the meridional transfer of energy to the south within the SICZ complex, ultimately leading to above/below-normal rainfall over central and southeastern regions of South Africa for positive/negative events respectively.

(b) The variations within the midlatitude circulation linked with a South Indian anticyclone ridging less/more towards the subcontinent from September-October (contributing to more than 15% of variance explained) appear to lead, in October-November, to below/above-normal rainfall over southeast regions of South Africa and northern Mozambique/southern Tanzania, with opposite effects over north Angola. It is worth noting that the changes in the atmospheric circulation in September-October seems to coincide with meridional dipole-like anomalies in the surface southwest Indian Ocean. This mode is characterized mostly by interannual variability and a marked seasonal signal.

Enhanced/reduced westerly wind regimes in the midlatitudes are found to decrease/increase moisture availability over southeastern coastal regions of South Africa where convection is inhibited/sustained, leading to below/above-normal rainfall. In particular during negative events related to this mode of variability, enhanced upper divergence driven by a strengthened subtropical jet helps favour convective activity over the subtropics. Reduced/sustained advection of moisture inland creates a deficit/excess in rainfall over regions to the north of Mozambique. Such results are consistent with the study of *Washington and Preston (2006)* who found that an anomalous cyclonic circulation corresponding with SST anomalies in the southwest Indian Ocean similar to those identified in September-October, could drive an easterly flux of substantial importance for southern African rainfall, particularly over the eastern tropics. Along the west coast, sustained/reduced offshore transport results in below/above-average rainfall over Namibia, while reduced/increased eastward advection of moisture leads to an opposite situation over north Angola.

(c) The modulations in intensity of the equatorial easterly flux from the Indian Ocean accompanied by changes within the northern trade latitudes, reflecting alterations of the meridional circulation in September-October (contributing to about 7% of variance explained), are found to alter rainfall regimes over central southern African tropics, to the south of Madagascar and over north Mozambique,

north Tanzania and north Kenya where impacts on land hydrology persist until December-January. This mode shows mostly variability at interannual time-scales. Such changes in the meridional circulation are found to reduce/enhance moisture input from the Indian Ocean at equatorial latitudes, leading to reduced/enhanced precipitated volumes at the coast. Simultaneous alterations in the Walker type circulation bring its descending/ascending limb over central southern Africa where less/more moisture is available, inhibiting/favouring convection (the excess in moisture at equatorial latitudes seems to be channeled at surface to the southern tropics, resulting in enhanced zonal moisture convergence west of the Great Rift valley during negative phases) and leading to below/above-average rainfall. South of 25°S , the westerly midlatitude circulation is strengthened/weakened, potentially bringing drier/wetter conditions to the south of Madagascar.

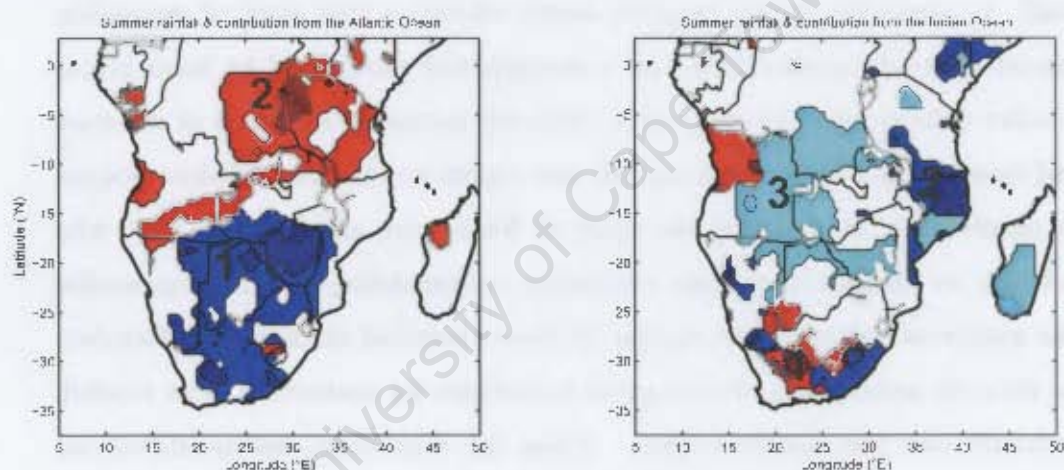


Figure 5.1: Contribution from neighbouring oceans expressed as correlation patterns (above 0.4) between CRU rainfall and primary modes in zonal moisture fluxes along the west/east coasts of southern Africa (left/right). Patterns for the South Atlantic relate to January-February rainfall. Those linked with the Indian Ocean reflect modulations in precipitated amounts in November-October, October-November and September-October for the first, second and third mode respectively.

As a summary, Figure 5.1 shows the contributions from both the South Atlantic and Indian Oceans to summer rainfall variability over the subcontinent. The

variability in zonal fluxes from the South Atlantic Ocean and southern African rainfall, affect mainly rainfall during the transition period between early and late summer season (January-February), while the variations in moisture input from the Indian Ocean regions studied here are found to rather modulate rainfall in the early summer season (from September-October to November-December).

In regards to the existing literature, this study is the first to present a tri-dimensional investigation of interannual variability in moisture fluxes from the South Atlantic and Indian Oceans onto the subcontinent. Such an approach helped provide further elements of description in terms of rainfall variability over southern Africa, and in particular over tropical areas which are of the less studied.

Of course these results are subject to several limitations linked to the chosen approach. Firstly, despite the comparison made between different rainfall dataset, there are still major caveats in rainfall data, in particular over tropical areas of southern Africa. As mentioned earlier, mass imbalance in the re-analyses has been identified in earlier studies (*Alexander and Schubert, 1990; Trenberth, 1991, 1997; Kanamaru and Salvucci, 2003*), and thus the present findings are subject to such uncertainties. Moreover, only 20 years of data were used and this is unfortunately one of the substantial limitations of this study. Finally, zonal features of water vapour transport were considered exclusively and their individual role along the west and east coasts separately, which is a big assumption. It is necessary to integrate their meridional component, as it could be of great importance too. Moreover, cross-interactions between water vapour transfer between the subcontinent and both oceanic basins need to be addressed in order to fully depict the contribution in moisture input from the oceans and how they interact together to enhance/inhibit rainfall during the whole summer season.

In addition further work is needed regarding some of the results:

(a) Some features of water vapour transport have been isolated as quite important for the development of anomalous events leading to wet/dry conditions over the subcontinent. Amongst them, the southern extension of the AEJ as well as the subtropical jet needs to be better documented. The southern extension of the AEJ for instance is not extensively referred to in the literature. It seems to be marked by a strong seasonality and the results show that it could be instrumental in the modulation of mid-summer rainfall over southern Africa. During ENSO events, it could be instrumental in the meridional transport of moisture to the south of the tropics. Some other components of water vapour transport, such as the easterly flow to the north of Madagascar in summer, which has been found to play a substantial role in modulating southern African rainfall (*Washington and Preston, 2006*), show strong signals in the multivariate analyses but have not been investigated in the present study. Focusing on different time-scales/seasons, their examination could definitely bring further elements of description in terms of rainfall variability.

Concerning ENSO, the events chosen for each mode contained years corresponding to the low/high phase of the Southern Oscillation for which typical anomalies have been identified. Generally in the selected samples, El Niño events are characterized by below/above-average rainfall over subtropical/eastern Africa, while an opposite situation is found during La Niña events. Such findings are consistent with the impacts of ENSO commonly described in the literature (*Harrison, 1984; Preston-Whyte and Tyson, 1988; Lindesay, 1988; D'Abreton, 1992; Nicholson and Kim, 1997; Nicholson, 2003*). Nevertheless, it would be worthy to examine deeper these modulations and in a more dedicated manner.

(b) More work is required regarding the links isolated between ocean-atmosphere systems in the neighbouring oceanic basins and modes of variability in zonal water vapour transport at the land-ocean interface.

Along the west coast, relationships between the leading modes of variability in zonal

moisture fluxes and surface pressures/SSTs in the tropical and midlatitude South Atlantic are most pronounced at the end of the preceding boreal summer/beginning of austral summer the year of their associated events respectively.

Along the east coast, the primary modes of zonal water vapour transport show interesting connections with SID-type surface pressures gradients at the end of the preceding boreal summer and during austral summer the year of their corresponding events. On the other hand, links with modulations in surface pressures to the northeast of the basin seem most pronounced from the end of the preceding austral summer until the beginning of the boreal summer following the events associated with the first and third modes. For these events, positive linkages with SSTs in the northeast Indian Ocean basin develop from the end of the preceding austral summer until the events peak, while an opposite relationship prevails after the events. Connections with a SID-type SST gradient across the basin are particularly significant from the end of the preceding boreal summer until the end of austral summer the year of the events.

These above relationships prevail according to variable phasings: are they artifacts of the statistical methods for the short period considered or do they reflect any seasonality in underlying mechanisms? Further investigation is needed to validate these findings. Moreover, relations with tropical modes of variability in each basin region were not considered, and this would be of substantial interest. Extending this approach to a longer period (such as 50 years, possible with ERA-40 dataset for instance) could provide relevant information regarding the robustness of some of the climatic signals identified here. Furthermore the potential relationships between ocean-atmosphere variability in the Indian/South Atlantic Oceans have not been considered in regards of the variability in zonal moisture transport along the west/east coast respectively, and this should be addressed. Regarding links with the neighbouring oceans, the use of other data related to the position of the thermocline, the surface heat budget and the heat content of the upper ocean is necessary, while atmospheric model simulations could help emphasize overlying atmospheric processes and their relation to potential conditioning in the neighbouring oceanic

basins. Such considerations could help identify potential processes and indicators in regards to southern African climate predictability.

(c) Substantial coherences are identified between land hydrology estimates and some rainfall dataset but more work needs to be done in order to characterize their relationships in more detail. NDVI shows particularly good results over semi-arid regions while soil moisture seems to reflect well the importance of runoff for regions located on the slope of escarpments; to the east of the Great Rift valley and over regions surrounding the Bie plateau for instance. The use of coarser AVHRR NDVI datasets such as in *Martiny et al.* (2005, 2006) have led to particularly interesting results regarding long-term memory effects over semi-arid regions of Africa, and would probably help to draw a better and more refined picture of dynamics involved in the hydrological cycle. Similarly, the use of other soil moisture dataset implemented regarding for example orography adjustment, non-homogeneous observations, tuning methods (more relevant to environmental conditions prevailing over the subcontinent), could provide better elements of description.

Moreover, recent studies have shown that vegetation as well as soil moisture could be instrumental in modulating the water cycle. For instance, numerous climatic and anthropogenic factors are causing desertification, but one of the major positive feedback mechanisms at play is between precipitation and vegetation (*Obasi, 1999*), which was not considered in this study. In this feedback, a reduction of the vegetal cover leads to reduced evaporation inducing a reduction in precipitation, thus initiating a positive feedback cycle, and this needs to be considered in the context of the present findings. Recent studies (*Todd and Washington, 2003; Cook et al., 2006*) have also emphasized a negative feedback between soil moisture and atmosphere over southern African regions which was not taken into account in the present investigation. Land-surface conditions and their interactions with convection mechanisms definitely need to be further examined in regards to the results presented here. In this framework, another aspect that was not integrated is the contribution from large water bodies such as Lake Victoria, and the Congo river

which are of crucial importance for the water budget of the surrounding regions. The use of related time-series could be of particular interest. In conclusion, this study has considered mainly the atmospheric part of the hydrological cycle, and one next topic would be to examine the possible mitigation/aggravation of the effect of precipitation through land hydrology.

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Appendix A

List of consensual ENSO years

University of Cape Town

Austral Summer	WRCC	CDC	CPC	MEI	Consensus
1979-1980			W-	W-	
1980-1981					
1981-1982					
1982-1983	W+	W	W+	W+	Strong El Niño
1983-1984			C-		
1984-1985			C-	C-	
1985-1986					
1986-1987			W	W	
1987-1988	W+	W-	W	W-	El Niño
1988-1989	C+	C-	C+	C	Strong La Niña
1989-1990					
1990-1991			W+		
1991-1992	W	W	W+	W+	Strong El Niño
1992-1993	W		W+	W-	El Niño
1993-1994	W+		W		
1994-1995	W+		W	W-	El Niño
1995-1996			C-	C-	
1996-1997					
1997-1998	W+	W	W+	W+	Strong El Niño
1998-1999	C+		C	C-	La Niña
1999-2000			C	C	
2000-2001	C	C	C-	C-	La Niña
2001-2002					
2002-2003	W	W	W	W	El Niño

Table A.1: List of consensual ENSO years (after *Fauchereau*, 2004). W/C stand for warm/cold events respectively.

Appendix B

Description of different rainfall estimates

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CHARM The Collaborative Historical African Rainfall Model dataset has been developed to overcome the above limitations embedded in rainfall estimates over the subcontinent. The two key sources of data for CHARM are the National Center for Environmental Prediction (NCEP) reanalysis time-series and gridded station data. The daily estimated precipitation fields from the reanalysis are smoothed with a specialized spatial filter. This generates a set of 'synoptic' rainfall fields at a resolution of 0.1 degree. The gridded monthly precipitation fields produced by Cort Willmott (*Willmott et al.*, 1994) are then used to provide a monthly bias correction of the daily rainfall fields. The resulting dataset offers 0.1 degree/daily precipitation fields for all of Africa for the period 1961-1996. Further details and validations of CHARM estimates can be found in *Funk et al.* (2003). It is worth notice that this dataset is currently in use for rainfall monitoring in the Famine Early Warning Systems Network (FEWS NET) part of the U.S. Aid program on food security in sub-saharan Africa.

GPCC The GPCC surface precipitation data set is derived from the monthly precipitation totals based on conventional surface raingauge measurements. The GPCC collects monthly precipitation totals received from CLIMAT and SYNOP reports via the World Weather Watch GTS (Global Telecommunication System) of the World Meteorological Organization (WMO). The GPCC also acquires monthly precipitation data from international/national meteorological and hydrological services/institutions. An interim database of about 6700 meteorological stations is defined. Surface rain-gauge based monthly precipitation data from these stations are analyzed over land areas and a gridded dataset is created, using a spatial objective analysis method. This dataset is comprised of monthly gridded area-mean rainfall totals for the period January 1986 to March 1999 on a 1 by 1 degree global grid. For more details the reader is invited to refer to (*Rudolf*, 1996) amongst others.

CAMS-OPI The NOAA NCEP climate anomaly monitoring system, outgoing longwave radiation precipitation index (CAMS-OPI) dataset is intended to be used

for real-time climate monitoring. The data set is produced from rain gauge data for land (from the GTS), and OPI estimates for land points without observation as well as over the oceans. It provides rainfall estimates at a 2.5 degree resolution for 1979 onwards. For more details the reader is invited to refer to *Janowiak and Xie (1999)*.

GPCP The Global Precipitation Climatology Project (GPCP) version 2 combines the precipitation information available from several sources into a final merged product, taking advantage of the strengths of each data type. The microwave estimates are based on Special Sensor Microwave/Imager (SSM/I) data from the Defense Meteorological Satellite Program (DMSP, United States) satellites that fly in sun-synchronous low-earth orbits. The infrared (IR) precipitation estimates are computed primarily from geostationary satellites (United States, Europe, Japan), and secondarily from polar-orbiting satellites (United States). Additional low-Earth orbit estimates include the Atmospheric Infrared Sounder (AIRS data from the NASA Aqua, and Television Infrared Observation Satellite Program (TIROS) Operational Vertical Sounder (TOVS) and Outgoing Longwave Radiation Precipitation Index (OPI) data from the NOAA series satellites. The gauge data are assembled and analyzed by the Global Precipitation Climatology Centre (GPCC) of the Deutscher Wetterdienst and by the Climate Prediction Center of NOAA. The resulting GPCP Version 2 Combined Precipitation Data Set covers the period January 1979 through present with rainfall estimates at a 2.5 degrees resolution. This data set is the successor to the GPCP Version 1 Combination. The Version 2 combination includes precipitation estimates from TOVS and AIRS, thus permitting filling data voids at high latitudes that occurred in Version 1. More details about this dataset can be found in *Adler et al. (2003)*.

CMAF The CPC Merged Analysis of Precipitation (CMAP) dataset merges satellite and rain gauge data from a number of satellite sources and rain gauge sources. The CMAP dataset has also an option with models, using precipitation from Numerical Weather Prediction (NWP) models. The CMAP dataset provides

rainfall estimates at a 2.5 degrees resolution from January 1979 to present. Details on the component datasets as well as the method used to merge these data are provided by *Xie and Arkin* (1996, 1997).

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Appendix C

On interannual summer rainfall variability over southern Africa

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Given the importance of vegetation cover but also subsurface water supply to the water cycle, a comparison of mean annual rainfall (from the Collective Historical African Rainfall Model dataset, but similar results are found using CRU data), annually integrated NDVI and mean annual monthly soil moisture (more details about the datasets can be found in Chapter 2) is presented in the following for the November-April period during 6 years, from 1990-1991 to 1995-1996.

Rainfall variability over southern Africa is generally characterized by a dipole pattern, with large positive/negative loadings over subtropical regions south of 15°S and eastern Africa respectively (Mutai *et al.*, 1998; Jury, 1992, 1997). Consequently, the domain chosen for comparison purposes has been divided between tropical and subtropical areas of southern Africa.

In order to have an idea of the validity of the results presented in the following, the 1974-2003 OFDA/CRED International Disaster Database (Emergency Disaster Database online at www.em-dat.net) from the University Catholique de Louvain, Brussels (Belgium) has been used, and a list of disasters extracted from this database for the 1990-1996 period is presented in Appendix D.

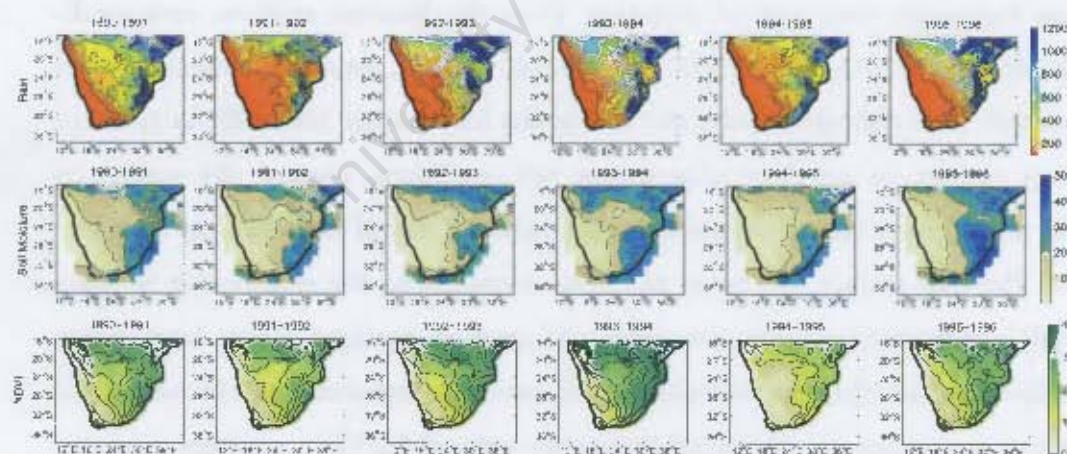


Figure C.1: Mean summer rainfall (top, in mm), monthly soil moisture (middle, in mm/mth) and integrated NDVI (bottom) during the summer period (November-April) of selected years over subtropical areas of southern Africa.

Figure C.1 shows November-April means over southern African subtropics from 1990-1991 to 1995-1996. Wettest conditions are found during the 1993-1994 summer, and correspond with flood conditions during this year (see Appendix B), in particular over Kwazulu-Natal (from October). On the other hand, 1994-1995 appears to exhibit the driest conditions which coincide with droughts that occurred in Lesotho (from April) and South Africa's northern provinces. The temporal extent of such a dry period cannot be diagnosed from these maps, but results for summer 1995-1996 suggest that it might affect the local hydrological budget. Mitigating conditions are noticeable in land hydrology data the following summer season (1995-1996) which does not appear to deviate much from rainfall means. In addition, large precipitated amounts are noticeable during 1995-1996 summer over southeastern regions of South Africa, where integrated NDVI and monthly soil moisture also exhibit higher values, reflecting the flood that occurred from December to February. Summer 1991-1992 shows the driest conditions, agreeing with droughts occurring in Botswana, Swaziland, Namibia and South Africa. Low values are also found in all fields the following summer (1992-1993), suggesting a potential inertia in these dry conditions within vegetation cover and soil moisture. From the disasters archives, such conditions correspond with the drought over Lesotho (April), Namibia (November) and Swaziland. It is difficult to interpret the results for summer 1990-1991 in land hydrology as they do not show anomalously wet conditions, instead, dry conditions which prevail since the previous summer (not shown) should be taken into consideration. Previous studies (*Farrar et al.*, 1994) investigating the relationship between rainfall and vegetation cover over semi-arid areas of southern Africa have shown that discrepancies could be attributed to soil nature in terms of both the rate of soil moisture generation per unit rainfall and the ratio of NDVI to moisture amongst other factors.

Similarly, Figure C.2 compares means over tropical areas of southern Africa, for the period from November-April from 1990-1991 to 1995-1996. It is worth noting that soil moisture data is output from models and thus can be affected by biases

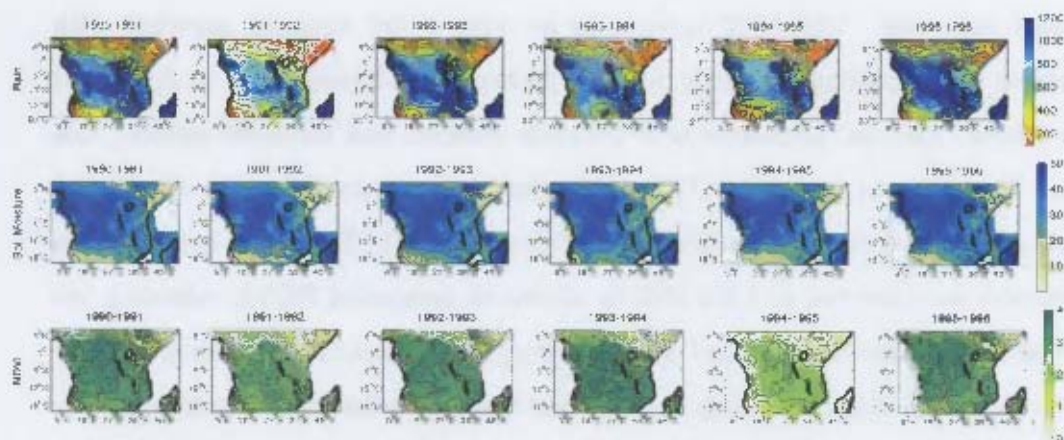


Figure C.2: Mean summer rainfall (top, in mm), monthly soil moisture (middle, in mm/mth) and integrated NDVI (bottom) during the summer period (November-April) of selected years over tropical areas of southern Africa.

contained in the input precipitation in particular over tropical regions. NDVI there is also expected to saturate over a certain threshold which is overpassed in the case of very dense vegetation cover as found over the wet Congo basin for instance. Nevertheless the annually integrated NDVI approach is well suited to reflect variations from one summer to the consecutive one. In order to take into account these saturation effects, the scales used for integrated NDVI does not consider values above 4, which correspond to the 0.7 individual NDVI value characteristic of dense vegetation. The results obtained for both fields reflect the above limitations which are discussed in more details in Chapter 2.

The years identified as anomalously wet (1990-1991, 1992-1993, 1993-1994 and 1995-1996) and dry (1991-1992 and 1993-1994) summers show the same characteristics in the integrated NDVI. Soil moisture does not exhibit much variability, in particular over the Congo basin. Nevertheless, the southward incursion of the 20 mm/mth contour in 1993-1994 confirms above normal soil moisture during this year. This might be the result of two consecutive wet summers (1992-1993 and 1993-1994) which could indicate a potential inertia within land hydrology data, as noticed

from maximums in integrated NDVI, and coinciding with floods over Tanzania in January this year. 1991-1992 appears to be quite a dry summer, agreeing with extensive dry conditions found in Angola, Malawi, and Mozambique in April and over Kenya, Zambia, Zimbabwe and Tanzania (central and northern regions) this year. The following summer (1992-1993) displays wet conditions in all dataset which coincided with floods in Tanzania (February-May). The driest conditions in vegetation are observed in 1994-1995 as shown in integrated NDVI, reflecting the drought in Zimbabwe, Kenya and Malawi during this period. The following summer (1995-1996) thus does exhibit greenness within the vegetation cover but less than in 1993-1994, which once more suggests that results in NDVI must be considered with the vegetation "memory" and feedback linked with prevailing historical conditions. Still these results agree with floods that occurred especially over Malawi (December) and Tanzania (March) this year.

Appendix D

Disasters list extracted from the
OFDA/CRED database

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Date	Country and Location	Disaster Type	Impacts
Dec 1990	Angola	Drought	-
1990-1991	Swaziland	Drought	35 000 affected
1991	South Africa	Drought	-
Apr 1992	Angola	Drought	-
Apr 1992	Namibia	Drought	250 000 affected
Apr 1992	Malawi	Drought	5 700 000 affected
Apr 1992	Mozambique	Drought	3 300 000 affected
Apr 1992	Zimbabwe	Drought	4 600 000 affected
1992	Kenya	Drought	2 700 000 affected
1992	Tanzania	Drought	800 000 affected
1992	Zambia	Drought	-
1992	Botswana	Drought	100 000 affected
Nov 1992	Namibia	Drought	-
1993	Swaziland	Drought	150 000 affected
1993	Malawi	Drought	7 000 000 affected
1993	Zimbabwe	Drought	5 000 000 affected
Feb 1993	Tanzania	Flood	54 killed, 30 injured, 201 513 affected
Apr 1993	Lesotho	Drought	-
May 1993	Tanzania	Flood	280 homeless
Oct 1993	South Africa	Flood	12 killed, 15 000 homeless
Jan 1994	Tanzania	Flood	31 killed, 7 000 homeless
Feb 1994	Malawi	Drought	3 000 000 affected
1994	Swaziland	Drought	45 000 affected
1994	Zimbabwe	Drought	5 000 000 affected

Date	Country and Location	Disaster Type	Impacts
Nov 1994	Somalia	Flood	100 killed
Dec 1994	Congo	Flood	16 500 affected
Jan 1995	Kenya	Drought	1 200 000 affected
Jan 1995	Malawi	Flood	1 killed, 1 000 homeless
Feb 1995	Botswana	Flood	20 killed, 3 500 affected
Mar 1995	Tanzania	Flood	1 850 affected
Apr 1995	Lesotho	Drought	331 500 affected
May 1995	Tanzania	Flood	3 killed, 20 000 homeless
1995	Zimbabwe	Drought	5 000 000 affected
Dec 1995	South Africa	Flood	207 killed, 4 500 affected
Dec 1995	Malawi	Flood	300 homeless
Jan 1996	South Africa	Flood	7 killed, 500 homeless
Feb 1996	South Africa	Flood	27 killed, 7 000 affected
Mar 1996	Tanzania	Drought	3 000 000 affected

Table D.1: List of disasters over southern Africa for the 1990-1996 period from the OFDA-CRED International Disaster Database, Universite Catholique de Louvain, Belgium.

Appendix E

On air moisture content

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As mention in Chapter 2, there are multiple measures of air moisture content. The mixing ratio for instance, represents the amount of water vapour in a given volume,

$$x = \frac{m_w}{m_d} \quad (\text{E.1})$$

where, m_w, m_d are respectively the mass of water vapour and the mass of dry air. Using vapour pressure measurements, Equation (E.1) can be expressed as,

$$x = \frac{0,622.e}{p} \quad (\text{E.2})$$

where, e is the vapour pressure and p the atmospheric pressure. During evaporation from a plane surface of water to the overlying air, an equilibrium is reached at a particular temperature (dew-point temperature) when no more exchange takes place between air and water. The air is said to be saturated and exerts a so-called saturation vapour pressure e_s , function of temperature. The saturated mixing ratio can then be defined as follow,

$$x_s = \frac{0,622.e_s}{p} \quad (\text{E.3})$$

In the absence of measurements, the vapour pressure e can be derived from dew-point temperature as described in previous studies (*McGee, 1971; D'Abreton, 1992; Fauchereau, 2004*),

$$e = e_s(T_d) = 6,11.10^{\left(\frac{7,5.T_d}{237,3+T_d}\right)} \quad (\text{E.4})$$

The specific humidity q is defined as the mass of water vapour per mass of moist air,

$$q = \frac{m_w}{m_a} \quad (\text{E.5})$$

where, m_w is the mass of water vapour and m_a the mass of moist air. It can also be calculated as,

$$q = \frac{0,622.e}{p - 0,378.e} \quad (\text{E.6})$$

The relative humidity RH is defined as the ratio of the observed mixing ratio to that which would saturate the air at the same temperature (*Preston-Whyte and Tyson, 1988*),

$$RH = \frac{x}{x_s} = \frac{e}{e_s} \quad (\text{E.7})$$