

Southern African synoptic climate dynamics and future change

Tropical Lows and the Congo Air Boundary

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A thesis submitted for the degree of
Doctor of Philosophy

Trinity Term 2019

Abstract

Southern Africa's development and economy is dependent on a rainfall resource that is projected to decline in coming decades as a result of anthropogenic climate change. This rainfall decline has been attributed to a half month delay in the onset of the wet season by the end of the 21st century based on RCP8.5 projections, shortening the growing season and reducing the agricultural viability of the region. Yet many aspects of the basic climatological state of the southern African atmosphere remain chronically understudied. Rainfall onset and regional rainfall climate depend on the behaviour of day-to-day weather systems along the southern margin of the African rain belt. However, the dynamics of these systems are not well understood. This thesis addresses this problem by characterising the frequencies, spatial distributions, dynamics, interannual variability and future change of two classes of weather systems: tropical lows and the Congo Air Boundary (CAB).

A vortex tracking algorithm is optimised to detect tropical lows in sub-daily reanalysis data and climate model output. It is found that tropical lows aggregate to form the synoptic expression of the summer Angola Low. Tropical lows are shown to exhibit a unique anchoring behaviour in Angola due to interactions with nocturnal low-level jets associated with differential surface heating. 70% of rainfall in south-east Angola and 31% of summer rainfall across the southern African tropical margin is attributed to tropical lows. Meanwhile, the majority of interannual rainfall variability in Angola, northern Namibia and western Zambia derives largely from tropical lows. Climate model projections indicate that tropical lows will become 15% less frequent by the end of the 21st century. Future change in the spatial distribution of tropical lows contributes to model disagreement over the spatial pattern of rainfall change in the summer season.

Computational edge detection is applied to identify surface drylines and convergence lines associated with the CAB, a forgotten feature of the southern African climatology. The CAB is shown to be a sharp discontinuity which marks the southern edge of the African rain belt, and is detected on at least 80% of days from August to October. Its seasonal southward progression over southern Africa is presented, and a close dynamic link between the CAB and the Angola Heat

Low is demonstrated. Seasonal rainfall variability is closely linked to the CAB, since tropical rainfall does not occur to the south of the CAB, with CAB variability explaining at least 55% of local monthly rainfall variability. The projected delayed wet season onset is shown to be related to a projected 0.7° CAB latitude increase and a 13% CAB frequency increase from October to December, with inter-model CAB variability explaining 46% of inter-model variance in projected rainfall change.

Acknowledgements

Thanks must go firstly to my supervisor, Richard Washington. Thank you for taking a chance on an out-of-the-blue email and welcoming me to your research group. Over the course of this degree, I have benefited immensely from your guidance, vision and enthusiasm. You have provided me with unique opportunities, through collaborations, travel and fieldwork, that have been the highlights of my PhD. These experiences have given me perspectives on climate science and academia that I would not have received anywhere else.

I must thank Keith Ayotte for teaching me how to think like a researcher. Your pearls of wisdom are hidden just beneath the surface throughout this document. I've tried to inflict them on my peers many a time since leaving Windlab, but they never sound quite the same.

The UMFULA climate science team have been absolute pleasure to work with. Thank you for being a supportive, cohesive and fun research group. Having a sympathetic research group to share ideas with has been invaluable in building my confidence as a scientist. Thanks for many patient explanations of your deep collective understanding of the African climate system. Special thanks must go particularly to Neil Hart, for your mentoring and for sharing your passion for dynamics, and to Callum Munday, for collaboration and for proof-reading parts of this thesis. More broadly, I must thank the UMFULA social science and impact modelling teams for helping me to understand how climate science feeds into your work. Thank you for allowing me to tag along to your annual meetings. Special thanks go to Estelle Rouhaud, for herding cats.

I am grateful to Kevin Hodges for his generous help in setting up TRACK; to Kerry Cook and Ned Vizzy, for welcoming me as a visiting student in Austin for a week and for several illuminating discussions on southern African dynamics; and to Glenn Carver, for support using the OpenIFS model, which unfortunately didn't make it into this thesis. This thesis is based on a series of meteorological and climate datasets and I am indebted to all those around the world who have

worked for many years to develop these, as described at the end of Chapters 3 - 6.

This project was supported by an Origin Foundation John Monash Scholarship. My most sincere thanks go to all who work with the John Monash Foundation for their efforts in building a community of Australian scholars around the world and in Oxford. The Monash family has been both a source of inspiration and a welcome connection to home throughout my degree.

The Climate Research Lab within the School of Geography is a special place, and I've been very lucky to call it my home for the last three years. Thanks must go to Karsten and Josie for all their efforts in making the CRL more than just a workplace.

Mum and Dad deserve a particularly special mention. I am deeply grateful for the work ethic you instilled in me through your example, and for the faith you've always had in us. Thank you for being so understanding as I've moved away for study, first across the country and then across the world.

To James: thank you for moving across the world with me on this fantastic adventure and continually coaxing plans out of me even when I thought I didn't know where this thesis was going. Thanks for dragging me out of countless ruts, for proof reading every line of this document, and for continually encouraging me to think about the broader significance of this work. Your support has made this PhD possible, and I'm looking forward to our next adventure.

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CHAPTER 1

Introduction

1.1. Motivation

The prosperity and economic growth of southern Africa is inextricably linked to climate. The region is home to nearly 350 million people, and growing at a rate of 2.5% per year (SADC, 2018). According to climate models, much of this region is projected to exhibit one of the few robust land-based drying trends. This drying trend is associated with a delay of wet season onset. The rainy season is projected to commence half a month later in 2080 - 2100 under a business-as-usual emissions scenario compared to the present day (Dunning et al., 2018). Presently, the wet season begins between October - December (OND) and lasts until March, with a large degree of spatial variation in the onset dates between sub-regions (Dunning et al., 2016). The core of the rainy season occurs between December and February (DJF). For a region dependent on water resources for growth, the risk to the continued development of southern Africa cannot be underestimated.

Climate risk is further compounded by the uncertainty inherent in future projections. Uncertainty is particularly acute over southern Africa because the many components of southern Africa's basic atmospheric state are poorly understood. Understanding of the fundamental regional climate dynamics must be improved before the future projections may be used with confidence to guide adaptation to a drier climate.

Southern Africa is strongly reliant on its water resource for domestic, industrial and agricultural use (SADC, 2016). Much of agriculture is rain-fed, while the irrigated remainder competes with hydropower for access to river flow. Nearly all the electricity generation in Zambia, Malawi and Lesotho is sourced from hydropower (Conway et al., 2015). The tourism industries of Victoria Falls, Lake Malawi and the Okavango Delta all rely on the health of the river basins. Steep orography ensures that no major rivers flow south across the 9°S latitude band. As a consequence, all freshwater south of this line must be sourced from local rainfall. The southern African economy and future prosperity is therefore highly sensitive to the annual rainfall total.

Although the rainfall climatology is of vital importance to the region, the basic dynamics of the region’s climatological mean state is still an area of active research (e.g. Nicholson (2018)). This is especially true of synoptic weather systems, which form the fundamental building blocks of the climatology. The local atmospheric dynamics, which ultimately play out on synoptic and mesoscale timescales, control the characteristics of the climatology and are critical for realistic simulation of the regional climate system. Therefore, it is essential that they are well understood. While variability of the climate on timescales from interannual through to decadal has been well studied (e.g. Mason and Jury (1997); Reason et al. (2006a); Morioka et al. (2015); Pohl et al. (2018)), the synoptic timescale has been neglected in southern Africa, particularly in the interior along the margin of the tropical rain belt. The exception here are synoptic tropical temperate troughs (TTTs), whose synoptic characteristics have been carefully documented, for example by Hart et al. (2012, 2013).

Advances in understanding the basic state of the climate system on synoptic timescales are necessary to reduce climate related socio-economic risk. The dynamics of both interannual variability and future change depend on the response of rainfall sourced from synoptic weather systems to external forcing, be it remote sea surface temperatures or temperature increases throughout the lower troposphere. Changes induced in synoptic weather systems will then carry through to the rainfall climatology. While the dynamics that drive these weather systems remain a mystery, and while our understanding of their simulation in climate models remains poor, seasonal forecasts and climate projections will continue to be subject to preventable uncertainty. Deeper understanding of the synoptic processes is vital for elucidating the context of climate projections on a range of timescales.

The southern African future drying signal is unique in the sense that it is land-based, and occurs primarily equatorward of the subtropical ridge (Scheff and Frierson, 2012*a,b*). The spatial extent of the region with a consistent drying signal across climate models is one of the two largest in the inhabited world. Projections of future change are subject to a high degree of uncertainty, not least because climate models struggle to represent the mean state of the southern African atmosphere (Munday and Washington, 2017, 2018). For these reasons, unravelling the dynamics of the southern African climate poses a worthwhile contribution to climate science in its own right.

This thesis will draw on the meteorology of southern Africa to characterise the synoptic expression of the southern African climate. By doing so, a set of weather systems that may be well simulated by climate models will be studied. The weather systems considered will be tropical lows and drylines/convergence

lines associated with the Congo Air Boundary (CAB). These systems have been chosen based on the following criteria:

- (1) Collocation with the region of maximum projected rainfall decline;
- (2) Association with rainfall on synoptic timescales and possessing a demonstrable statistical relationship with interannual rainfall variability;
- (3) Possessing large enough length scales so as to be resolvable in typical climate models and reanalysis products; and
- (4) Being relatively understudied to date, such that no adequate feature tracking methodology has been previously applied to identify the feature on daily timescales to the author's knowledge.

The overarching goal of this work is to express the climatology, dynamics and future change of the southern African tropical margin from the viewpoint of synoptic weather systems. In pursuit of this goal, this thesis will present a climatology of the spatial and temporal frequencies, key characteristics, interannual variability and influence on interannual rainfall variability of two classes of weather systems, based on observed and reanalysed datasets. This thesis then examines the historical representation of these weather systems and their projected future change in CMIP5 models.

The remainder of the introduction proceeds as follows. First, the rationale for focussing on synoptic weather systems to constrain understanding of rainfall climate is presented in Section 1.2.1. This section emphasises the distinction between weather and climate, and the symbiotic relationship between the two concepts. In Section 1.2.2, the climate models that will be considered in this thesis are introduced, and the synoptic process-based approach is placed within the context of these models. Third, Section 1.2.3 precisely defines the region of

interest within southern Africa. The thesis then proceeds with a literature review that introduces current understanding of the synoptic climate of southern Africa. Section 1.3 focuses first on the observed climatological basic state, second on the classes of synoptic weather systems that aggregate to this mean state, and third on observed interannual variability. When considering synoptic systems, feature-tracking methodologies that have been applied to similar features across the world are also outlined. Section 1.4 reviews the ability of climate models to simulate southern African climate and projections of future change in this region. Section 1.5 describes the overarching goals and aims of the thesis and further justifies the focus on tropical lows and the CAB. The means by which the four papers submitted in association with this thesis will achieve these aims is outlined in the final section.

1.2. Rationale

This section provides a rationale for the key approaches taken in this thesis. The focus on synoptic events is justified in Section 1.2.1, the use of climate models and methods of climate model evaluation is explained in Section 1.2.2, and the primary study area is defined and justified in Section 1.2.3.

1.2.1. Synoptic constraints

When studying the rainfall climate of a given location, rainfall is often considered as a climatological variable, averaged over seasonal intervals (e.g. Washington and Preston (2006), Aloysius et al. (2016)). There are advantages to this approach, since monthly, seasonal and interannual rainfall is generally smoother and more predictable than daily rainfall. Furthermore, this is a timescale on which climate information, such as seasonal forecasts, is often distributed (e.g. Scaife et al. (2016)). On seasonal timescales, global mean rainfall is constrained

by the global energy budget, since the latent heat of condensation is the primary means by which the mid- and upper- troposphere warms (Vallis, 2017).

However, seasonal averages do not expose the rainfall mechanisms which are responsible for the climatology. Neither do seasonal means capture the temporal pattern of rainfall, which, outside the deep tropics, often occurs in intense bursts separated by longer periods of dry conditions. Pendergrass and Knutti (2018) showed that at gauge stations across the world, half the annual mean rainfall would fall in the 12 wettest days, on average. This low value reflects the skewed temporal distribution of daily rainfall.

The temporally uneven distribution of rainfall arises because rainfall may only occur as isolated convection, as a part of infrequent synoptic events during which static instability is present, or if the large-scale flow contains forcing large enough to allow static stability to be overcome (Emanuel, 1994). The occurrence and characteristics of such synoptic events therefore places additional constraints on local rainfall, which may be used to characterise the rainfall climatology and future projections. This approach has been applied successfully in the southern hemisphere at high latitudes, where extra-tropical cyclones are responsible for much of the local precipitation, and where extra-tropical storm track shifts have been robustly linked to future rainfall change (Frederiksen and Frederiksen, 2007).

The temporal distribution of rainfall in the tropics is generally steadier in the tropics than the extra-tropics, where synoptic eddy fluxes are an order of magnitude larger. However, the synoptic variability is nevertheless important at lower latitudes, in particular in southern Africa where the rainy season can at times consist of only a few significant rain events (Cook et al., 2004; Blamey and Reason, 2009). Based on the methodology of Pendergrass and Knutti (2018),

the median across southern Africa of the number of days in which half the annual rain occurs is 12.5 days. This value is based on the TRMM satellite-based gridded rainfall dataset at its native resolution, which Pendergrass and Knutti (2018) showed matched well with stations at adjacent grid cells. This implies that intermittency in southern Africa is only slightly lower than in the gauge station mean. Therefore, the characteristics of the synoptic weather systems that deliver precipitation in southern Africa determine the rainfall climatology. By studying the dynamics of synoptic weather systems, the spatial distribution of rainfall may be explained.

Because synoptic weather events control the delivery of precipitation, it is important to study the rainfall climate from the viewpoint of synoptic components of the local atmospheric circulation. To do this, individual weather systems must be identified within observed, reanalysed and modelled atmospheric data. Automated algorithms, bespoke designed and/or optimised for each class of weather system, have proven successful at tracking features through time and space (e.g. Hodges (1999), Thorncroft and Hodges (2001), Hart et al. (2012), Hurley and Boos (2015) and Hunt et al. (2016)). Once identification and tracking is completed, spatial distributions, frequencies and rainfall contributions can be easily assessed. In addition, the dynamics of these weather systems may be examined using techniques such as centred composites and budget analysis.

In light of the importance of synoptic timescale for rainfall, this thesis examines synoptic events in southern Africa. Firstly it identifies tropical lows and the CAB as key understudied classes of synoptic systems that influence rainfall over southern Africa, and investigates their climatologies in observational datasets using feature-tracking algorithms. The degree to which interannual rainfall variability is driven by rainfall associated with these systems is addressed. This thesis

then attempts to express rainfall changes projected in climate models as changes in the spatial distributions and characteristics of these synoptic events that enable convection. This final step follows the principles of process-based climate model evaluation.

1.2.2. Climate models

General circulation models (GCMs) are a primary tool available for predicting the spatial distribution of future climate change (Taylor et al., 2012). GCMs are mathematical models that solve discretised versions of the Navier-Stokes equations for the atmosphere and the oceans. Typically, they are discretised to resolutions that do not enable them to resolve processes such as atmospheric convection and oceanic eddies, and so these processes which must be parametrised (Neelin, 2010; Griffies, 2004). Parametrisation schemes introduce a considerable degree of error into the model's skill at simulating the climate system. This error contributes to uncertainty in the problem of climate prediction, and is a key source of divergence in future projections generated by different models. However, there is a general consensus that useful information may still be drawn from GCMs despite their limitations, particularly if an ensemble of different models is considered (Stocker et al., 2013). Therefore, GCMs will be used to study the future change of tropical lows and drylines in the final stage of this thesis.

The GCMs studied in this thesis were drawn from the coupled model inter-comparison project (CMIP), a framework in which scores of GCMs may be compared. CMIP5 consists of a set of GCMs developed around the world that have been forced by an identical set of forcing scenarios. As the ability of GCMs improves over time, the CMIP project repeats every 5 - 10 years with state of the art models. At the time of writing, CMIP has just completed its 6th round

(Eyring et al., 2016), however, daily model output from the 6th round was not available from all models. Therefore, the 5th round, CMIP5, has been used instead. In CMIP5, model scenarios included historical simulations intended to represent the climate of the 20th century and a set of future climate scenarios simulating the 21st century and beyond with varying projections of greenhouse gas forcing, along with some other scenarios (Taylor et al., 2012). These scenarios include RCP2.6, which assumes that strong global action will be taken to mitigate emissions; RCP8.5, which represents a future in which no action is taken to curb emissions, and RCP4.5, which lies between these two poles. The southern African drying trend is present in all three of these scenarios, however it is most prominent in RCP8.5 (Stocker et al., 2013). Therefore this thesis focusses on comparing change between the projected end of 21st century climate under the emissions scenario labelled RCP8.5 and the historical 20th century simulation.

While CMIP5 models were primarily developed for examination at climatological timescales, each model simulates its own version of the Earth's weather (Taylor et al., 2011). In order to achieve numerical stability, all CMIP5 GCMs are run with a time-step between one minute and one hour, which allows for the temporal evolution of synoptic scale weather systems to be properly resolved. Climate may be defined as the statistical distribution of weather, taken over a long enough period to properly characterise the variation of synoptic weather systems. Therefore, in order to properly resolve the climate system, a model must be broadly capable of producing the correct types of weather processes (Neelin, 2010). Indeed, modelling centres around the world have adopted an integrated approach, favouring the use of models which can be run in different modes for numerical weather prediction, seasonal forecasting and climate prediction (e.g. Martin et al. (2011)).

This close relationship between weather and climate, and the requirement that climate models must, to some degree, resolve weather systems, allows the synoptic constraints described in Section 1.2.1 to be examined in climate models. This technique is a part of a broader methodology known as process-based evaluation of climate models (e.g. James et al., 2015; Dieppois et al., 2015; Daron et al., 2019; Woollings et al., 2018). Process-based evaluation focusses on determining whether components of the Earth system that govern its dynamical and thermodynamical evolution are correctly simulated in climate models, rather than focussing only on model outputs that are directly relevant for impact studies. This approach may lead to insight into why a particular model bias may exist, and allows for a screening process to discount models that do not appropriately simulate important weather processes (James et al., 2018).

Meanwhile, projections in the changes of synoptic weather processes may be considered as a proxy for precipitation projections. This ‘process-based projection’ approach has value in cases where these synoptic processes are better resolved in historical simulations than regional precipitation. Linking precipitation projections to dynamical projected changes in synoptic processes enables conceptual mechanisms of future change to be developed. The development of these mechanisms reduces the extent to which climate models are relied on as ‘black boxes’ and also has the potential to lead to the generation of falsifiable hypotheses. In this way, the perceived uncertainty surrounding rainfall projections may be reduced.

The task that remains is one of translation: learning which features are of high relevance to rainfall and then developing methods to easily trace them in climatological datasets. Once this is achieved, a deeper understanding of the basic climatology may be realised. Questions relating to the climatological timescale,

such as whether these systems are well resolved in GCMs, how these weather systems respond to different phases of ENSO, and to increasing global temperature due to anthropogenic CO₂, can then be addressed.

1.2.3. Key study area

This thesis considers the meteorology, climate and climate change of the African continent south of the equator. The climate of this region stretches from equatorial rainforest at its northern edge, through humid subtropical and semi-arid/arid climates and finally to a warm Mediterranean climate at the southern tip near Cape Town, where winter a rainfall regime dominates (Beck et al., 2018).

The study area for this thesis is determined by taking the following considerations into account. Firstly, the study area must contain the overlapping regions within southern Africa where future rainfall decline is projected to be robust. Secondly, since a process-based approach is taken, it must avoid splitting the loci of any key climatological features of southern Africa. Thirdly, the approach of this thesis intends to focus on atmospheric features that have previously received little attention. Therefore, the core region of attention is the margin of the tropical rain belt. Less focus is given to the subtropics located further south, where South African climate scientists have already built a comprehensive literature base, particularly regarding TTTs (e.g. Harrison (1984), Todd and Washington (1999), Washington and Todd (1999), Vigaud et al. (2009), Hart et al. (2012) and Macron et al. (2014)).

Figure 1.1 shows the projected rainfall change in southern Africa across three different global warming scenarios, for which more detail is provided in sections 1.2.2 and 2.1.5. This figure indicates that in the scenario with the least warming

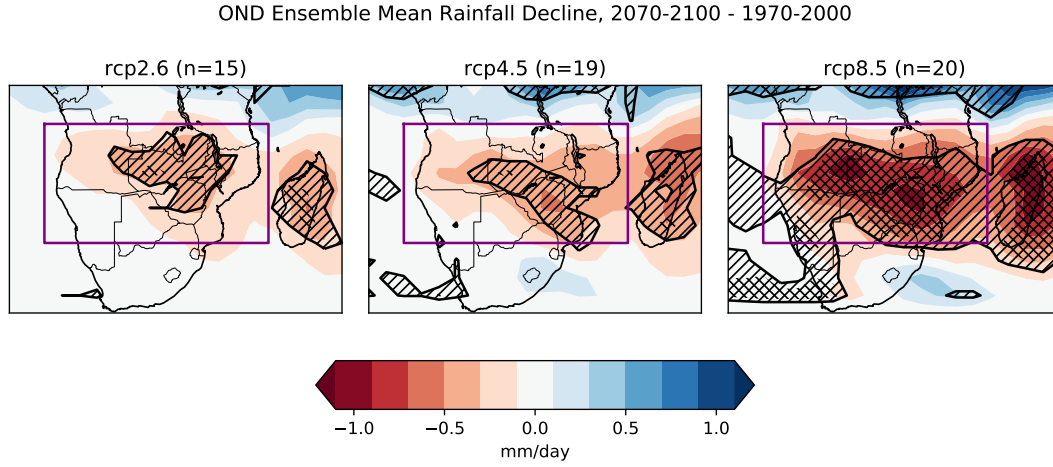


Figure 1.1. Region of interest with spatial patterns of future rainfall change under three RCP scenarios. Colours indicate the ensemble mean of projected rainfall change across a selection of 15-20 CMIP5 models (as per Table 6.1). Single (double) hatching indicates regions where 66% (80%) of the models feature a consistent rainfall decline or increase whose magnitude is larger than the model's 30-year internal variability. Hatching is only shown where the models with a significant change possess a consistent direction of change. The purple box indicates the primary area of interest for this theses.

(RCP2.6), 60% of models feature drying that is larger than their own internal variability over much of Zambia, Malawi and Zimbabwe. Under the scenario with the most warming (RCP8.5), this proportion increases to 80%, and the magnitude of drying doubles. Figure 1.1 indicates that the consistent projected rainfall decline is contained between latitudes of 10° - 24° S from October-December. Therefore, allowing for a buffer zone around this rainfall decline. the latitude band from 8° to 25° S is chosen as the focus region for this study. This region contains all of Zambia, Malawi, and Zimbabwe, large parts of Mozambique, Botswana and Angola, the northern half of Namibia and smaller proportions of the Democratic Republic of the Congo, Tanzania and South Africa.

1.3. Observed climate of southern Africa

The following sections form a literature review which outlines the present understanding of the observed climate of southern Africa. First, the major atmospheric features of the climatological mean state are presented. As much of the understanding of this mean state is derived from reanalysis products rather than direct observations, the extent to which reanalysis products agree in southern Africa is then reviewed. Third, this thesis investigates current literature that describes how this mean state disaggregates into a series of synoptic weather systems. While the literature examined is sound, gaps are identified regarding the structure and dynamics of these weather systems. Finally, the leading order drivers of interannual rainfall variability are considered.

1.3.1. Climatological mean state

Figure 1.2 displays key climatological atmospheric features of southern Africa in the spring and summer seasons. In the following section, the current literature, these features will be defined and literature concerning their dynamics will be discussed.

1.3.1.1. *Tropical rain belt*

The African tropical rain belt is a broad region of high humidity and convective precipitation which moves north and south across the African continent through the seasons (Dunning et al., 2016). It is analogous to the ITCZ over adjacent oceans, although use of this term over continental Africa has recently been questioned (Nicholson, 2018). In West Africa, the rain belt shifts north rapidly at the end of June (Sultan et al., 2003), forming the west African monsoon. In East and Central Africa, the September-November (SON) southward passage and the March-May (MAM) northward passage of the rain belt are referred to as the

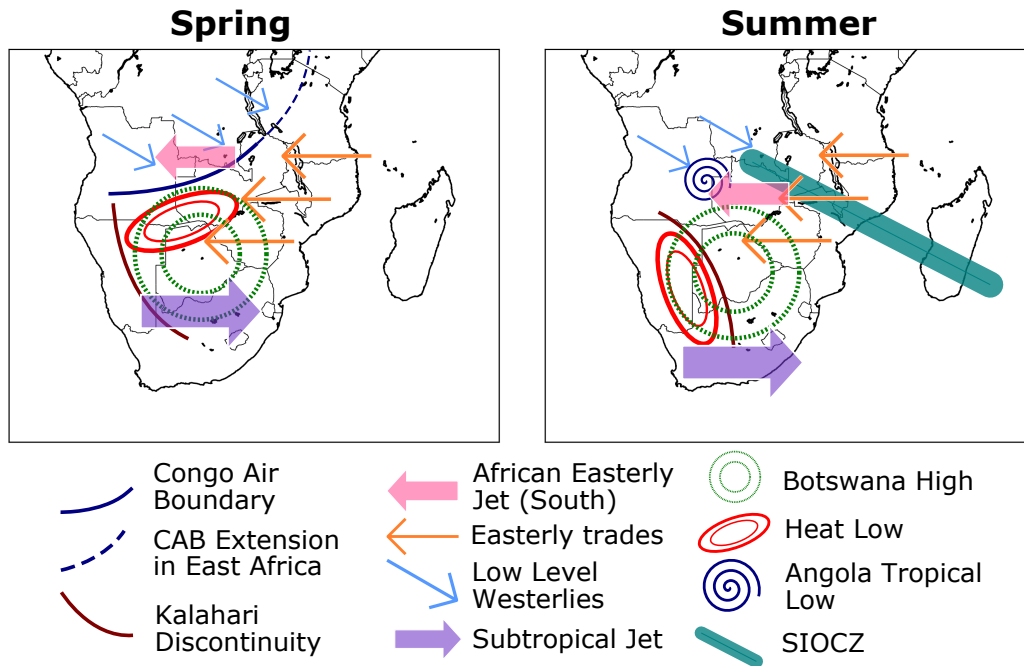


Figure 1.2. Schematic of key mean-state features of the southern African atmosphere during spring (OND) and summer (DJF). This figure has been adapted from Figure 5.13. The location of the SIOCZ has been added, and is based on Lazenby et al. (2016). The features shown in this figure are discussed in Section 1.3.1.

short rains and the long rains respectively. The rain belt shifts slowly towards southern Africa in the October-December (OND) season, reaches its south-most latitude in January and February, and retreats between March and May.

The southern African rainy season bears resemblances to monsoon systems in other regions of the world, however, whether it is a monsoon circulation itself depends on the definition applied. The criteria applied by Ramage (1971) require a reversal of winds by at least 120° between January and July and mean wind speeds exceeding 3 m/s in each season. Slingo (2002) demonstrates that these conditions are not met in southern Africa, as the wind speeds are too weak and do not change direction sufficiently. However, the tropical rain belt reaches further

south over Africa than it does over the oceans due to the temperature gradients associated with the land surface, which suggests that the processes driving the seasonal cycle of tropical rainfall may be qualitatively similar to those that occur in monsoon regions. The Monsoon Westerly Index, (Wang and Ding, 2008) which does not set a minimum wind speed threshold, does classify southern Africa as a monsoon region. However, this thesis takes the historical stance, following Tyson and Preston-Whyte (2000), that the southern African wet season is not a monsoon circulation.

Interestingly, the strong meridional temperature gradient is caused by contrasts between the warmer subtropical landmass with the relatively cool equatorial rainforest, rather than with equatorial oceans as in the standard monsoon regions of the world. Since surface temperature gradients affect the momentum budget through the generation of divergent low-level jets (Parker et al., 2005), it may be expected that these jets are instrumental in the dynamics of local synoptic weather systems. However, this has not yet been explicitly proven.

Despite being arguably the most important climatic feature of tropical Africa, there are many unanswered questions regarding the tropical rain belt (Nicholson, 2018), including questions relating to its southern edge. It is unknown whether the southern edge of the rain belt is a clear discontinuity, like the inter-tropical discontinuity at the northern edge, or whether it is a more gradually varying frontier. The degree to which convection along the tropical edge is organised has implications regarding the prevalence of convective aggregation in the tropics. While the seasonal shift of the rain belt is clearly in phase with the annual cycle, the precise dynamics of this oscillation are unclear. Furthermore, the degree to which subseasonal variability in the rate of the seasonal progression dominate rainfall

variability along its south-most extent is unknown. The characteristics and dynamics of the rain belt's southern margin determine the climatology, variability and change of rainfall in the tropical edge region. Therefore, it is important that these dynamics are better understood. This thesis will answer these questions in Chapter 5.

1.3.1.2. *Subtropical highs*

The subtropical highs are a series of high pressure cells located approximately 30° north and south of the equator. The highs may be defined as a set of closed contours in climatological mean-sea pressure over each subtropical ocean basin (Rodwell and Hoskins, 2001). They are associated with the subsiding branch of the Hadley Circulation (Marshall and Plumb, 2008).

The climatological highs of leading order importance to southern Africa are those over the adjacent oceans - the **Mascarene High** in the Indian Ocean, and the **Atlantic High** in the South Atlantic Ocean (Sun et al., 2017). The easterly geostrophic winds associated with the northern edge of the Mascarene High are the **Indian Ocean trade winds**, and advect moisture onto the southern African continent (Jury and Nkosi, 2000).

Since the south west Indian Ocean sea surface temperatures (SST) are much warmer than the south eastern Atlantic Ocean SSTs, there is much more moisture flux associated with the geostrophic inflow from the Mascarene High than the St. Helena High (Sun et al., 2017). However, under the influence of the land-sea temperature gradients, the Atlantic Ocean trade winds recurve near 5°S to become the primarily divergent **Low-Level Westerlies** (LLW), which flux moisture from the warm equatorial Atlantic Ocean into the Congo basin (Pokam et al., 2014).

1.3.1.3. *Angola Low*

The Angola Low is a cyclonic vortex that exists in the summer climatological mean wind and geopotential height fields in the lower troposphere, often studied at vertical pressure levels between 850 and 700 hPa. Centred near the border of Angola and Zambia, the Angola Low is co-located with a local spatial maximum in the January - March (JFM) precipitation climatology. As a climatological feature, it was first described by Mulenga (1998), who identified its key relationship with rainfall variability (further studied by Cook et al. (2004)). The vertical depth of the Angola Low varies across the summer season, starting as a shallow heat low from September through to November, and then transitioning to a deeper tropical low from December to March (Munday and Washington, 2017). The dynamics of the Angola Low on a synoptic timescale have not been well studied. It is not known, for example, whether the Angola Low is the result of one stationary system, or numerous systems that aggregate together to form the Angola Low in the climatological mean. The dynamics of the transition from heat low to tropical low are also not clear. Chapter 3 will describe the dynamics of the Angola Low. Current understanding of the interannual variability of the Angola Low will be described in section 1.3.4.2.

1.3.1.4. *Botswana High*

The Botswana High is a semi-permanent mid-tropospheric anticyclone located over subtropical southern Africa at approximately 500 hPa. It is thought to be a thermally induced response to latent heat release further north in the tropics (Reason, 2016). The geostrophic winds to the north of the Botswana High make up the **southern African Easterly Jet** (AEJ-s), which modulates rainfall over Central Africa (Kuete et al., 2019; Nicholson and Grist, 2003). As the rainy

season progresses from September through to March, the Botswana High and associated circulations move south. The Angola and Kalahari Heat Lows are located below the Botswana High (Kouete et al., 2019; Vizzy and Cook, 2016). Rainfall over subtropical southern Africa is suppressed when the Botswana High is strong (Reason, 2016), possibly due to enhanced subsidence associated with the anticyclone.

1.3.1.5. *The Congo Air Boundary*

The CAB, also referred to as the Zaire Air Boundary, is traditionally defined as the location where the LLW winds, which originate as recurved Atlantic south-easterlies, meet with the easterly Indian Ocean trade winds at the surface (Taljaard, 1972; Torrance, 1979). The CAB is also a dryline because the converging airmasses differ strongly in humidity, with the LLW being close to saturation after their journey over the Congo rainforest and the easterly trades having been ‘continentalized’ and dried out over the great escarpment (Torrance, 1979; Leroux, 2001). Taljaard (1986) and Van Heerden and Taljaard (1998) describe the CAB as the mean location of the core of a broad zone where active weather dominates, and discuss tropical lows that travel westward along the CAB. In the Austral summer, the CAB extends from Angola eastwards through Zambia, and then turns northward to extend up through east Africa. In winter, it retreats northwards to the Congo basin. This historical view of the CAB differs slightly from the interpretation presented in this thesis (and in Figure 1.2), which for practical purposes defines the CAB as a sharp dryline that breaks down at the end of spring.

Studies of the CAB have primarily focussed on synoptic charts and drawn heavily on the individual forecasting experience of expert meteorologists in the

pre-satellite era (Taljaard, 1972; Torrance, 1979). In the intervening decades, the CAB disappeared from mainstream climate science, though it remained well-known in the weather forecasting and paleoclimate communities (Tierney et al., 2011). No in-depth study of the representation of the CAB in models or reanalysis products has been undertaken. Nor is the seasonality, the frequency or the extent to which the CAB is a discernible feature of the daily climatology well-known. This gap in the literature will be rectified in Chapter 5.

1.3.1.6. *The South Indian Ocean Convergence Zone*

The South Indian Ocean Convergence Zone (SIOCZ, also referred to as the South Indian Convergence Zone) is a diagonally oriented convergence zone that extends from central and southern Africa through Madagascar and into the Indian Ocean (Cook, 2000). It is a region of high precipitation, which may be viewed as the south-most extent of the tropical rain belt. The SIOCZ is weaker and more intermittent than the equivalent features in the other southern hemisphere ocean basins: the South Pacific and South Atlantic Convergence Zones. TTTs form the synoptic expression of the SIOCZ, and will be discussed in Section 1.3.3.4 (Todd et al., 2004).

1.3.2. **A note on reanalysis products**

Because the availability of observational data, and particularly of upper atmospheric measurements, is very low across sub-Saharan Africa (Washington et al., 2006), understanding of the dynamical mean state in the picture described above is primarily based on reanalysis products. However, this same lack of observational data ensures that reanalyses are less constrained in southern Africa than in other parts of the world. Therefore, reanalysis products are vulnerable to biases that may originate from their modelled components. This section examines the

extent to which our view the southern African mean state depends on reanalysis data and the degree to which reanalysis products vary.

Engelstaedter and Washington (n.d.) compared ERA-Interim surface pressure, 2m air temperature, 2m specific humidity and 10m wind speeds to weather station data across the African continent. They found negative temperature and positive humidity biases were present in the tropical rain belt, and that the correlations between the observed and reanalysed time series were weaker in the tropics than on the subtropical margins. Temperature and pressure biases were partially explained by the reanalysis model topography. Very few stations were available in this thesis's key study area.

Zhang et al. (2013) compared the rainfall climatologies of eight reanalysis datasets. They found that the seasonal cycles of precipitation were well represented across the models. Spatial patterns of climatological December - March rainfall were highly variable across reanalysis products, with all models showing too much rainfall in the Congo Basin in this season compared to a gridded precipitation product. This is especially the case in ERA-40, ERA-interim and JRA-25, where the SIOCZ was poorly defined. A mechanism based on differences in moisture fluxes around the Angola Low was proposed to be accountable for this difference.

Due to the uncertainties in the climatologies of reanalysis products, especially in data sparse regions, many authors elect to repeat their methodologies across a selection of reanalysis products. This allows for the state of the climatologies of reanalysis products to be compared. For example, Crétat et al. (2018) studied the 700 hPa vorticity anomaly associated with the Angola Low in the ERA-Interim, MERRA2 and NCEP2 reanalysis products. They found that the anomaly was somewhat weaker in ERA-interim than the other two reanalysis

products, and that the variability of vorticity anomalies was reduced by a factor of approximately two between NCEP2 and the other two reanalysis products. Spatial patterns of lower tropospheric winds were qualitatively comparable over the three reanalysis products. Pokam et al. (2014) found that NCEP1, NCEP2, ERA-Interim and MERRA2 all agree on the structure and dynamics of the LLW winds, though they found that increased resolution in the newer reanalysis products provided a clearer depiction of the circulation dynamics.

Overall, there is sufficient agreement between reanalysis products to suggest that their mean state circulation features of the southern African atmosphere are plausible. However, the practice of presenting results from multiple reanalysis datasets is advisable, as a check on reanalysis model behaviour. Furthermore, the disagreement between spatial rainfall patterns (Zhang et al., 2013) is a result of the fact that reanalysis models, just like climate models, must parametrise rainfall processes. Atmospheric components that are properly resolved, such as geopotential, large-scale winds and horizontal temperature gradients, are more reliably simulated than components such as rainfall and vertical profiles. Therefore, in this thesis, an emphasis is placed on fields that are linked to the large-scale circulation rather than convection, and multiple reanalysis products are compared in each chapter. More detail is provided in Chapter 2.

1.3.3. Synoptic weather systems

One of the goals of this thesis is to characterise the climatology of southern Africa from the viewpoint of local synoptic weather systems. This section will summarise major classes of weather systems in southern Africa. In some instances these weather systems are under-studied in southern Africa, and so this section draws on studies of similar weather systems in other regions of the world, as

well as idealised experiments. Where possible, focus is given to bespoke tracking algorithms that have been used to identify these weather systems in southern Africa and elsewhere. The following discussion focuses on tropical lows, heat lows, tropical discontinuities, TTTs and mesoscale convective systems (MCSs). Two of these are then selected as the primary focus of this thesis.

1.3.3.1. *Heat lows*

Heat lows occur primarily in subtropical regions, where large-scale subsidence associated with the subtropical jets allows clear-sky conditions to persist, such that the land surface may reach high temperatures. Under these conditions, shallow dry convection may occur. The idealised study of RÁCZ and SMITH (1999) provides a detailed description of the dynamics of heat lows. These systems feature a pronounced diurnal cycle. They are a response to high midday surface temperatures, which cause the formation of a shallow low pressure region in the mid-afternoon. The high temperatures excite dry turbulent convection, which inhibits a dynamical response to the low pressure until sunset, when the turbulence dies down and a temperature inversion forms (Parker et al., 2005). At this point, a radial inflow develops at low levels as a direct response to the low pressure, which undergoes inertial rotation overnight to form a cyclonic vortex in the early hours of the morning, and breaks down before midday due to the re-establishment of the turbulent mixing. This means that the pressure and vorticity maxima are out of phase by half a day, and the system is in geostrophic balance only in the daily mean. The spatial pattern of the cyclonic vorticity anomaly is typically quite elongated in idealised studies. The cyclonic vorticity is quite shallow, extending up to about 750 hPa, before being capped by a broader anticyclone associated with the divergent outflow of the shallow convection.

In southern Africa, heat lows form in southern Angola in spring, and in the Kalahari Desert in summer. A temperature inversion and a nocturnal low-level jet were observed by the SAFARI campaign in September and October 1992, which may have been associated with the heat low (Zunckel et al., 1996). The capping anticyclone is the Botswana high, as described earlier, which undergoes a similar southward shift (Kueete et al., 2019). The climatological expression of the heat low is discussed in many studies of southern African climatology, (e.g. Taljaard (1953), Cook (2000), Munday and Washington (2017)) but no dedicated study of the spatial distributions or vertical structures of synoptic heat lows in southern Africa has been published to the author’s knowledge.

In this thesis, the synoptics of southern African heat lows will be considered in Chapter 3. This chapter will present the climatological seasonal cycle and frequency of the Angola Heat Low. A link between heat lows and the CAB will be established in Chapter 5.

1.3.3.2. *Tropical lows*

Tropical lows are cyclonic vortices that propagate westward across southern Africa in the Austral summer. They strongly influence local rainfall and aggregate in the climatological December - February (DJF) mean to form the Angola Low. Case study investigations of southern African meteorology have identified that strong tropical lows in Angola are associated with heavy rainfall and wet spells (Cook et al., 2004; Hart et al., 2010). The prominence of tropical lows over southern Africa was first described by Taljaard (1953), who noted that these lows are quasi-stationary in the Angola region, and are associated with low level convergence.

Tropical lows are well studied across the world, including in India (Godbole, 1977; Hunt et al., 2016; Hurley and Boos, 2015), northern Australia (Berry et al., 2012; McBride and Keenan, 1982) and the low-latitude oceans (Wallace, 1971; Fine et al., 2016). Hurley and Boos (2015) included southern African tropical lows in a study of monsoon depressions across the tropics. They found that tropical lows in Australia, India and the Western Pacific feature a common structure with a tall column of potential vorticity, a warmer anomaly near the surface and a cold anomaly further aloft. The horizontal extent of the associated wind anomalies is around 1000 km, and the vorticity anomaly typically extends from the surface to about 400 hPa. These findings are consistent with detailed studies of Indian and Australian monsoon lows and depressions (Hunt et al., 2016; Kilroy et al., 2017). Hurley and Boos (2015) did not find this characteristic structure in a composite of southern African tropical lows, but this may be due to heat lows also being detected by their algorithm. Tropical lows are generally transient features, and the quasi-stationary behaviour described by Taljaard (1953) seems to be unique to southern Africa. The dynamics of this stationarity are not well understood.

Of the references mentioned above, most studies published in the modern era applied feature-tracking algorithms to identify and track tropical lows and monsoon depressions through reanalysis datasets. Hurley and Boos (2015) and Fine et al. (2016) used the tracking algorithm described by Hodges (1994, 1999) to identify and link cyclonic weather systems through time, while Hunt et al. (2016) and Berry et al. (2012) developed their own algorithms. The extrema of a variety of atmospheric fields were used to identify the locations of tropical lows, including mean sea-level pressure, relative vorticity and potential vorticity. In most cases, the field was first smoothed to a coarser resolution and thresholds

were set on the magnitude of the extrema. These steps were taken to prevent spurious noise from causing the false detection of tropical low events.

To date, no comprehensive study has been conducted focussing solely on the characteristics of southern African tropical lows. Their tracks, longevities, spatial distributions, frequencies, contribution to the climatological rainfall and variability have not yet been systematically examined. Chapter 4 of this thesis fills this gap, by tracking tropical lows through reanalysis data and compiling statistics of track locations, their relationship with ENSO and the proportion of precipitation that they directly account for.

1.3.3.3. *Drylines and convergence lines*

The CAB, defined in section 1.3.1.5, marks the confluence zone of the Congo airmass, which has high near-surface relative humidity and is convectively unstable, with drier subtropical air, which is stabilised by subsidence due to the descending branch of the Hadley Cell (Leroux, 2001). As, in the context of mainstream climate science, the CAB has primarily been studied in the context of synoptic charts, it is reasonable to assume that it may be associated with drylines and/or convergence lines. However, to the author's knowledge, no systematic study of the frequency, climatology of dynamics of either such synoptic feature in southern Africa has been published. The exception is the original CAB literature described in section 1.3.1.5, which does not focus on synoptic timescales so much as the mean climatological state.

Drylines and convergence lines are associated with the edges of monsoons in other regions of the world. In West Africa, the monsoon edge is referred to as the Inter-Tropical Discontinuity (ITD), (Sultan and Janicot, 2000). The ITD is typically identified as a confluence zone between the West African monsoon and

the northerly Harmattan Winds. Unlike the CAB, whose climatological position shifts gradually, the ITD undergoes a rapid jump from 5° to 10° N at the beginning of the boreal summer. This has been linked to advection due to the nocturnal low-level jets associated with the Sahara heat low (Parker et al., 2005; Hagos and Cook, 2007).

A number of methodologies have been applied in regions across the world to identify drylines. The most basic method assumes the existence of the dryline and simply identifies the dryline as a contour of constant humidity. For example, the ITD in West Africa is often defined as the 14°C contour of near-surface dew-point, or the 0.010 kg/kg contour of near-surface specific humidity (Roberts et al., 2015). However, this method may identify a dryline even when no step-change in humidity is present, provided that the humidity changes from above to below the threshold inside the domain. Instead, more complex methods use thresholds of the horizontal gradients of near-surface humidity fields, such as mixing ratio gradients (Arnup and Reeder, 2007, 2009) and specific humidity gradients (Hoch and Markowski, 2005).

Drylines and convergence lines associated with the CAB, and the newly identified Kalahari Discontinuity (KD) are the focus of Chapter 5. This chapter develops an algorithm to detect these features in reanalysis data and presents their historical climatology and relationship with interannual rainfall variability.

1.3.3.4. *Tropical temperate troughs*

TTTs are elongated cloud-bands that form over southern Africa and provide a large proportion of the subtropical summer rainfall to the south of the tropical edge, where annual rainfall is typically low (Harrison, 1984). TTTs occur when

an upper level ridge in the subtropical jet interacts with moist tropical convection from tropical rain belt and allows convection to spread in a band across the region (Todd and Washington, 1999; Macron et al., 2014). TTTs are a particularly important source of rainfall for arid regions, such southern Zimbabwe, eastern Botswana and eastern South Africa, which are limited by subsidence and experience little rainfall from other sources. On average, 15.1 TTT events occur over continental southern Africa per year (Hart et al., 2013). Since TTTs often link with the Congo airmass in Angola, it has been proposed that the Angola Low may be instrumental to their formation (e.g. Todd and Washington (1999); Cook (2000)). The dynamics of the interaction of TTTs with the tropical rain belt and any role of the Angola Low or the CAB in facilitating this interaction on a synoptic timescale are not fully understood.

Hart et al. (2012) developed a bespoke ‘MetBot’ algorithm to detect and track synoptic TTT events in the southern African region. This algorithm used thresholds of daily outgoing longwave radiation (OLR) to detect blobs with OLR below a fixed threshold, subject to constraints on characteristics including latitudinal extent and minimum area. This MetBot has been applied with great success to characterising the dynamics, seasonality, contribution to rainfall and interannual rainfall of TTT events.

While TTTs are very important for subtropical summer rainfall in southern Africa, this thesis is more focussed on tropical rainfall, and its projected decline. Furthermore, in part due to the MetBot described above, the distributions and frequencies of TTTs has already been thoroughly studied. Research into the representation of TTTs in coupled models is also already under-way (James et al., n.d.). Therefore, the decision has been made not to consider TTTs as one of the

synoptic systems of interest in this study. Potential synoptic timescale relationships between TTTs and both tropical lows and the CAB will be considered in Chapters 4 and 5 respectively. However, the climatologies and future change of TTTs will not be presented.

1.3.3.5. *Mesoscale convective systems and complexes*

Mesoscale convective systems (MCS) are large, long lived convective storms that form when clusters of thunderstorms amalgamate (Houze, 2004). Mesoscale convective complexes (MCC) are a special class of MCS, typically characterised by quasi-circular agglomerations of cold cloud in infrared satellite imagery of order 10^5 km². MCCs form primarily in summer and over land, and often occur in transition zones between regions of high and low OLR, and in the lee of elevated terrain (Laing and Fritsch, 1997). Blamey and Reason (2012) identified 70 MCC systems in southern Africa south of 18°S over a period of 8 years, suggesting a frequency of 8.75 per summer, based on Meteosat infrared satellite observations. Though this is relatively infrequent compared to other regions of the world, the intermittency of rainfall over southern Africa means that these events are nevertheless important for rainfall over southern Africa. One case study examined by Blamey and Reason (2009) found that one MCC system contributed two thirds of monthly rainfall in eastern South Africa.

More generally, MCSs are characterised as large contiguous rain areas (Houze, 2004), and so may be identified as such in gridded precipitation datasets. By counting connected sets of grid-cells in TRMM with rainfall over 100 mm/day in each grid-cell and with a minimum area of 3 grid cells as an MCS, Liu et al. (2019) found that over 80% of summer rainfall in sub-Saharan Northern Africa was associated with MCSs.

MCCs are too small to be resolved in climate models, which typically have horizontal resolutions of around 200 km. Therefore, they do not fit all the criteria required for further study in this thesis. Despite their likely importance for southern African rainfall and rainfall change, they will not be considered further. However, it is worth noting that larger weather features in the real atmosphere, such as TTTs, tropical lows and the airmass north of the CAB, often contain embedded MCSs which are likely associated with much of their rainfall.

1.3.4. Interannual rainfall variability

Southern Africa is a hotspot for interannual rainfall variability. This high variability, coupled with a high dependence on rainfall for food, energy and water needs, leaves the region highly vulnerable to drought and flooding events (Conway et al., 2015).

The interannual variability in southern Africa is complex primarily because it is linked to multiple interrelated sources of variability, not all of which are well understood. The foremost source of variability is the El Niño Southern Oscillation, which is linked to southern Africa by a combination of atmospheric and oceanic teleconnections (Lindesay, 1988; Cook, 2001; Ratnam et al., 2014; Blamey et al., 2018). Other factors influencing rainfall variability, which have been identified as variability in land-based stationary pressure anomalies known locally as the Angola Low and the Botswana High, act to modulate the ENSO signal (Reason and Jagadheesha, 2005; Reason, 2016; Driver and Reason, 2017; Crétat et al., 2018; Pohl et al., 2018). Additional modes of variability traced to sea surface temperatures in the adjacent Indian and Atlantic Oceans have also been shown to have an interannual control on southern African rainfall (Goddard and Graham, 1999; Rouault et al., 2003; Reason et al., 2006a). The influencers

of interannual variability considered in this thesis are ENSO and the variability of the Angola Low, which will be reviewed in detail below.

1.3.4.1. *El Niño Southern Oscillation*

The El Niño Southern Oscillation is the leading order cause of rainfall variability in southern Africa (Lindesay et al., 1986). ENSO consists of an interannual oscillation in the sea-surface temperatures and mean sea-level pressures in the equatorial eastern Pacific Ocean, driven by the build-up and subsequent relaxation of warm water in the Eastern Pacific linked with oscillations in the equatorial zonal winds. The variation of such a large body of water has implications across the globe, and southern Africa is no exception.

Southern Africa tends to experience dry conditions during the warm phase of ENSO (El Niño) and wet conditions during the cold phase (La Niña). Further north in East Africa, the opposite response is observed. The resultant spatial anomaly pattern often resembles a diagonally oriented dipole. However, the relationship between ENSO and southern African rainfall is not straightforward: while the most severe and widespread droughts in southern Africa in recent years have been associated with El Niño, not all El Niños have been associated with low rainfall years (Blamey et al., 2018). The primary example of this relationship breaking down was the 1997 - 1998 El Niño, when widespread drought was predicted but not realised (Lyon and Mason, 2007). Similarly, while La Niñas have been associated with unseasonally wet summers, this relationship does not always hold.

The mechanisms by which changes in the Pacific Ocean may propagate through to southern Africa have been frequently studied over the last few decades,

however, the underlying drivers of the teleconnection are still not completely understood. There are three major mechanisms understood to be important for the teleconnection:

- (1) A change in the large-scale atmospheric conditions in southern Africa and the adjacent Indian Ocean due to a shift in the atmospheric Walker Circulation due to ENSO (Lindesay, 1988; Cook, 2001);
- (2) A change in the moisture flux from the Indian Ocean onto the southern Africa due to ENSO-related changes in Indian Ocean SSTs (Goddard and Graham, 1999; Nicholson and Kim, 1997); and
- (3) An extra-tropical response related to the Pacific South American (PSA) anomaly and its influence on the subtropical ridge (Colberg et al., 2004).

Lindesay (1988) first characterised the response of southern Africa as a shift in the Walker Circulation, with the SIOCZ shifting north-eastwards away from (towards) the African continent towards (away from) the Indian Ocean in the El Niño (La Niña) phase. This shift is accompanied by a warming of the Indian Ocean, which makes it a more favourable location for convection. Reason et al. (2000) found that the thermal response of the Indian Ocean was driven by changes in cloud cover and wind strength. As the SIOCZ is a region of enhanced precipitation, its shift into the ocean results in lower rainfall over southern Africa (Cook, 2000). Cook (2001) used an idealised model to show that such a shift in the Walker Circulation may be achieved via the excitation of stationary atmospheric waves by a steady-state Gill-Matsuno like forcing in the eastern Pacific (i.e., the location of ENSO SST signal). This standing wave pattern features statistically significant anti-cyclonic circulations over the north-east Indian Ocean and southern Africa, and a less significant cyclonic circulation over the south west

Indian Ocean at 850 and 500 hPa (Ratnam et al., 2014). The standing wave creates conditions which are less (more) favourable to uplift, moisture convergence and instability over southern Africa (the south west Indian Ocean).

The second mechanism relies on the response of the Indian Ocean to transfer the ENSO signal to southern Africa. Goddard and Graham (1999) showed using model experiments that the correct rainfall dipole response across southern and eastern Africa could only be achieved if the ENSO signal was present in the Indian Ocean. Warming (cooling) of the tropical Indian Ocean due to El Niño (La Niña) reduces the land-sea contrast between itself and southern Africa, which in turn reduces the moisture flux onto the continent. The warming of the Indian Ocean due to ENSO may still be achieved by the atmospheric processes described earlier, however, Goddard and Graham (1999) and Nicholson (2003) state that the rainfall change over southern and eastern Africa comes primarily from the changes in this moisture flux.

In addition to providing a link between southern Africa and the eastern Pacific Ocean, the Indian Ocean possesses its own variability on interannual time-scales which also modulates southern African rainfall (Washington and Preston, 2006). Anomalously warm SSTs in the tropical Indian Ocean, synonymous with the positive phase of the Indian Ocean Dipole (IOD), tend to draw the SIOCZ into the Indian Ocean, causing dry anomalies over southern Africa (Lazenby et al., 2016; Gaughan et al., 2016). Further south, however, rainfall between 12° and 18°S is positively correlated with ($R=0.2 - 0.3$) warm anomalies in the western subtropical Indian Ocean (Behera and Yamagata, 2001). This region, together with a pole further north off Western Australia, forms a spatially coherent dipole pattern, known as the Subtropical Indian Ocean Dipole (SIOD) (Behera and Yamagata, 2001; Reason, 2001; Hoell et al., 2017; Daron et al., 2019). Modelling

studies have indicated that warmer waters to the south west of Madagascar lead to enhanced local evaporation and onshore winds in Mozambique, causing an increase in the moisture flux into southern Africa (Reason, 2002).

The final ENSO teleconnection mechanism is related instead to the extratropical response to ENSO. The PSA pattern is a well-studied ENSO teleconnection. During the warm phase of ENSO, a high pressure anomaly and an associated wave-train are present in the Southern Ocean to the west of the Drake Passage. This acts to disrupt the climatological jets and highs throughout the mid-latitudes. Near southern Africa, the main impacts of the PSA are an eastward shift of the South Atlantic High towards the African continent. Associated subsidence may then inhibit rainfall (Ratnam et al., 2014) and modified winds may induce changes in the Atlantic Ocean SSTs (Colberg et al., 2004).

Whether it occurs via oceanic or atmospheric means, the north-eastward shift of the SIO CZ has been well studied, including on a synoptic timescale. Fauchereau et al. (2009) found that the phase of ENSO influences the frequency of occurrence of a series of spatial convective modes on a daily timescale. The warm phase of ENSO increases the likelihood of two spatial patterns: one featuring increased rainfall over East Africa, and one featuring suppressed rainfall over southern Africa. The modes affected by ENSO correspond to TTTs, described in Section 1.3.3.4. By studying these TTTs directly, Hart et al. (2018) found that these cloud-bands occur less frequently over southern Africa and more frequently in the south west Indian Ocean during the warm phase of ENSO. During the cold phase, they found a lower magnitude increase of events over southern Africa. They propose that these changes may be related to the shift in the location of convective instability (i.e. mechanism 1), as well as changes in the subtropical jet and its separation from the polar jet (linked to mechanism 3). This work

traces the delivery of the ENSO signal to a specific class of synoptic system in the southern extent of southern Africa. To date, no such analysis has attributed the ENSO signal to classes of synoptic weather systems further north along the tropical edge of the rain belt.

It is likely that all three mechanisms are important in conveying the ENSO signal to southern African rainfall, and that the weighting of each component is different from event to event. Blamey et al. (2018) contrasted the dynamical mechanisms of the ENSO signal in the three strongest El Niño events of the last 30 years: 1982 - 1983, 1997 - 1998 and 2015 - 2016. They found that each of these three mechanisms was in play in each El Niño event, but that there were differences in the spatial patterns and intensities of the teleconnection patterns that played into differences in the manifestation of the El Niño signal across the events. The modulation of local climatic features, namely the Angola Low and the Botswana High, were also identified to vary between the three El Niño seasons.

1.3.4.2. *The Angola Low*

The interannual strength of the Angola Low has been identified as a key indicator of whether or not an El Niño event will have a strong response in southern Africa. Dry conditions are associated with El Niños with weak Angola Lows, including 1982 - 83 and 2015 - 16, while conditions closer to the climatological mean have eventuated from El Niño summers when the Angola Low was strong, such as 1997 - 98 (Lyon and Mason, 2007; Blamey et al., 2018). This observation has led to the proposition that the Angola Low may modulate the ENSO rainfall signal over southern Africa. It has been suggested that the behaviour of the Angola Low may explain the failure of ENSO to predict southern African rainfall in some seasons (Reason and Jagadheesha, 2005; Crétat et al., 2018).

As noted above, ENSO is the leading order cause of interannual rainfall variability in southern Africa. The role of the Angola Low in modulating this relationship was first identified by Reason and Jagadheesha (2005), who performed a model experiment with an atmosphere-only GCM studying 5 summer seasons with strong ENSO anomalies. They studied 3 El Niño and 2 La Niña events, including the 1997 - 98 event. Aside from the 1997 - 98 El Niño, the remaining seasons studied showed the typical El Niño response in observations, with decreased (increased) rainfall and a weaker (stronger) Angola Low in El Niño (La Niña) seasons. 1997 - 98 showed less severe dry conditions than the over two El Niño seasons considered, and featured an Angola Low in a near-climatological state. They found that the GCM did not simulate sufficient interannual variability in the strength of the Angola Low compared to NCEP reanalysis, and so in all seasons a near-climatological Angola Low was simulated. Consequentially, only the 1997 - 98 El Niño rainfall was accurately simulated. They concluded that accurate simulation of the Angola Low is crucial for proper simulation of the ENSO signal over southern Africa.

Taking a closer look at climatological anomalies in the summer of 1997 - 98, Lyon and Mason (2007) confirmed that the Angola Low was unusually strong in JFM 1998 as compared to other El Niño seasons, and that it contributed an enhanced moisture flux from the Congo and the equatorial Atlantic into southern Africa. They found that the enhanced Angola Low was a part of an anomalous planetary wave pattern extending across the western subtropical South Atlantic Ocean.

By applying k-means cluster analysis to identify strong, weak, northward and southward states of the Angola Low, Crétat et al. (2018) confirmed that the Angola Low was typically weaker in the warm phase of ENSO. In the La Niña

phase, they found that the Angola Low was more frequently located further south. They found that the weak Angola Low state was associated with drier conditions across southern Africa, and that the southward state was associated with wet conditions.

By considering a 3500 year simulation to bypass problems surrounding the number of years for which observations are available, Pascale et al. (2019) found that the latitude of the Angola Low features variability at daily through to multi-decadal timescales. They again confirmed the existence of a strong relationship between Angola Low latitude and rainfall, and a weaker relationship between Angola Low latitude and the Niño 3.4 SST index.

It is unclear whether variability of the Angola Low’s modulation of the ENSO signal has prognostic or diagnostic power. If it were the former, a stronger Angola Low may cause an increase in moisture flux into southern Africa, enabling increased rainfall despite El Niño conditions. If the latter were true, a stronger Angola Low may simply be a side-effect, driven by a mode of variability external to ENSO that also increases rainfall over southern Africa. Furthermore, in light of the finding of Munday and Washington (2017) that the Angola Low transitions from a shallow heat low to a deeper tropical low in early summer, it is unclear whether either or both of these phases modulate the relationship with ENSO. It is also unclear whether the Angola Low’s interannual relationship with southern African rainfall is local or remote: i.e. whether the increased rainfall during strong Angola Low seasons is rainfall that is directly associated with the low pressure weather systems that form the synoptic expression of the Angola Low, or remotely driven by an overall increase in humidity and/or decrease in stability associated with the flux of oceanic air brought onto the continent by the stronger Angola Low. Chapters 3 and 4 will answer these questions.

1.4. Simulated climate of southern Africa

In this section, CMIP5 simulations of southern African climate are reviewed. Historical simulations are considered first, with particular regard to current understanding of sources of the present day rainfall bias. Then, literature regarding the southern African drying trend and variability in climate model project model projections is presented.

1.4.1. Historical representation in CMIP5 models

The CMIP model experiments are the foremost tool available for the generation of projections of future climate change due to anthropogenic warming. However, realistic representation of the current climate is deemed likely to be a largely necessary (though not sufficient) condition in order for these projections to be plausible. This includes a representation of both climatological processes and historical rainfall amounts. Where model biases in precipitation do exist, an understanding of what model processes may be causing them may be able to justify use of a delta factor projection of change despite the bias. A large body of literature has been compiled over the last decade in assessing the rainfall and circulation biases over southern Africa in the historical CMIP and AMIP simulations of the CMIP5 project.

There is a well-established rainfall wet bias across the summer rainfall region of southern Africa in CMIP5 models, with a spatially coherent ensemble mean magnitude of approximately 1 mm/day. This ranges 30 - 70% of the observed annual mean rainfall, varying between regions of high and low rainfall respectively (Stocker et al., 2013, chapter 9). Dunning et al. (2017) found that the rainfall seasonality is generally well simulated in southern Africa, with climatological onset and cessation dates matching nicely with gridded rainfall observation

products in both the CMIP and AMIP simulations. They found a high degree of spatial agreement in the large-scale pattern of the progression of the rain belt between models and observations, suggesting that the processes driving this progression are well simulated. At the south-west extreme of the continent, they also found that the winter rainfall region was overly large compared to observations, stretching to 25°S on the west coast.

Further north in tropical southern Africa, the wet bias has been studied in the context of central Africa. Creese and Washington (2018) found that the warm SST bias in the equatorial Atlantic contributes to a wet bias near the west coast of central Africa in the SON rainy season. They also found that models that are drier to the east of central Africa have one unified African Easterly Jet located near the equator in SON, whereas wetter models have separate northern and southern jets.

The rainfall bias in subtropical southern Africa is present all year round in the ensemble mean, but is largest in January and February, the height of the rainy season (Munday and Washington, 2017). There is broad inter-model variation in the magnitude of this bias, ranging from 0 - 70% of the observed rainfall magnitude, which (Munday and Washington, 2017) attributes to an over-representation of the Angola Low in CMIP5 models. They find that the low level geopotential height anomaly of the Angola Low varies across models, and is generally stronger than that present in ERA-Interim and NCEP2 reanalyses. The Angola Low also extends too far south in many models, encroaching on the location of the Kalahari heat low and preventing shallow dry convection in this region in favour of deep moist convection. Munday and Washington (2018) refines this argument, showing that an under-representation of the topography and buoyant stability over the eastern escarpment of southern Africa allows the Indian Ocean trade winds to

penetrate further into southern Africa, increasing moisture flux into the interior of the continent. This invigorates the Angola Low and also diverts moisture flux away from Madagascar, creating a dry bias to the east of southern Africa.

Dieppois et al. (2015) found that the interannual variability of precipitation over subtropical southern Africa in CMIP5 models is anomalously low, and that in particular the ENSO teleconnection between southern Africa and the eastern Pacific is under-represented. They showed that ENSO is present in the 24 CMIP5 models that they considered, though in many cases the SST anomaly extends unrealistically far into the west Pacific. However, the teleconnection between ENSO and the Indian Ocean is not properly simulated, which Dieppois et al. (2015) link to a lack of an eastward shift in the SIOCZ and a consequential over-prediction of rainfall during modelled El Niño years. The south and west portions of southern Africa retain a weak teleconnection with ENSO, however, in the north-east, near northern Mozambique and Malawi, the signal disappears and sometimes reverses its sign. Lazenby et al. (2016) found that the SIOCZ is well represented in the mean states of CMIP5 models, although it is overly zonally orientated. Consistent with Dieppois et al. (2015), they found that while the modelled SIOCZ features a realistic degree of interannual variability in the form of north-easterly and south-westerly shifts, this variability does not correspond to Indian or Pacific SSTs, in contrast to observations.

Overall, while the rainfall bias over southern Africa is significant, a great deal of work has gone into attributing the bias to various circulation changes. By considering projected future rainfall changes in the context of these circulation biases, we may therefore make a judgement of whether the circulation bias may invalidate the projection.

However, the majority of the studies presented above have been based on monthly resolution CMIP5 data (or, in the case of (Dunning et al., 2017), focussed on features at a timescale.) No analysis has been conducted on the representation of tropical southern African synoptic weather systems in CMIP5 models. The extent to which climate models are able to realistically simulate common weather processes is fundamental to whether they are capable of realistically projecting future change. Chapter 6 will fill this literature gap by considering the representation of tropical lows and drylines associated with the CAB in CMIP5 models on a daily timescale.

1.4.2. Projected future change

CMIP3 and CMIP5-based rainfall projections exhibit a rainfall decline over southern Africa which is strongest in October, November and December (James and Washington, 2013; Dunning et al., 2018). This rainfall decline has been linked with a delay in the onset date of the rainy season (Dunning et al., 2018), and in many models comes with an increase in rainfall in central Africa in the same season (Aloysius et al., 2016; Creese et al., 2019). However, while most models agree on the sign of rainfall change, they disagree widely on the magnitude of change, with the projected end of 21st century RCP8.5 rainfall change ranging from 0 to 34% of the historical seasonal total (Munday and Washington, 2019).

James and Washington (2013) found that the southern African September-November drying signal becomes statistically significant across climate models that have undergone at least two degrees of global warming. They demonstrated that rainfall is projected to decrease by 0.2-0.4 mm per day in the ensemble mean during the simulated time periods where CMIP3 models have reached 2 degrees above pre-industrial levels. By the time warming levels increase to 4 degrees,

this rainfall decline exceeds 0.8 mm/day in regions of southern Africa. Figure 1.1 showed a similar result for the CMIP5 model experiments, and demonstrated that the projected end of 21st Century ensemble mean rainfall decline over the region of interest between October and December has a magnitude of 0.2-0.4 mm/day in a low emissions scenario (RCP2.6), and 0.7-0.9 mm/day in a high emissions scenario (RCP8.5).

At first glance, drying in southern Africa appears to follow a broader trend of subtropical precipitation decline (SPD), a widely accepted projection of future change. However, while most regions of SPD are largely ocean based and feature only small regions of land. Hydrological projections in these regions feature both a decline in precipitation and precipitation minus evaporation ($P - E$). The southern African drying, meanwhile, is land-based and features a proportional decline in evaporation, so that $P - E$ remains roughly constant. This is likely because evaporation in southern Africa is moisture limited, as climatological $P - E$ must be positive or very small over land (Roderick et al., 2014). Furthermore, by studying idealised experiments that feature an abrupt doubling of CO₂, (He and Soden, 2017) found that SPD occurs on fast timescales, of approximately 1 year, while southern African drying emerges much more slowly, after about 70 years. This result emphasises that there are key differences in the mechanisms that drive SPD and southern African drying. Therefore, southern African drying is globally unique and requires a bespoke approach.

Lazenby et al. (2018) applied the Chadwick et al. (2013) decomposition of rainfall change to the ensemble mean OND drying over southern Africa. The Chadwick decomposition divides rainfall change into components due to: the thermodynamic increases in specific humidity (Held, 2005), the ‘upped-ante’ mechanism due to decreases in relative humidity over land (Chou and Neelin, 2004), the

reduction in convection due to the weakening of the overall tropical circulation (Ma et al., 2012), a residual term that contains changes due to spatial changes in circulation patterns and a cross term due to non-linear interactions. As such, the Chadwick decomposition determines whether simple theory-based mechanisms of change based on moist processes dominate the future change signal (the former three components) or whether more complicated changes are at play. Lazenby et al. (2018) found that although the terms associated with the upped-ante and the weakening of the tropical circulation were negative, appearing as if they could be sources of the drying signal, they balanced out with the change due to thermodynamic increases in specific humidity. This leaves the shift term, whose spatial pattern matches well with the pattern of overall change, solely responsible for the ensemble mean drying. They found that the drying associated with a dynamical northward shift in the location of convection rather than a direct thermodynamic response to a warmer atmosphere.

Meanwhile, Dunning et al. (2018) linked the delayed onset of OND rains in southern Africa to relative changes in the strengths of the Saharan and Angola heat lows. Munday and Washington (2019) found a similar result, that models with a deeper future climatological Angola Heat Low showed a higher intensity of drying, while Cook and Vizzy (2013) observed a strengthening of the Angola Low in regional climate model simulations. At subtropical latitudes, James et al. (n.d.) linked rainfall decline to a decrease in the frequency of TTTs peaking in October. As such, some work has already been done in investigating the local circulation changes relating to the OND drying. However, processes linked to drying in the tropical margins have only been investigated on climatological and seasonal timescales to date.

Less work has looked at rainfall change in the peak of the wet season, when model projections disagree, although Lazenby et al. (2018) suggest that inter-model differences in SST trends in adjacent oceans may explain the diverse range of model projections. Once again, the characteristics of the future change patterns of atmospheric features on synoptic timescales has not been studied. This knowledge gap ensures that the synoptic context of the future change signal, and whether or not a specific class of weather system may deliver the projected change, is unknown. Chapter 6 will frame future Austral spring and summer southern African projected rainfall changes as consequences of projected changes in the behaviour, frequency and spatial distribution of synoptic tropical lows and drylines.

1.5. Thesis aims

The overarching goal of this thesis is to express the climatology, dynamics and future change of the southern African tropical margin from the viewpoint of synoptic weather systems. The focus area is the margin region between the tropics and subtropics in southern Africa. This region is subject to a unique projection of future rainfall decline.

Based on the above literature review, we identify three key candidate synoptic systems that may be used to characterise this future change: Angola Heat Lows, Angola Tropical Lows and the CAB. As we will discover in Chapter 5, the heat lows and the CAB are intrinsically linked and can be considered together. This reduces our features of interest to two classes: tropical lows and drylines/convergence lines associated with the CAB.

Both tropical lows and the CAB satisfy the criteria set out in Section 1.1, as follows. Both the CAB and tropical lows occur in the region of maximum

drying, while tropical lows also occur in the region where interannual variability is maximised. Drylines, convergence lines and tropical lows are all meteorological phenomena that are associated with rainfall, and this thesis will demonstrate statistical relationships with interannual rainfall variability. Both the CAB and tropical lows have length scales in the order of 1000 km, and are therefore resolvable in climate models with resolutions of 200 km, although mesoscale details will not be resolved. Finally, as indicated in the literature review, both the CAB and tropical lows are understudied, particularly in the last 20 years and since the advent of high resolution reanalysis products.

In order to express the rainfall climatology, variability and future change from the viewpoint of the CAB and tropical lows, the following aims must be achieved.

- Aim 1: Develop flexible algorithms that are capable of identifying the CAB and tropical lows in daily and subdaily reanalysis data and model output across a range of spatial resolutions;
- Aim 2: Characterise the behaviour, spatial distributions and frequencies of the CAB and tropical lows in climatological-length datasets;
- Aim 3: Study the dynamics that drive the behaviour of the CAB and tropical lows;
- Aim 4: Determine the extent to which the CAB and tropical lows deliver interannual rainfall variability signals; and
- Aim 5: Characterise future rainfall change in the context of the CAB and tropical lows.

1.6. Notes on the format of this thesis

This work is presented as a thesis by publication. The main results take the form of research articles, which have been submitted to peer reviewed journals.

Each paper is presented as a separate chapter, prefaced by a foreword which explains how the papers achieve the aims of this thesis. An afterword that emphasises relationships between the results of the separate papers is also included in each case. Where appropriate, this will include a small amount of extra analysis. These papers, their journals, and publication status at the time of the submission of this thesis, are listed below.

- Howard, E. and Washington, R. (2018) Characterising the Synoptic Expression of the Angola Low, *Journal of Climate* 31(17), 7147-7165. doi: 10.1175/JCLI-D-18-0017.1;
- Howard, E., Washington, R and Hodges, K. (forthcoming) Tropical Lows in southern Africa: tracks, rainfall contributions and the role of ENSO, *Journal of Geophysical Research: Atmospheres* (Accepted September 2019) doi: 10.1029/2019JD030803;
- Howard, E and Washington, R (forthcoming) Drylines in southern Africa: rediscovering the Congo Air Boundary. *Journal of Climate* (Accepted August 2019). doi: 10.1175/JCLI-D-19-0437.1; and
- Howard, E and Washington, R. Tracing future early summer drying in southern Africa to tropical lows and the Congo Air Boundary. *Submitted to Journal of Climate* (submitted October 2019).

The first paper, which forms the basis of Chapter 3, contributes to Aims 2, 3 and 4 for tropical low events. It also contributes towards aim 2 for heat low events, which will later be shown to be closely related to the CAB. In the second paper (Chapter 4), the focus is solely on tropical lows, and Aims 1, 2 and 4 are achieved. The CAB is formally defined in paper 3 (Chapter 5), and Aims 1, 2, 3 and 4 are addressed with respect to drylines and convergence lines associated

with CAB. Finally, in Chapter 6, paper 4 focusses on Aim 5 with both tropical lows and the CAB in mind.

1.6.1. Authorship declaration

I conducted all the analysis and writing for the four papers listed above. Richard Washington discussed the results, methodology and structure of each paper with me, and provided comments on the drafted manuscripts. He also assisted me in responding to reviewer comments during the peer review process, particularly on paper 2. Kevin Hodges built the main algorithm used in paper 2, helped me to optimise the algorithm for use on reanalysis data in southern Africa, and provided comments on the drafted manuscript before its initial submission.

CHAPTER 2

Data and Methods

This chapter outlines the datasets and methodologies that have been used in this thesis. The analysis presented in Chapters 3 - 5 is based primarily upon reanalysis datasets and gridded precipitation products, while Chapter 6 studies the CMIP5 climate model database. These three types of data are described below, together with Niño 3.4 SST index and the SASSCAL WeatherNet database.

This thesis has relied on a combination of methodological techniques to study the tropical margin of southern Africa. These include feature-tracking algorithms, composite analysis, dynamical budgets and process based assessments of climate models. The step-by-step details of these methodologies will be described in each paper, and so to avoid repetition they will not be presented here. Instead, this section will provide broader context for the feature-tracking methodologies, which cut across chapters in this thesis. It will justify some decisions made and thresholds set during the algorithm design process. Focus will be given to the TRACK algorithm used in Chapters 4 and 6, and Canny edge/ridge detection applied in Chapters 5 and 6.

2.1. Datasets

2.1.1. Reanalysis Products

Reanalysis datasets are the output of a series of atmospheric forecast models which assimilate weather measurements from around the world. They may be

considered to be a best guess self-consistent approximation of the global atmospheric state at subdaily resolution over decadal time scales. Reanalysis products maintain internal consistency by using the same forecast model and data assimilation scheme to assimilate an extended segment of historical data. The accuracy of reanalysis data is constrained by the availability of measurements to assimilate. In regions where measurements are plentiful, confidence in reanalysis is higher than in places where measurements are sparse. Southern Africa falls into the latter category, where reanalysis is more model-based than observation-based. However, these datasets still bear merit due to a dearth of alternative data sources. Certain aspects of the circulation, such as the winds and geopotential, vary more gradually away from assimilated grid cells and so are easier for the models to reproduce than other aspects such as precipitation and cloud processes. Therefore, dynamical aspects of the circulation may be trusted to a greater degree.

Reanalysis models typically assimilate data from conventional and satellite-based data sources. Conventional observations derive from wind profilers, surface-based ship, land station and buoy reports, radiosondes, pilot balloons, dropsondes and aircraft pilot reports. Parameters such as atmospheric motion vectors, surface winds over the ocean, temperature and ozone profiles and microwave sounding radiances are obtained from satellite-based remote sensing (Dee et al., 2011; Gelaro et al., 2017). These observations are assimilated using variational analysis techniques. There are two main types of variational analysis used in the considered reanalyses: 3D, which assimilates instantaneous measurements, and 4D, which assimilates the tendencies of measured quantities.

Reanalysis products are used for three purposes in this thesis. First, they are studied on a subdaily time scale to identify the CAB and tropical lows. Following

this, reanalyses are then used to characterise the climatological frequencies, distributions, variability and dynamics of these weather systems. Thirdly, reanalysis is used to compare the historical representation of tropical lows and the CAB in CMIP5 models to those in the reanalysis products.

The main parameters used to identify the CAB and tropical lows are mid-tropospheric vorticity, surface specific humidity and near-surface wind convergence. Temperature profiles are also used to separate dry and moist convecting lows. Composites of winds, geopotential, humidity and temperature are then studied, along with the first order differentials of the wind field. Reanalysis-derived estimates of precipitation are wholly avoided due to their reliance on unresolved convection, and gridded rainfall products derived from satellite and gauge data are used instead. Three reanalysis products are used in this study: ERA-Interim, ERA-5 and MERRA2. By using a range of reanalysis products, it is intended that some of the uncertainty in the estimated atmospheric state is captured.

2.1.1.1. *ERA-Interim*

ERA-Interim was released in 2011 at a spectral resolution of T255 (approx 75 km resolution at the equator) with 60 vertical levels (Dee et al., 2011). It was developed by ECMWF using version Cy31r2 of the IFS model. The assimilation scheme is a 4D-Var scheme and has a 12 hour analysis window. ERA-Interim applies 4D variational analysis every 12 hours, constraining the upper-air atmospheric state evolution. ERA-Interim is the focus of Chapter 3, and one of the three products that is used in Chapter 4. It is also used to validate the historical representation of tropical lows in CMIP5 models in Chapter 6.

2.1.1.2. *ERA-5*

ERA-5 is an updated ECMWF reanalysis with superior resolution and an updated version of IFS as compared to ERA-interim (Copernicus Climate Change Service (C3S), 2017). The version of IFS used is Cy41r2, and is 10 years more modern than ERA-Interim. Like ERA-Interim, ERA-5 also uses 12-hourly 4D variational analysis. The spectral truncation of the main model is T639, equivalent to 31 km at the equator. ERA-5 also consists of 10 ensemble members at a reduced resolution (T319, or 63 km at the equator). Only the high resolution reanalysis is used in this thesis. ERA-5 is used in Chapter 4 and is the focus of Chapter 5. It is also used for validation of tropical lows in CMIP5 models in Chapter 6. The decision to restrict Chapter 5 to ERA-5 is made because the CAB features mesoscale structure that is not captured in lower resolution reanalyses or models. The large-scale structure of the CAB in ERA-5 is compared to that of CMIP5 models in Chapter 6 by regridding to a $2^\circ \times 2^\circ$ grid using nearest neighbour interpolation.

2.1.1.3. *MERRA-2*

MERRA-2 is a retrospective reanalysis product produced by NASA's Global Modelling and Assimilation Office (NASA, 2016). MERRA-2 uses the GEOS atmospheric forecast model, which has a finite volume dynamical core. The assimilation scheme uses a grid point statistical interpolation scheme with 3D variational analysis calculated every 6 hours and applied through incremental analysis updates (Gelaro et al., 2017). It has 72 vertical levels stretching up to 0.01hPa, and a horizontal resolution of approximately 0.5 degrees in the north - south direction, and 0.625 in the east - west direction. The horizontal grid geometry is based on a cubed sphere. MERRA-2 is available from 1980 through

to the near-present, with a latency of a few months. This dataset is used to cross-check the analysis of Chapter 3, and has a larger role in Chapters 4 and 6.

2.1.2. Gridded Rainfall Datasets

Gridded rainfall datasets are used in this thesis to assess the degree to which interannual variability in the identified synoptic weather systems controls interannual rainfall variability over southern Africa. The rainfall datasets used in this study all combine rain gauge station data with satellite imagery, and are chosen to optimise the length of data overlapping with the reanalysis period considered (1979 - present). Due to a shortage of observational data over the African continent (Washington et al., 2006), gridded rainfall datasets tend to be undesirably dependent on the algorithms used to construct precipitation estimates from the input data (Awange et al., 2016). For this reason, multiple rainfall datasets are considered in this thesis, and key results are required to be robust across multiple datasets. Three datasets are used: GPCP, TAMSAT and CHIRPS. These are detailed below.

2.1.2.1. GPCP

Version 2.3 of the Global Precipitation Climatology Project (GPCP) monthly dataset is described in detail by Adler et al. (2003) and Adler et al. (2018). This product dates from 1979 through to the present day. GPCP has global coverage and has a spatial resolution of $2.5^\circ \times 2.5^\circ$. A series of satellite-based rainfall estimates are combined with each other and with point data from the Global Precipitation Climatology Centre's (GPCC) gauge station network.

Stations from Climate Assessment and Monitoring System (CAMS) and Global Historical Climate Network (GHCN) are used for the period of 1979 - 1985, when GPCC was not yet available. From 1979 to 1987, the satellite data

was sourced from outgoing longwave radiation (OLR) measured by satellites in low-Earth orbits. From 1987 onwards, this was replaced by rainfall estimates based on microwave emissions (over ocean) and scattering (over land) obtained by low-Earth orbit satellites, and infrared-based estimates from geostationary and polar-orbiting satellites. Adjustments made to account for changes in satellite observation techniques over the years are described by Adler et al. (2018).

Because GPCP contains only monthly and pentad values, it could not be used to attribute daily precipitation to tropical low tracks. In this thesis it has been studied in Chapter 3 and used to validate results from other products in Chapter 5.

2.1.2.2. *CHIRPS*

The Climate Hazards Infrared Precipitation with Stations (CHIRPS) dataset uses machine learning techniques to interpolate cold cloud duration (CCD) estimates based on infrared satellite observations with station data (Funk et al., 2015). CHIRPS is available at land points between 50°N and 50°S on a $0.05^\circ \times 0.05^\circ$ grid, from 1981 until the present day.

CCD estimates are used to estimate the proportion of climatological precipitation occurring at each location on a given day. This is then combined with a monthly climatology built from 5 satellite-based estimates of rainfall, infrared brightness temperature and land surface temperatures. Station data is then used to apply a bias correction, using an inverse distance weighting algorithm. CHIRPS is computed on pentadal time intervals, and then aggregated to monthly and dekadal, and disaggregated to a daily resolution using the CFS forecast.

In this study, monthly CHIRPS estimates are used to validate results in Chapters 3 and as the main rainfall dataset in Chapter 5, and daily CHIRPS estimates are used in Chapter 4.

2.1.2.3. *TAMSAT*

Version 3.0 of the Tropical Applications of Meterology using SATellite and ground based observations (TAMSAT) rainfall dataset is an Africa-specific daily rainfall dataset. TAMSAT is available from 1983 to the present. TAMSAT uses Meteosat thermal-infrared observations to derive CCD.

Using the historical satellite-based dataset, CCD is calculated daily and then summed into pentads. This pentadal data is then calibrated against station data to derive calibration parameters, which are applied in real-time to the pentadal CCD estimates to obtain pentadal rainfall estimates. The daily distribution of CCD in each pentad is then used to disaggregate pentadal rain into a daily rainfall estimate.

It is important to note that observational data is only used to calibrate the TAMSAT rainfall estimates, and so is not blended with satellites in real time. This allows for the inclusion of historical station datasets that are no longer operational, which is valuable in Africa, where the number of stations has reduced dramatically over recent decades (Creese and Washington, 2016). However, it may reduce the temporal accuracy of rainfall estimates.

In this study, monthly TAMSAT estimates are used to validate the results of Chapters 3 and 5, and daily TAMSAT estimates are used in Chapter 4.

2.1.3. Niño 3.4 SST index

As described in Section 1.3.4.1, interannual rainfall variability in southern Africa is strongly dependent on ENSO. Therefore, links between the interannual

variability of southern African synoptic weather systems and ENSO are considered in this thesis. ENSO is a complex phenomenon and undergoes its own internal variability, largely in the spatial pattern of the Pacific Ocean SST. During the El Niño phase, variability often manifests in the location of the warm SST anomaly, either located in the Eastern Equatorial Pacific or in the Central Equatorial Pacific (El Niño Modoki). This makes it difficult to capture the entire variability of ENSO in a single index.

However, for simplicity, this thesis uses the Niño 3.4 SST index, which is the historically preferred index for use in southern Africa (e.g. Reason et al. (2006a), Lyon and Mason (2007)). This index is the anomalous sea surface temperature in the region 5°N - 5°S , 120°W - 170°W (Trenberth, 1997). This index overlaps the region both the central and eastern Pacific regions, so that it captures both El Niño and El Niño Modoki events. The Niño 3.4 SST index is used in Chapters 3 and 4.

2.1.4. SASSCAL Weathernet Weather Stations

SASSCAL Weathernet is a network of Automatic Weather Stations (AWS) established by the Southern African Science Service Centre for Climate Change and Adaptive Land Management (SASSCAL) (Kaspar et al., 2015; Helmschrot et al., 2015). Stations are located in Namibia, Zambia, Botswana, Angola and the north-west of South Africa. Most of the data records started in 2014 or 2015, but many of the Namibian stations are older, dating back to 2010.

Most SASSCAL stations measure rainfall, solar radiation, wind speed and direction, relative humidity and barometric pressure at hourly or subhourly time intervals. Incoming data are subject to automatic homogeneity and consistency quality control before being published. However, as with all observational

datasets, more extensive screening may be required to filter inevitable errors. To address this, unrealistic values are filtered from the dataset before use, and analysis using SASSCAL data is designed so that it is not sensitive to random errors. This is achieved by presenting scatter-plots of raw data rather than averaged values.

In Chapter 5, 17 of the 149 stations are used to validate the existence of the CAB and the KD by comparing specific humidity and air temperature to the north and the south of the CAB, and to the east and the west of the KD.

2.1.5. 5th Coupled Model Intercomparison Project

The CMIP5 dataset is a collection of coordinated GCM experiments run by 20 climate modelling groups around the world. Its experiment design and rationale are described by Taylor et al. (2007) and Taylor et al. (2012), which are the main references for this section. Each associated model has been used to simulate a set of 7 core long term (century time scale) experiments, together with shorter near-term integrations and two tiers of additional long-term simulations. This thesis will concentrate on the results of two of the core simulations: historical and RCP8.5, which will be described below.

All GCMs in CMIP5 consist of an atmospheric model coupled to an ocean model. Some of the models considered are Earth System Models (ESMs), which directly model the entire carbon cycle. Flexibility was provided within the CMIP5 framework for modelling centres to include additional components, such as dynamic vegetation and super-parametrisation of convection, so long as they met the basic requirements. This means that the resultant set of models is quite diverse.

However, many components of different models have been developed based on the same literature, and models share code and parametrisation schemes. Therefore, the models in the CMIP5 dataset are not statistically independent from each other (Knutti et al., 2013). Many models share the same biases, such as an inability to properly resolve the stratocumulus cloud decks in eastern ocean basins (Richter, 2015), which results in similar patterns in SST biases. These SST biases tend to drive similar circulation biases across the model dataset.

2.1.5.1. *Historical*

The historical CMIP5 simulation guidelines are designed to simulate the climate from 1850 to 2005. In order to achieve this, time-evolving boundary conditions are imposed based on real-world observations. This includes changing concentrations of CO₂ and various aerosols due to anthropogenic and natural emissions and processes, changing solar forcing, and changing land use type.

This experiment is therefore the most suitable coupled model experiment for comparison with reanalysis products and satellite era observational records. However, since the historical models are coupled experiments, all forms of internal climate variability, including ocean variability, must be simulated. This means that, for example, even a model with a good representation of ENSO is almost entirely unlikely to be in phase with the real ENSO oscillation. Therefore, climate models cannot be compared year-on-year with observations. Rather, climatological distributions of parameters over 20 years or more must instead be compared.

2.1.5.2. *RCP8.5*

The RCP8.5 forcing pathway experiments feature boundary conditions which ensure that the radiative forcing reaches around 8.5 Wm^{-2} by the end of the 21st century. The core experiment time-frame is 2006 - 2100. RCP stands for

‘Representative Concentration Pathway’, and indicates that the concentration of greenhouse gases is prescribed to the model. These models are initialised from their respective historical simulations. In ESMs, which fully resolve the carbon cycle, CO₂ emissions are prescribed, rather than concentrations.

RCP8.5 is designed as a ‘business as usual’ experiment, which projects greenhouse gas emissions in the event that no attempt is made to curb them. Other concentration pathways, including RCP2.6, RCP4.5 and RCP6.0, have also been included in the core or tier 1 experiments, but have not been included in this thesis. This is because the aim of this thesis is to characterise future change in terms of changes in circulation systems, rather than to predict magnitudes of change. As the highest magnitude concentration pathway simulated, RCP8.5 is best suited to showing up robust changes in precipitation since it has a higher signal to noise ratio than the other concentration pathways.

2.2. Methodologies

The overarching methodological approach of this thesis involves identifying and analysing the dynamics and characteristics of two synoptic weather systems: tropical lows and the CAB. Following this, these dynamics and characteristics are used to evaluate synoptic processes in climate models and process based projections of regional climate change are constructed. A recipe setting out the steps of this methodology was presented as a list of aims in Section 1.5. Justification of this synoptic weather system approach was provided in Section 1.2.1. Meanwhile, a rationale of process based model evaluation and process based climate projections used to address Aim 5 was given in Section 1.2.2. The present section presents the two key algorithms used in Chapters 4, 5 and 6 to identify synoptic weather systems, addressing Aim 1. The methodologies used to characterise

the behaviour, spatial distributions, dynamics and impact on interannual rainfall variability (Aims 2 - 4) are unique to each feature and are therefore described in situ in Chapters 3 - 5.

2.2.1. TRACK

TRACK (Hodges, 1994, 1999) is a feature-tracking algorithm that has widely been used to track quasi-circular vortices through meteorological datasets. It has been extensively applied to detect midlatitude storm tracks, however in recent years it has also been used for tropical and subtropical applications at smaller length scales. For example, Hurley and Boos (2015) applied TRACK to build a global climatology of monsoon depressions, Hodges et al. (2017) used it to study tropical cyclones, and Curio et al. (2019) studied Tibetan Plateau vortices. While an extensive range of vortex tracking algorithms have been developed for atmospheric applications (Neu et al., 2013), TRACK is chosen for use in this thesis due to its strong precedent of use at tropical and subtropical latitudes.

TRACK is used to detect tropical lows in Chapter 4. Thresholds and parameters set in the TRACK algorithm are set based on the length-scales of southern African tropical lows as established in Chapter 3, and on the advice of its developer, Kevin Hodges. Vortices are detected according to their lower tropospheric relative vorticity. This procedure produced a superset of lows which includes continental tropical lows associated with the Angola Low, but also coastally trapped Kelvin waves, extra-tropical cyclones and heat lows. A set of criteria are applied as follows to extract tropical lows:

- (1) Tracks must spend at least one time step over the southern African mainland;
- (2) Track longevity must be at least one day;

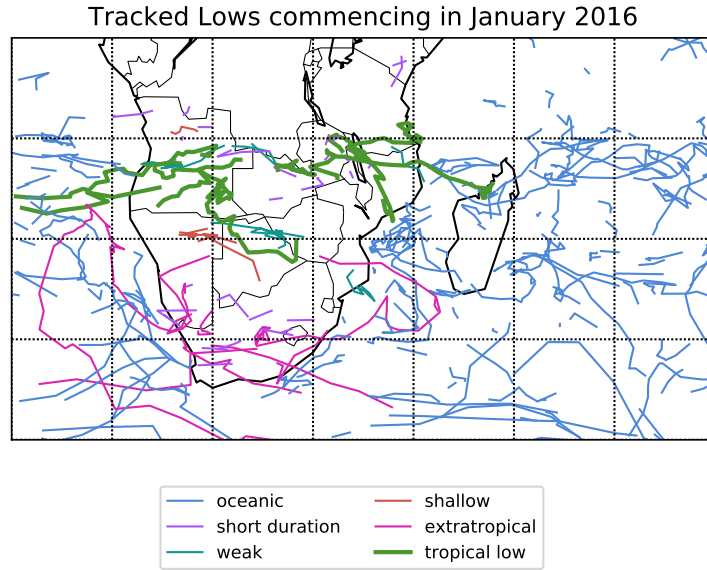


Figure 2.1. Track paths of all lows detected by TRACK commencing in January 2016, coloured by stage of post-TRACK filtering at which they are disqualified.

- (3) The filtered vertical mean relative vorticity must satisfy $\zeta < -3e^{-5}s^{-1}$ in at least one 6-hourly time-step;
- (4) There must be coincident cyclonic vorticity at 600, 700 and 800 hPa for a continuous 24 hour period;
- (5) The genesis location must not be in the Atlantic; and
- (6) The genesis location must be north of 25°S.

Figure 2.1 shows the track locations of all detected vortices in January 2016. Blue tracks indicate those which are excluded by criterion (1) for being entirely oceanic and/or Madagascan, and not making landfall onto the southern African continent. Short duration tracks (criterion (2)) are coloured purple, while weak vorticity tracks are coloured turquoise (excluded by criterion (3)). Tracks that do not feature cyclonic vorticity at every level (criterion (4)) are coloured red, and in most cases occur in the Kalahari, suggesting that these events are shallow

Kalahari heat lows. A cluster of oceanic lows can be seen heading down the Namibian coast. These are coastally trapped Kelvin waves, and although in this sample they are all excluded by the other criteria, the occasional Kelvin wave was erroneously labelled as a tropical low without criterion (5). Finally, extra-tropical cyclones are excluded by criterion (6). Pink lines indicate cyclones that are extra-tropical in origin. The remaining tracks, coloured green and designated in bold, are tropical lows and are studied in Chapter 4, where further validation is provided.

2.2.2. Canny Edge and Ridge Detection

Chapter 5 sets out to detect drylines and convergence lines in atmospheric data. In this section, we detail the methodology by which these features are detected. Separate algorithms have been developed to detect the two forms of lines. The dryline algorithm is discussed first, followed by the convergence line algorithm.

Drylines are sharp boundaries between airmasses with different values of surface humidity. As such, they are characterised by regions of elevated humidity gradient that manifest as one dimensional lines in two dimensional space. Drylines have traditionally been identified by contours of some intermediary humidity value, often a dew-point temperature. However, this methodology will detect false positives when gradual humidity gradients are present. Furthermore, if humidity values either side of the dryline change over the season, this methodology may become biased. For this reason, this thesis uses humidity gradients, rather than contours.

An ‘edge’ in a function of two dimensional space is characterised by a local maximum of the magnitude of the gradient, when restricted to a line parallel to

the direction of the gradient vector. If:

$$\nabla q = M\hat{\theta} = M(\cos \theta \hat{x} + \sin \theta \hat{y})$$

Then an edge must exist where $\frac{\partial M}{\partial \hat{\theta}} = 0$.

The Canny algorithm was originally designed to detect edges in image data, with applications such as photo sharpening and machine vision (Canny, 1986). It discretises the above calculus so that it may be applied to gridded data. Smoothness and convergence properties have been proven for the Canny algorithm provided that the input data is appropriately smooth. In order to meet this smoothness criteria, data must first be smoothed with a Gaussian filter. This process involves convolving the data with a Gaussian kernel of fixed standard deviation (0.5° in Chapter 5 and 2° in Chapter 6), first in the y -direction and second in the x -direction.

The Canny algorithm is explained in more detail in Chapter 5. In short, it uses Sobel operators to calculate a proxy for gradient of q . Sobel operators calculate centred difference derivatives smoothed in the perpendicular direction to the derivative. The magnitude and direction of this gradient are then calculated. At each interior grid point, depending on direction of the gradient θ , combinations of horizontal, vertical and diagonal finite difference forward and backwards derivatives are combined to calculate the discretisations of $\frac{\partial M^+}{\partial \hat{\theta}}$ and $\frac{\partial M^-}{\partial \hat{\theta}}$ are calculated. If these two terms are of opposite sign, then an equivalent continuous function would have to pass through zero. Edges are therefore identified at these grid points.

To calculate convergence lines, a similar approach is used. In this case, points along a ridge-line of high convergence may be identified by examining the roots of

the partial derivative in the direction perpendicular to the ridge axis. However, while the angle of an edge is by definition perpendicular to the gradient vector, the ridge angles are more difficult to calculate. To do so, the methodology of Weller et al. (2017) is adapted.

Weller et al. (2017) calculate the angle of potential ridges by considering an analogy to the inertial tensor. The inertial tensor of a rigid body encodes information regarding the spatial distribution of mass within that body. In particular, the inertial tensor's eigenvectors are the principal axes of inertia. For example, if the body is an ellipsoid, the principal axes are the major and minor axes.

In two dimensions, the inertial tensor is defined as follows, where x_c, y_c is the centroid of the body and ρ is its density.

$$\begin{bmatrix} \iint_S \rho(x - x_c)^2 dx dy & \iint_S \rho(x - x_c)(y - y_c) dx dy \\ \iint_S \rho(x - x_c)(y - y_c) dx dy & \iint_S \rho(y - y_c)^2 dx dy \end{bmatrix}$$

This concept is applied to convergence lines as follows. At each grid cell where the wind convergence exceeds a fixed threshold, a connected region of adjacent grid cells within a set radius is identified. This connected region is then taken as the rigid body, with density (ρ) replaced by convergence $\nabla \cdot \bar{u}$. Hence, summed over all points in the connected region, the tensor becomes:

$$\begin{bmatrix} \sum_i (\nabla \cdot \bar{u}_i)(x_i - x_c)^2 & \sum_i (\nabla \cdot \bar{u}_i)(x_i - x_c)(y_i - y_c) \\ \sum_i (\nabla \cdot \bar{u}_i)(x_i - x_c)(y_i - y_c) & \sum_i (\nabla \cdot \bar{u}_i)(y_i - y_c)^2 \end{bmatrix}$$

The eigenvalues and eigenvectors of this matrix may then be calculated. At a grid cell that is at the centre of a ridge, the eigenvector associated with the larger eigenvalue will then be in the direction of the ridge, and the smaller will be

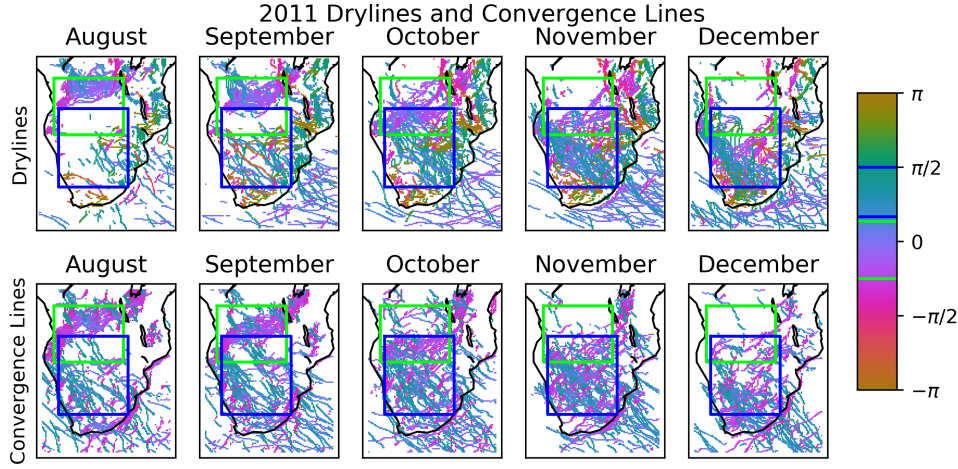


Figure 2.2. Drylines (top) and convergence lines (bottom) identified in 2011. The angles of the respective line are indicated by line colours. Locations and angle thresholds for the CAB are shown in green, and the KD in blue.

perpendicular to the ridge. The direction of this second eigenvector, v_2 , is the equivalent of the gradient vector in the edge-finding algorithm. The methodology therefore proceeds by searching for grid cells where $\frac{\partial \nabla \cdot u}{\partial v_2}$ is minimised, as described above for the edge finding algorithm.

In Chapter 5, thresholds are applied to the locations and angles of drylines and convergence lines in order to classify subsets of the lines associated to the CAB or to the KD. These thresholds are indicated in Figure 2.2, which shows all lines identified in 2011. Drylines aligned parallel to the coast have been removed. A cluster of drylines and convergence lines oriented near-zonally along the southern edge of the Congo is apparent in pinks and purples from August to October. The CAB thresholds have been optimised in order to isolate these features. Similarly, the KD lines stretch across the Kalahari in blue colours, in all seasons as convergence lines and from October to December as drylines. The KD thresholds are set in order to detect only these lines. The remaining drylines and

convergence lines are generally associated with transient weather systems and do not exhibit clear clustering.

In Chapter 6, this algorithm was modified in order to detect drylines in lower resolution models as well as ERA-5. In order to compare across resolutions, all data was regridded to a $2^\circ \times 2^\circ$ resolution. Nearest neighbour regridding is used so that humidity gradients in higher resolution datasets would not be differentially smoothed compared to the coarsest datasets.

The Canny edge and ridge algorithms provide a robust means of identifying drylines and convergence lines in meteorological datasets. This methodology may be applied to drylines in other regions of the world, and, with a little modification, may also be used to identify fronts in other fields. The code used to calculate the locations of drylines and convergence lines in daily ERA-5 reanalysis data has been made available at <https://github.com/EmmaHoward/drylines>.

CHAPTER 3

Characterising the Synoptic Expression of the Angola Low

The Angola Low is a key feature of the southern African wet season atmosphere which influences precipitation across the continent. This chapter uses ERA-Interim reanalysis to show that the synoptic expression of the Angola Low is a combination of dry heat lows and moist tropical low-pressure systems. Angola Heat Low and Angola Tropical Low composites are contrasted against similar lows observed in other continental tropical regions and found to be broadly comparable. The implications that the distinction between dry and moist events has for the interannual relationship between the Angola Low, precipitation and ENSO are examined. The tropical lows exhibit unusual semi-stationary behaviour by lingering in the Angola region rather than travelling offshore. This behaviour is proposed to be caused by an integrated sea breeze-anabatic wind which enhances (inhibits) cyclonic vorticity stretching and convection inland (near the coast). The combined effect of the heat lows and the anchored tropical lows creates the Angola Low in the climatological average. By elucidating the mechanisms of the Angola Low, this research improves the foundation of process-based evaluation of southern African present and future climate in CMIP and AMIP models.

3.1. Foreword to Chapter 3

Chapter 3 addresses the link between weather systems and climate features in southern Africa during the Austral spring and summer. It considers the climatological Angola Low, determining which weather systems provide its cyclonic

vorticity anomaly on synoptic timescales. Two important classes of weather systems are identified: heat lows and tropical lows. Composite vertical structures and spatial distributions of these lows are presented in accordance with Aim 2. A regression analysis reveals that the frequency of tropical lows is correlated with interannual rainfall variability, while the frequency of heat lows is not (Aim 4). A dynamical interaction between tropical lows and the west-coast seabreeze is then presented, elucidating the dynamics of tropical low behaviour in southern Africa (Aim 3).

This chapter was published as a journal article in the *Journal of Climate* in September 2018, following acceptance in April 2018. Some minor errors in the regression statistics presented in Tables 3.1 and 3.2 have been corrected; these did not impact on the statistical significance of the results. Subsequent to publication, the relationship between the synoptic heat lows identified in this chapter and the Congo Air Boundary discussed in Chapter 5 was discovered. For clarity, this result is best presented using the compositing framework discussed in this chapter. Therefore, it is included in the afterword at the end of this chapter.

3.2. Introduction

Precipitation underpins the lives of 150 million people in southern Africa¹. Shifts in rainfall can undercut agricultural production, undermine water, food and energy security, and ultimately threaten the economic viability of the region (Conway et al., 2015). In this region, rainfall is strongly variable on a wide range of timescales, from intra-seasonal through to decadal (Reason et al., 2006a). The local climate dynamics driving precipitation variability are complex, and a comprehensive understanding of the processes which force the local climate

¹Loosely defined here as mainland Africa south of 10°S.

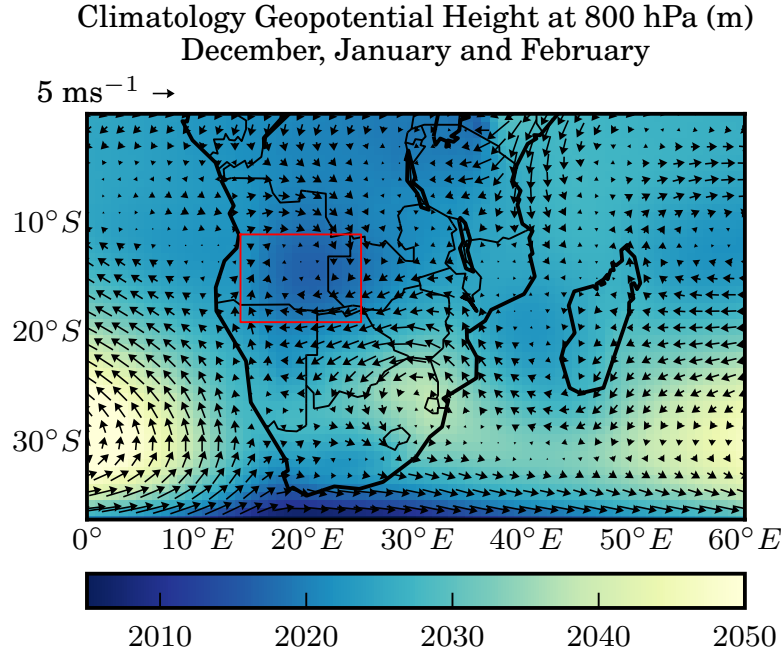


Figure 3.1. ERA-Interim mean geopotential height (filled contours) and winds (vectors) at 800 hPa over southern Africa over the months of December, January and February from 1979 to 2015. The Angola Low is visible as a low-pressure system featuring cyclonic circulation centred at 13°S and 20°E. The red box indicates the primary region of interest for this study.

remains elusive. With approximately 60% of the region over 800 m above sea level (NOAA, 1988), and spanning over 20° of latitude from the tropics to the midlatitudes, southern Africa exists in the nexus of competing climatic features. One such feature, the Angola Low, is known to have a central influence over wet season precipitation across the subcontinent (e.g. Reason and Jagadheesha (2005), Cook et al. (2004)).

The Angola Low is a semi-permanent low-pressure system associated with cyclonic circulation. It is easily identifiable in the December-January-February climatology of near-surface geopotential height (e.g. Munday and Washington (2017)) as shown in Figure 3.1. The system is centred over eastern Angola at

about 13°S and extends into surrounding countries. It is associated with the convergence of moisture flux originating from the western Indian and south east Atlantic Oceans, and thus modulates moisture transport into the subcontinent (Rouault et al., 2003).

The Angola Low was first named as a distinct feature by Zunckel et al. (1996) although the first comprehensive analysis of its meteorology was developed by Mulenga (1998). Perhaps because of the dearth of circulation data over remote western Zambia and war-torn eastern Angola, the integrity of the closed Angola Low circulation had remained elusive to pioneers of southern African climate who described the broad convergence zone in which the feature tends to develop as the Congo Air Boundary (Taljaard (1953) and Taljaard (1972)), and also as the Zaire Air Boundary.

3.2.1. Angola Low significance

The development of climate models and reanalysis products has shone light on the vital role of the Angola Low in the dynamics of southern African climate. Cook et al. (2004) studied composites of wet and dry spells based on South African rain gauge data. They found that wet spells are associated with stronger Angola Low circulations in NCEP2 reanalysis data than dry spells. In addition to this, Munday and Washington (2017) have shown that in some regions of Southern Africa, 40 - 60% of the inter-model precipitation variability between historical CMIP5 models can be explained by the simulated depth of the Angola Low.

The dominant mode of interannual variability in southern African precipitation is the El Niño Southern Oscillation (ENSO) (e.g. Lindesay (1988)). Through the El Niño phase, ENSO is a key driver of some of the most severe recent droughts in the region. Based on an examination of the circulation over southern Africa

in a well spread sample of 3 El Niño and 2 La Niña summers, Reason and Jagadheesha (2005) suggested that the Angola Low has a modulating influence of ENSO on southern African precipitation.

There is also evidence that the Angola Low may be a precursor to tropical-temperate cloud band (TTT) formation (Todd and Washington (1999) and Hart et al. (2010)), which provide a significant proportion of rainfall across the whole of the southern African region (Harrison, 1984). The Angola Low enables southward transport of atmospheric water vapour from the tropics, crucial to the development of TTTs. In an idealised model experiment, Cook (2000) found that an idealised thermal low similar to the Angola Low acts as an origin region for a land based convergence zone analogous to the South Indian Convergence Zone (SICZ), the time-mean manifestation of TTTs.

Future changes in the Angola Low are likely to impact southern African precipitation. Vizzy and Cook (2016) observe a recent strengthening of the Angola Low in multiple reanalysis datasets from 1982 through to 2013, by examining trends in mean sea-level pressure and surface winds. They find that this is associated with sea surface temperature warming off the Angola Coast in concert with a decrease in coastal upwelling in the eastern South Atlantic. Similarly, Vizzy et al. (2015) studied a regional climate model representing projections of southern African climate into the late 21st century. In their simulations, a shortening of the wet season over Malawi was linked to a projected strengthening of the continental lows, including the Angola Low, in April, associated with increased surface heating due to anthropogenic climate change.

Considering its importance to the regional climate, it is crucial that the mechanisms which drive the Angola Low are well understood so as to increase confidence in future projections of southern African precipitation. However at

present, the dynamics of the Angola Low, particularly on a synoptic timescale, are not clear. Furthermore, a paucity of measurement data in tropical Africa means that process-based evaluation of climate models is often more feasible than performance-based precipitation evaluations (James et al., 2018). An understanding of the mechanisms of the Angola Low will allow for an evaluation of the southern African synoptic scale processes simulated by climate models. This will reduce the uncertainty around future projections of regional precipitation changes.

3.2.2. Angola Low dynamics

Traditionally, the Angola Low has been considered to be a dry thermal low, following the theoretical framework of Rácz and Smith (1999). The idealised work of Spengler et al. (2005) and Reason (1996) predict that at tropical and subtropical latitudes, continental heat lows will form on the western sides of continents due to the interaction of the background easterly flow with the surface heating and topography. Many aspects of the Angola Low are consistent with this framework. Mulenga (1998) found that the Angola Low could be formed in a quasi-geostrophic model of southern Africa as a Matsuno-Gill response to surface heating, using a similar method to Leslie (1980).

However, there is emerging evidence that moist convection may be as important to the Angola Low as dry convection. Mulenga (1998) remarks that the Angola Low may act as an anchor point for deep tropical convection. Further to this, Munday and Washington (2017) found that the convection driving the Angola Low shifts from being shallow and dry to moist and deep at around midsummer each year. Particularly notable instances of deep convection in the Angola region have occurred when Indian Ocean tropical cyclones, including Eline

in February 2000 (Reason and Keibel, 2004) and Bonita in January 1996, (Mudenda and Mumba, 1996) have crossed onto the African continent in Mozambique and traversed up the Zambezi River Basin to merge with Angola Low.

3.2.3. Chapter aims and structure

Despite its evident importance to southern African climate system on many timescales, the Angola Low is typically considered as a feature of the seasonal mean. Little attention has been paid to its dynamics or synoptic expression. The focus of this chapter is to perform a detailed analysis of the Angola Low as modelled in the ERA-Interim reanalysis database.

We diagnose the Angola Low as a combination of early season heat low events and late season transient tropical low events, which we denote the Angola Heat Low and the Angola Tropical Low. We investigate the bearing this division has on the relationship between the Angola Low and precipitation. Our findings indicate that the interannual variability of the tropical lows is correlated to interannual summer precipitation variability. In contrast, we find that the interannual variability of the heat lows has no bearing on precipitation variability. While heat low events fit nicely into the idealised theory described by Rácz and Smith (1999) and Spengler et al. (2005), the tropical lows are dynamically similar to transient monsoon lows and depressions that have been observed across India and northern Australia (Hunt et al., 2016). These southern African tropical lows, which form within the tropical rainband, are semi-stationary and linger in the Angola Low region, in contrast to those which form elsewhere. We find evidence suggesting this behaviour may stem from the interactions of the west coast sea breeze with an anabatic wind associated with the steep escarpment along the Angola-Namibia

Coast. This integrated sea breeze-anabatic wind will be referred to as an anabatic sea breeze for brevity.

The remainder of this chapter proceeds as follows. Section 3.4 classifies the Angola Low as a series of thermal lows (the Angola Heat Low) and tropical lows (the Angola Tropical Low). We then examine the synoptic characteristics of each of the two phases. In section 3.5 we establish the influence of ENSO on the Angola Low, and then examine the different effects of the Angola Heat Low and the Angola Tropical Low on precipitation. The chapter then moves towards an understanding of the local drivers of the Angola Low in section 3.6. The final section summarises the research findings and discusses its implications.

3.3. Data and methods

This study uses the ERA-Interim reanalysis dataset at native resolution (0.75 degrees), as described by Dee et al. (2011). 37 Austral summers were analysed, starting in September 1979 and ending in March 2016. Data is analysed on a daily timescale unless otherwise indicated. The primary region of interest is southern Angola and northern Namibia (11 - 19°S, 14 - 25°E). This analysis was repeated using 3 hourly MERRA2 data, spanning from September 1980 through to March 2016. The results of this analysis were qualitatively similar to those obtained from ERA-Interim and led to the same conclusions. For brevity, only the ERA-Interim results are presented in this chapter.

To identify Angola Low events, we consider the daily mean vorticity within the region of interest described above. For each day in the sample period, grid cells with vorticity at 800 hPa less than $-4 \times 10^{-5} \text{ s}^{-1}$ were classified as Angola Low grid cells, and classified as a heat low or tropical low. The choice of the vorticity threshold and the definition of classification system are described in

section 3.4.1. The centre of an Angola low event at a point in that time is then calculated as the centroid of a group of adjacent Angola Low grid cells of the same phase. This method allows for multiple events to exist in the region of interest at a period of time, provided that none of their constituent grid cells are adjacent. The average number of grid cells which were identified in each cluster was 6.2, implying a radius of about 100 km. Each cluster represents the core of a cyclonic system, and so the total radius of the cluster is often larger than this core. 90% of the clusters contained less than 15 grid cells. However, the distribution of the cluster sizes was highly skewed, and the largest cluster identified contained 57 grid cells. This cluster occurred on the 18th of January 1996, when Mudenda and Mumba (1996) report that ex-Tropical Cyclone Bonita had merged with the Angola Low. The sensitivity of all results to the relative vorticity threshold has been tested using a threshold of $-3.5 \times 10^{-5} \text{ s}^{-1}$. This increased the number of Angola Low grid cells flagged, but all other results were qualitatively unchanged and the statistical significance of the results still held in all cases.

We have used composite analysis to study the structure of various atmospheric fields during Angola Low events. In order to remove the effect of the seasonal cycle, a 14 day running mean climatology is subtracted from both the composite sample and the population before testing. A two-tailed Welch's t-test (Welch, 1947) is then applied to test the null hypothesis that the composite mean anomaly is the same as that of the climatology. Autocorrelation within the composite samples has been controlled for by assuming that the data follow a first order autoregressive process and calculating an effective sample size using the lag-1 autocorrelation coefficient. The false discovery rate is controlled by calculating a threshold level p_{FDR}^* based on an FDR control value of $\alpha_{FDR} = 0.05$ (Wilks, 2011). A combined value of p_{FDR}^* is calculated for each multi-panel figure and

is presented in each figure caption. Despite the use of anomalies to calculate significance, we have elected to show the full fields in the composite plots. We found this displayed greater clarity of the overall results.

3.4. Synoptic characteristics

In order to study the synoptic events which comprise the Angola Low, we generated a set of time and space coordinates in the study area which featured strong cyclonic relative vorticity. We then studied the phase space of various atmospheric variables at the identified coordinates. This revealed two clear clusters of synoptic events. In this section, we describe the method used to classify the two clusters, the synoptic characteristics of the two clusters of low-pressure systems, and finally the behaviour of the systems. Despite the low latitude of the study area, we find that geostrophic balance is still a useful approximation. Above the boundary layer, the overall magnitude of the ageostrophic component is about 33% of the overall magnitude of the geostrophic component.

3.4.1. Classification system

The distinction between the Angola Heat Low and the Angola Tropical Low on the daily timescale has been characterised by considering the dry static stability of events with strong cyclonic vorticity. For each day in the study period, grid cells in the primary region of interest (11 - 19°S, 14 - 25°E) with daily mean relative vorticity at 800 hPa less than $-4 \times 10^{-5} \text{ s}^{-1}$ have been identified. A heat map of the static stability at 700 hPa against the 800 hPa relative vorticity is given in Figure 3.2. From this it can be seen that there are two distinct clusters of events, those with high static stability (above 0.0033 K/m and coloured blue) and those with low static stability (below 0.0033 K/m and coloured red). The lower panel of Figure 3.2 shows a parameter space heat map of specific

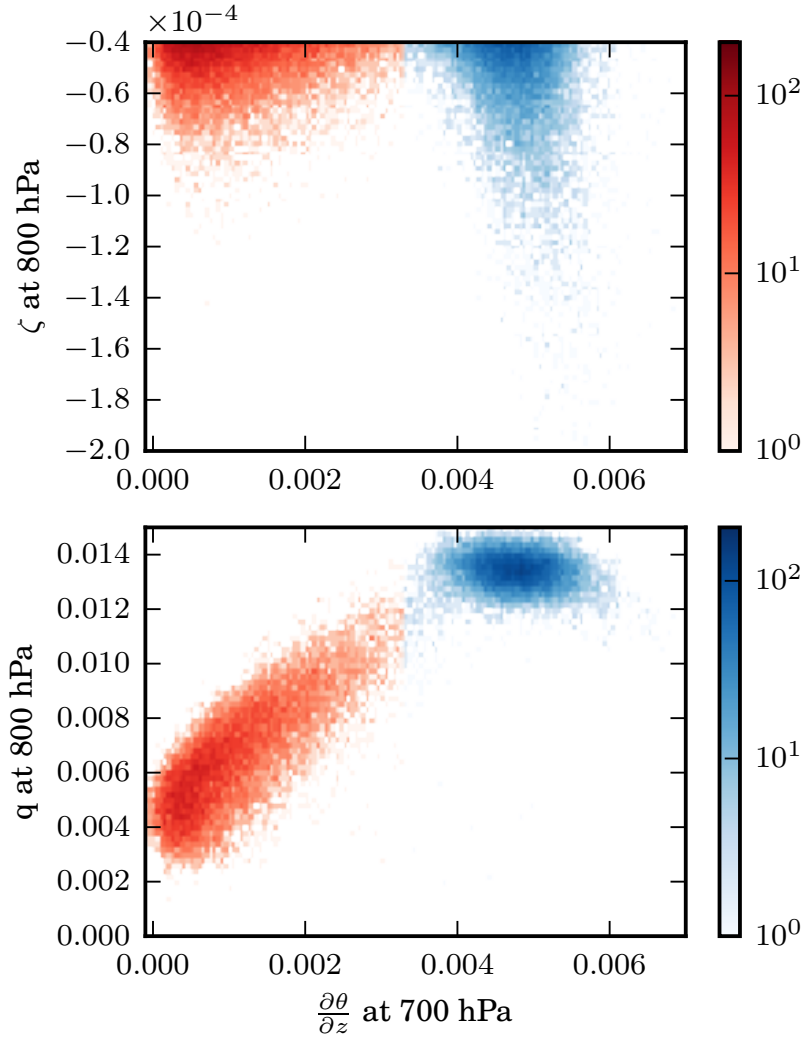


Figure 3.2. Log-scaled phase space heatmaps of relative vorticity, stability and humidity in the Angola Low region on days featuring cyclonic relative vorticity exceeding $4 \times 10^{-5} \text{ s}^{-1}$. Top: 800 hPa Relative Vorticity against 700 hPa stability, and bottom: 800 hPa specific humidity against 700 hPa stability. Blue areas show tropical low grid cells, while red areas show heat low grid cells.

Consistent with the finding of Munday and Washington (2017), a seasonal distinction between occurrences of the Angola Heat Low and the Angola Tropical Low is apparent. Figure 3.4 shows a 2D histogram map of where intense cyclonic circulation associated with dry and moist convection occur from October through

to March. The method used to identify heat low and tropical low grid cells is as described above, however, the results are shown for southern Africa. Dry convection is strongly evident in Angola from October to November, and moist convection is present from December through to March.

Also apparent is the moist convecting Mozambique Channel trough and the dry convecting Kalahari heat low, which are not the focus of this research. It is clear that the Angola Low presents as the Angola Heat Low from October through to November, and then transitions to the Angola Tropical Low during December when the wet season begins, and remains as the Angola Tropical Low until March. This leads us to investigate the synoptic structure of these two phases separately below.

3.4.2. Angola Heat Low dynamics

Key diurnal characteristics of the Angola Heat Low for comparison to the literature surrounding idealised heat lows are presented in Figure 3.5. The figure shows diurnal vertical west-east cross-sections of winds and potential temperature during heat lows identified using the methodology described in section 3.4.1, centred on the centroid of the heat low grid cells. These cross-sections are consistent with Figures 6 to 9 of RÁCZ and SMITH (1999) and Figure 3 of SPENGLER and SMITH (2008), which demonstrate vertical cross-sections of potential temperature and winds in idealised heat low experiments. The authors found that the radial wind inflow is strongest overnight and rotates into a geostrophic tangential wind in the early morning, and that the potential temperature at the centre of the heat lows is unstable in the middle of the day. A mid-level anticyclone sits above the heat low and is strongest in the morning.

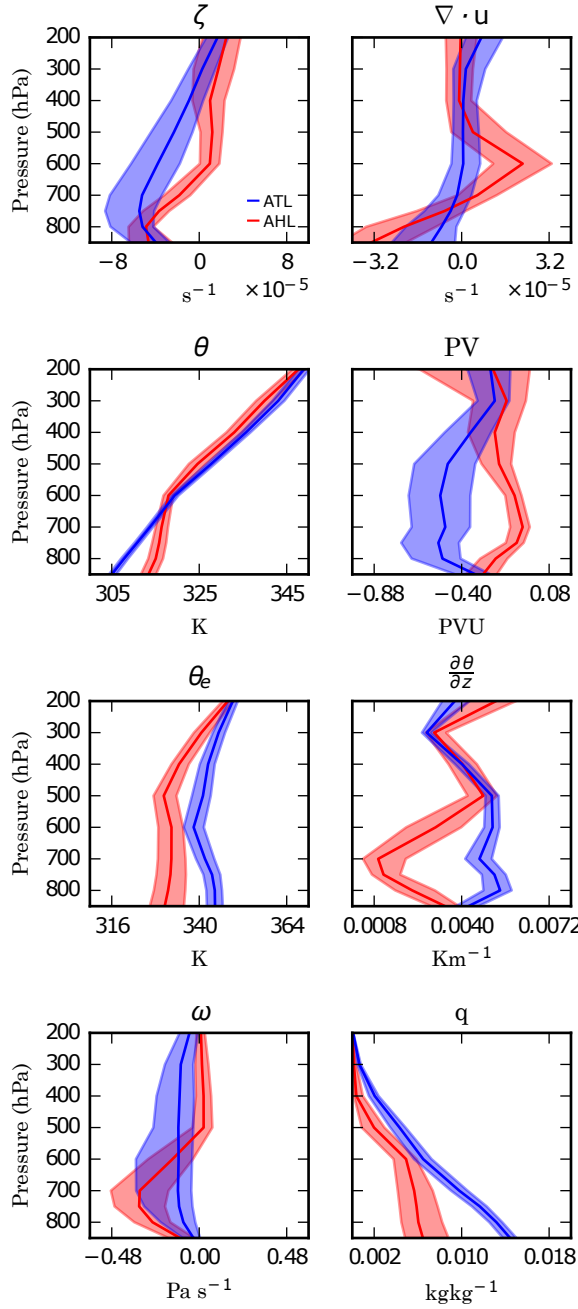


Figure 3.3. Vertical profiles of relative vorticity (left, first row), divergence (right, first row), potential temperature (left, second row), potential vorticity (right, second row), equivalent potential temperature (left, third row), potential temperature lapse rate (right, third row), vertical velocity (left, fourth row), and specific humidity (right, fourth row) during Angola Low events. Heat low profiles are shown in red and tropical low profiles are blue. Solid lines indicate the median value of the distributions, while the coloured bands represent one standard deviation either side of the median.

The main difference between the idealised models and our ERA-Interim based analysis is that the instability is weaker in the reanalysis. The weak instability may result from the averaging of many heat lows in our composite. The westerly

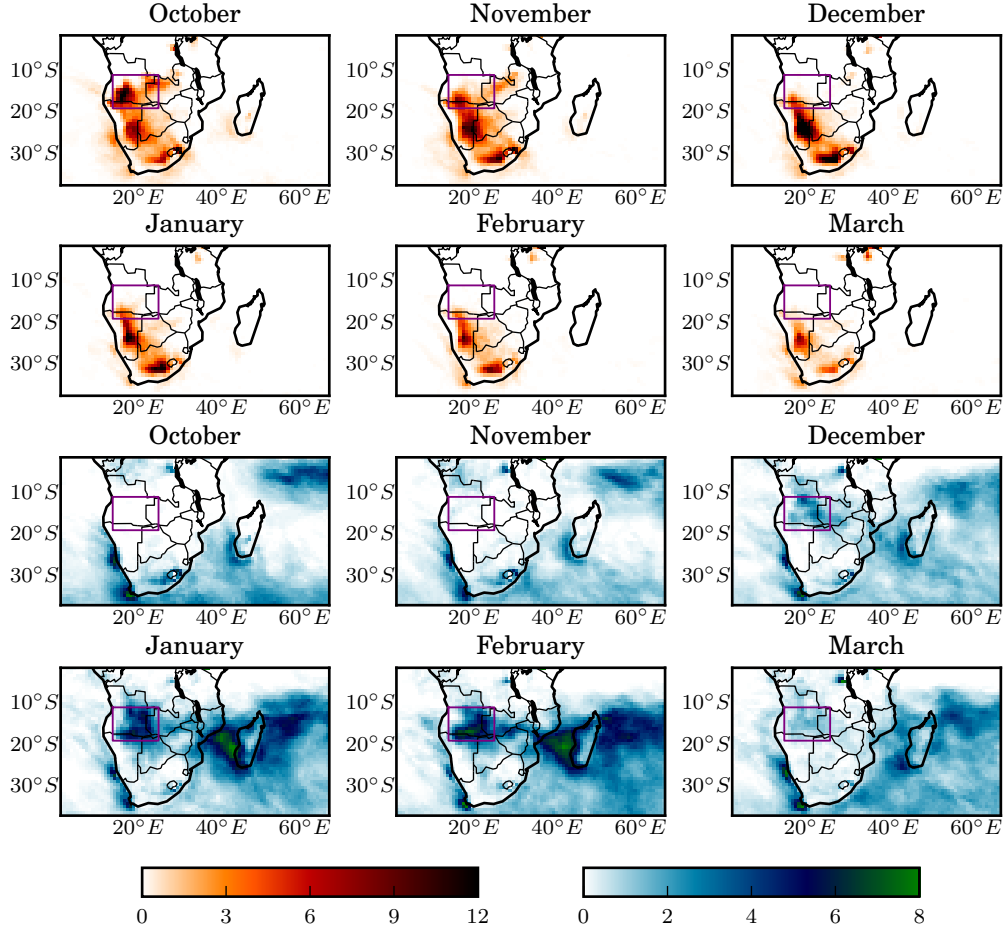


Figure 3.4. Monthly heat map histograms of the locations where cyclonic circulations ($\zeta < -4 \times 10^{-5} \text{ s}^{-1}$) with neutral and unstable dry static stability have been identified in each month. The panels show monthly occurrences of neutrally stratified cyclones with $\frac{\partial\theta}{\partial z} < 0.0033 \text{ Km}^{-1}$ at 700 hPa (left column) and stably stratified cyclones with $\frac{\partial\theta}{\partial z} > 0.0033 \text{ Km}^{-1}$ at 700 hPa (right column). The colour-scale represents the average number of events occurring at each grid point in a given year.

zonal inflow resembles a sea breeze, which will be further discussed in section 3.6. We also find that the upper-level jet which caps the upper-level anticyclone is significantly stronger during the heat low than in the climatology, despite the fact that the climatological seasonal cycle was removed when statistical significance

was calculated. Overall there is satisfactory evidence that the Angola Heat Low is indeed a thermal low in the traditional sense.

3.4.3. Angola Tropical Low dynamics

As cold-cored synoptic-scale lows that track over a tropical landmass, the tropical lows in the Angola region bear resemblance to tropical low-pressure systems, including monsoon depressions. Monsoon depressions have been most intensively studied over the Indian Subcontinent (e.g. Hunt et al. (2016), Godbole (1977)), but have also been studied over northern Australia (Berry et al., 2012). Hurley and Boos (2015) conducted a comprehensive study of these features across low latitude land masses and noted their similarities and differences across different regions of the globe, including southern Africa.

The Angola Tropical Low consists of a deep column of potential vorticity, extending from the surface to about 300 hPa (Figure 3.6). The panels in Figure 3.6 show daily vertical west-east cross-sections of various atmospheric variables during tropical lows identified using the methodology described in section 3.4.1, centred on the centroid of the tropical low grid cells. The temperature anomaly field features a dipole which is cool near the surface and warm in the upper-troposphere. In contrast with the Angola Heat Low, the upper-level zonal winds during the Angola Tropical Low are easterly, suggesting that the tropical lows are embedded in the tropical easterly jet. The cyclonic circulation anomalies reach out from the centre of the system approximately 500 km, giving the total system an average diameter of 1000 km. These observations are consistent with the structures of the Indian and northern Australian monsoon lows and depressions observed by Hunt et al. (2016), Berry et al. (2011) and Hurley and Boos (2015).

This implies that the growth and propagation mechanisms of these circulations may resemble those of the tropical lows in Angola.

The implication that some low-pressure systems over southern Africa are dynamically similar to monsoon depressions in Australia and India is not immediately reconcilable with the work of Hurley and Boos (2015). The southern African composites of Hurley and Boos (2015) do not show the characteristic temperature or PV structure of a typical monsoon low. However, their composite sample contains data from December to February and is performed over an area which extends down to 25°S. Hence the sample will contain Kalahari and Angola heat lows as well as tropical lows, which would be expected to obscure the signal of the tropical depressions. Therefore we conclude that the tropical lows in the Angola region are dynamically related to the monsoon lows that have been observed over Australia and India. Hurley and Boos (2015) identified on average 12.5 low-pressure systems from November to February in southern Africa, in contrast to 25 over the same period in Australia and 18 from May to August in India. Even before accounting for the fact that some of these systems may be heat lows, tropical lows are less common in Southern Africa than in these other regions.

3.4.4. Movement of synoptic events

Figure 3.7 shows the longitudes and timing of grid cells in seven selected years which meet the threshold criteria of heat low and tropical low events. The years displayed in Figure 3.7 have been chosen to represent a range of ENSO phases. In this instance, the domain has been extended to (11 - 19°S, 0 - 55°E) and the classification has been run over six hourly data. Heat lows, shown in red, develop in two longitudinal bands centred on 18 and 22°E, which sometimes merge and

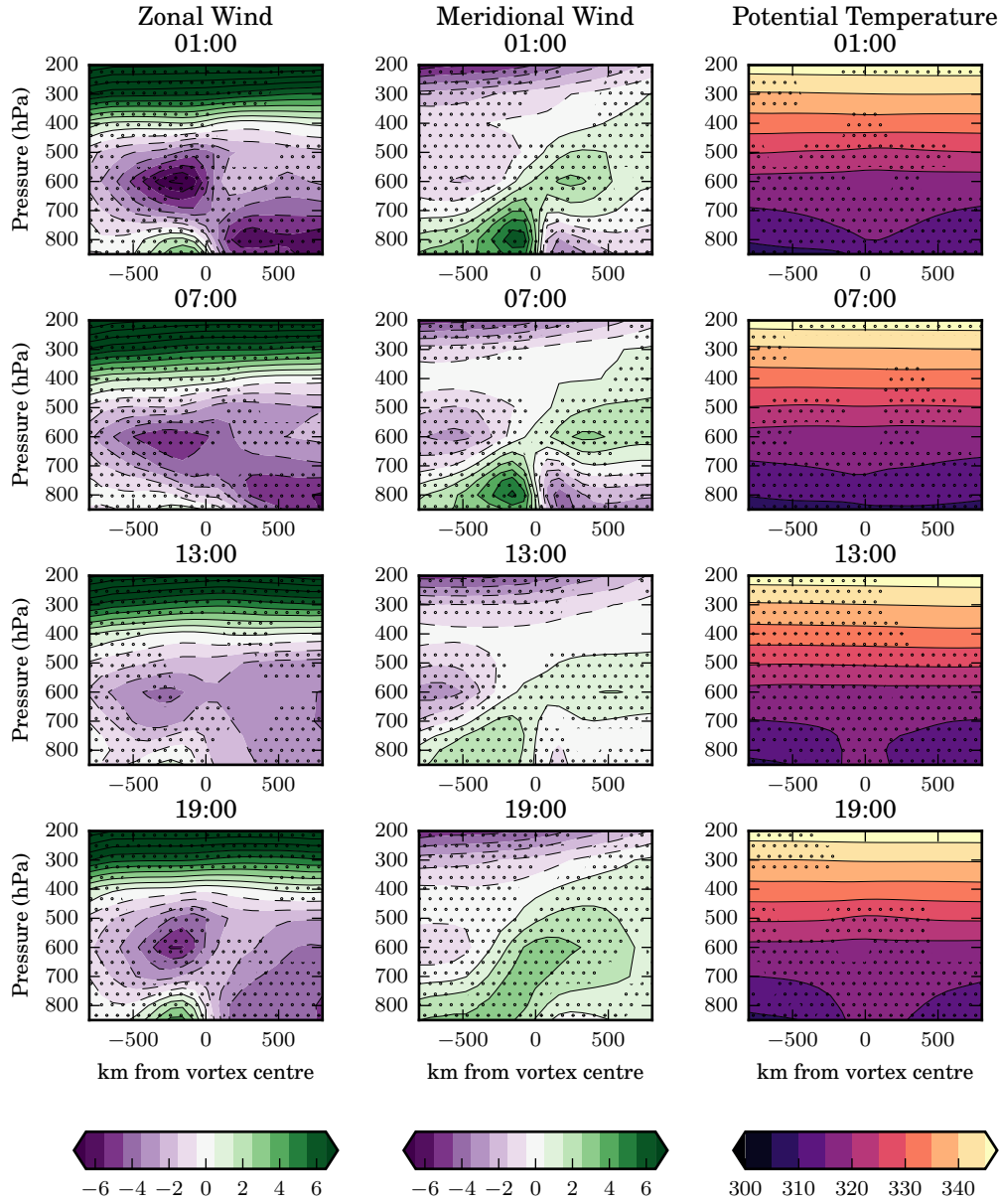


Figure 3.5. Composite west-east cross-sections with height of zonal wind (left column), meridional wind (centre) and potential temperature (right column) for heat low events (see text for definitions) at 01:00 (top row), 07:00 (second row), 13:00 (third row) and 19:00 (fourth row). Stippling shows the statistically significant points using a threshold of $p_{FDR}^* = 0.037$.

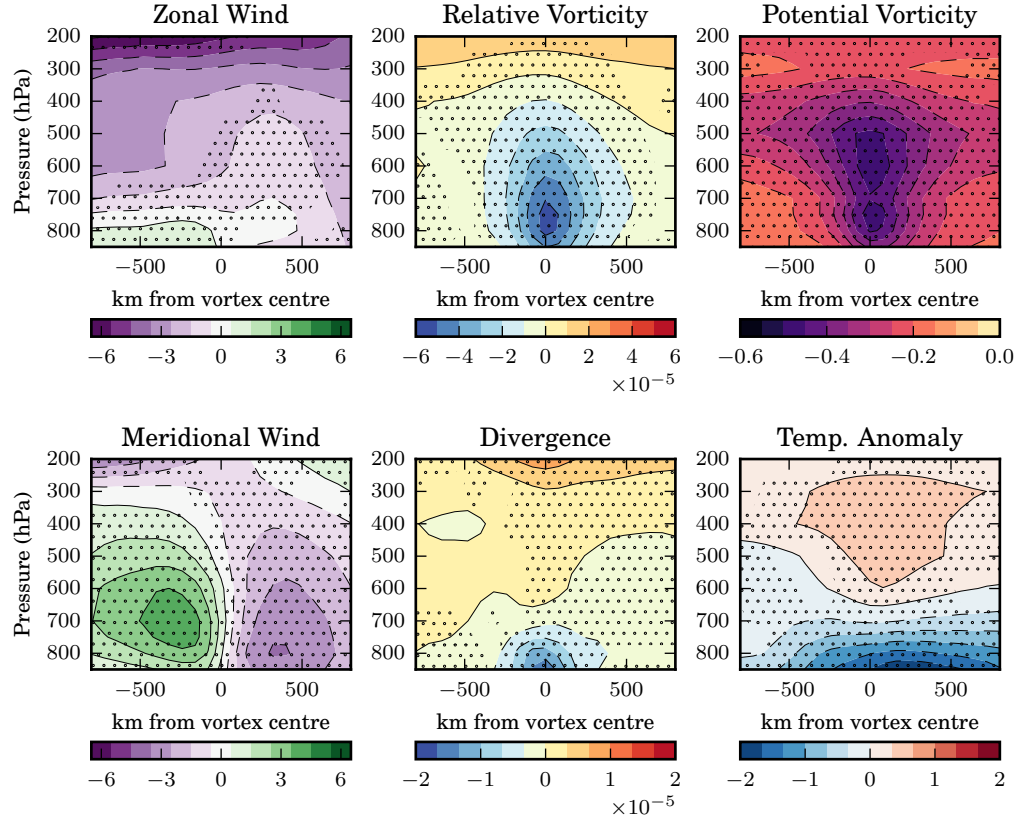


Figure 3.6. Composite vertical west-east cross-sections with height of zonal wind (top left), meridional wind (top right), relative vorticity (centre left), divergence (centre right), potential vorticity (bottom left), and temperature anomaly (bottom right) for tropical low events (see text for definitions). Stippling shows the statistically significant points using a threshold $p_{FDR}^* = 0.040$.

rarely move more than 5 degrees. The heat lows appear to be geographically locked and form only over the Angola region. By contrast tropical lows, shown in blue, travel east and west across the African continent. However, these circulation features linger in the region of interest, appearing to become anchored at around 20°E . This is at odds with tropical low-pressure systems observed in Australia and India, which are predominantly transient systems (Hunt et al. (2016), Berry et al. (2011) and Hurley and Boos (2015)). Although a small number of tropical

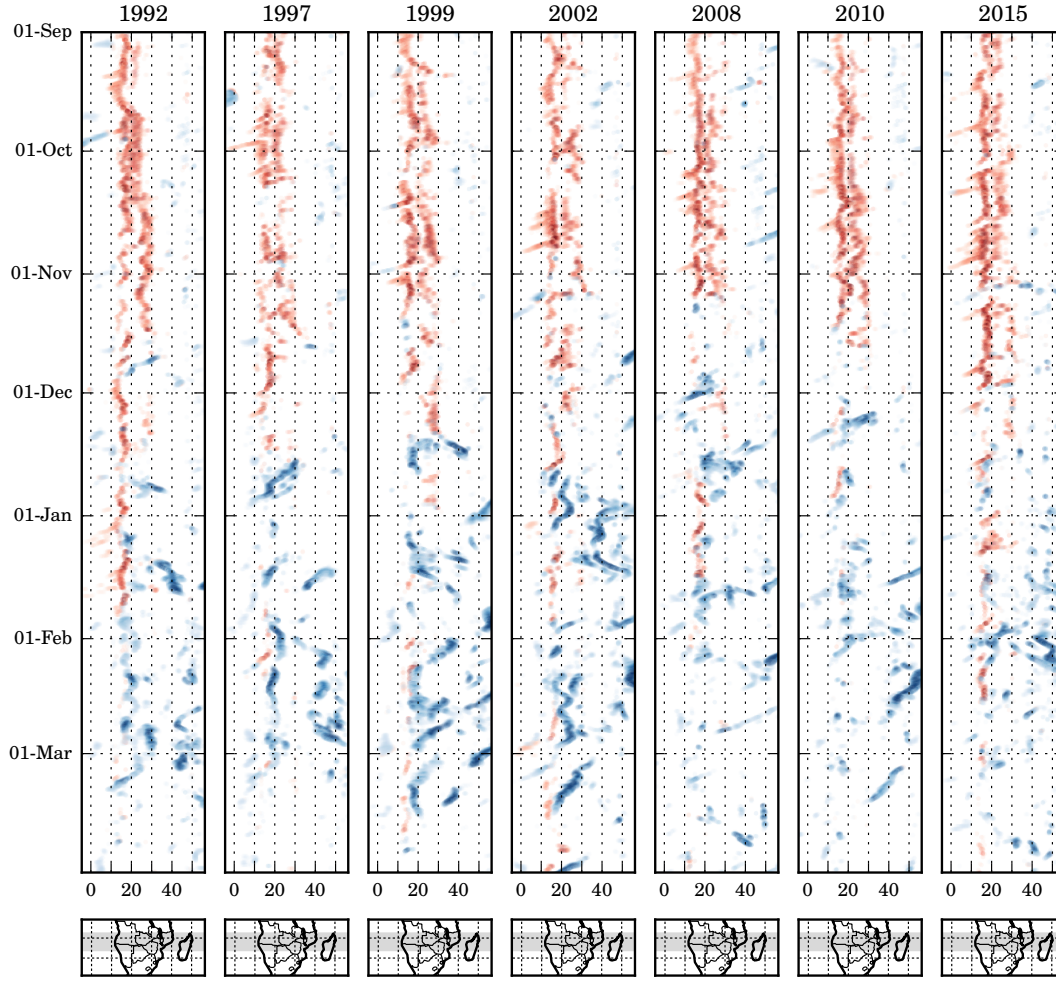


Figure 3.7. Top row: Longitude - time plots of the locations where cyclonic circulations ($\zeta < -4 \times 10^{-5} \text{ s}^{-1}$) have been identified in selected years between 11 and 18°S. Red dots: Neutrally stratified with $\frac{\partial\theta}{\partial z} < 0.0033 \text{ Km}^{-1}$ at 700 hPa, Blue dots: stably stratified with $\frac{\partial\theta}{\partial z} > 0.0033 \text{ Km}^{-1}$ at 700 hPa. Colour intensity represents cyclonic vorticity. Years shown are, from left to right, 1992 - 1993, 1997 - 1998, 1999 - 2000, 2002 - 2003, 2008 - 2009, 2010 - 2011 and 2015 - 2016. The bottom row shows a map of southern Africa with the domain of the above panel coloured in grey and is provided for context.

lows form in the Atlantic ocean, they only rarely cross either east or west across the West African coast. This behaviour is reflected across the all the years in the study period from 1979 to 2015 (not shown).

In the dry El Niño summer of 2015 - 2016, the Angola Heat Low lasted well into February (Figure 3.7). Meanwhile, the moist circulation features rarely reached the Angola region at all. By contrast, the wet El Niño summer of 1997 - 1998 featured numerous semi-stationary tropical lows in the Angola Low region. The extremely wet La Niña summer of 1999 - 2000 featured a large number of tropical low events tracking across the African continent from December onwards, many of which lingered in the Angola region. Of particular note is ex-Tropical Cyclone Eline, which penetrated mainland Africa in late February after crossing the Indian Ocean and reached 20°S (Reason and Keibel, 2004). However, in the drier La Niña summer of 2010 - 2011, although tropical low events were identified over southern Africa, none persisted in Angola for over a week. Inspired by these qualitative observations, the next section aims to clarify the interannual relationship between the phases of the Angola Low, ENSO and precipitation.

3.5. The Angola Low and precipitation

On an interannual timescale, the Angola Low is believed to have an important connection to regional precipitation across southern Africa (Mulenga, 1998). This may have a modulating impact on the relationship between southern African precipitation and the El Niño Southern Oscillation (ENSO). The El Niño phase of ENSO is typically associated with drought in southern Africa, a result of the shift in the Walker Circulation. The 1982 - 1983 and 2015 - 2016 El Niños both occurred in years where the Angola Low was weak, and resulted in severe drought. The 1997 - 1998 El Niño, however, coincided with a strong Angola Low and a drought was not observed.² The differences between these El Niño

²These three summers featured the strongest El Niño events of the study period, with average Niño 3.4 SST indices from November to March respectively 2.15, 2.28 and 2.14. The November - March southern African GPCP precipitation anomaly for the two drought summers was over

Table 3.1. Regression statistics from the regressions of the annual Angola Tropical Low Index (ATLI), Angola Heat Low index (AHLI) and combined Angola Low Index (ALI) on the Niño3.4 SST Index.

	ATLI	AHLI	ALI
R^2	0.13	0.04	0.03
N	37	37	37
Constant Coefficient	0.48	0.45	0.54
Constant Standard Error	0.04	0.03	0.03
Constant Coefficient P Value	<0.001	<0.001	<0.001
Niño 3.4 SST Coefficient	-0.08	0.03	-0.03
Niño 3.4 SST Standard Error	0.04	0.03	0.03
Niño 3.4 SST P Value	0.029	0.265	0.299

Table 3.2. Regression statistics for five regressions of southern African November - February southern African GPCP precipitation. (1): precipitation regressed onto ENSO, (2 - 4): partial regressions of the residual of precipitation onto the residuals of the ATLI, the AHLI and the ALI and (5): precipitation regressed onto the ATLI.

	Niño 3.4 Only	ATLI+Niño 3.4	AHLI+Niño 3.4	ALI+Niño 3.4	ATLI Only
R^2	0.52	0.26	0.00	0.13	0.35
N	37	37	37	37	37
Variable	ENSO	ATLI	AHLI	ALI	ATLI
Partial Regress?	No	Yes	Yes	Yes	No
Constant Coeff.	17.55	N/A	N/A	N/A	14.46
Constant S.E.	0.30	N/A	N/A	N/A	0.78
Constant P Val.	<0.001	N/A	N/A	N/A	<0.001
Variable Coeff.	-1.79	4.09	-0.53	3.96	6.34
Variable S.E.	0.29	1.15	1.60	1.68	1.47
Variable P Val.	<0.001	0.001	0.740	0.024	<0.001

summers have been well studied. Reason and Jagadheesha (2005) found that the interannual variability of the Angola Low modulates the rainfall impacts of ENSO. Lyon and Mason (2007) confirmed the role of the Angola Low and also found that high sea surface temperatures near Southern Africa and anomalous shifts in Walker circulation all contributed to the increase in precipitation in 1997 - 1998 as compared to 1982 - 1983.

1.5 standard deviations below the 1979 - 2015 mean, while the 1997 - 1998 precipitation anomaly was within 0.25 standard deviations of the mean.

The separation of the Angola Low into the Angola Heat Low and the Angola Tropical Low adds clarity to its relationship with ENSO and precipitation. An Angola Heat Low Index (AHLI) and an Angola Tropical Low Index (ATLI) have been created by counting the number of days per year when each class of Angola Low has been identified from November to March, and normalising such that the maximum value of the index is 1. The normalised sum of the AHLI and the ATLI is referred to as the Angola Low Index (ALI). It may be expected that the indices are anti-correlated, since both indices will be dependent on the date of the transition from the Angola Heat Low to the Angola Tropical Low. If the transition is earlier (later) than normal, then there will be more (fewer) tropical low days and less (more) heat low days. However, we found that this anti-correlation was in fact very weak, with $R=-0.12$.

Table 3.1 shows the results of linear regressions of the ATLI, the AHLI and the ALI onto the November - March average Niño 3.4 SST index. The AHLI is not dependent on Niño 3.4 SST ($R^2=0.04, p=0.27$). The ATLI has an R^2 coefficient 0.13 ($p=0.03$). However, the regression parameter of the Niño 3.4 SST index switches sign. As a consequence, the ALI does not exhibit a significant dependence on the Niño 3.4 SST index ($R^2=0.03, p=0.30$). This suggests that considering the Angola Heat Low and the Angola Tropical Low as a single feature obscures the relationship between the Angola Low and ENSO.

Average GPCP V2.3 November to March precipitation over southern Africa (south of 15°S) was regressed first against the Niño 3.4 SST index alone, and partial regressions were performed on the residual precipitation against the residual Angola Low indices, with regression statistics displayed in Table 3.2. Regression statistics of precipitation on the ATLI index alone are also shown. Niño 3.4 SST

alone was found to explain 52% of the variance ($p < 0.001$), and the partial regression onto the ATLI explained a further 26% of the variance ($p = 0.001$). However, the partial regression onto the AHLI did not increase variance explained and was not significant at the 0.05 level. This suggests that it is the Angola Tropical Low, and not the Angola Heat Low, which modulates the impact of ENSO on southern African precipitation, and that combining the effects of the Angola Tropical Low and the Angola Heat Low adds noise to this signal ($R^2 = 0.13$, $p = 0.024$).

Figure 3.8 shows scatter plots of the variables used in the first two regressions described in Table 3.2, with colours representing the calculated and predicted GPCP precipitation per summer over southern Africa. The coloured dots show the GPCP precipitation per summer while the coloured lines show the predictions based on the respective regressions. The black line indicates the regression of the ATLI on the Niño 3.4 SST index.

Adding the ATLI as a variable in the regression (Figure 3.8, lower panel) explains the variation among the three strongest El Niño events, in contrast to the ENSO only regression (Figure 3.8, upper panel). Furthermore, the difference between the strong La Niña summers starting in 1988, 2007 and 1999 is also explained by the inclusion of the ATLI in the regression. Neither regression predicts the precipitation of 1994 and 2005, which both had strong rainfall anomalies but occurred during the neutral ENSO phase and featured moderate tropical low indices.

These regression results imply that the component of the variation in the Angola Low that is independent of ENSO is correlated to the summer mean precipitation across southern Africa. We do not attempt to further examine this correlation here, or make any statements regarding causation or modes of variability. However, we note that future attempts to characterise the modulation

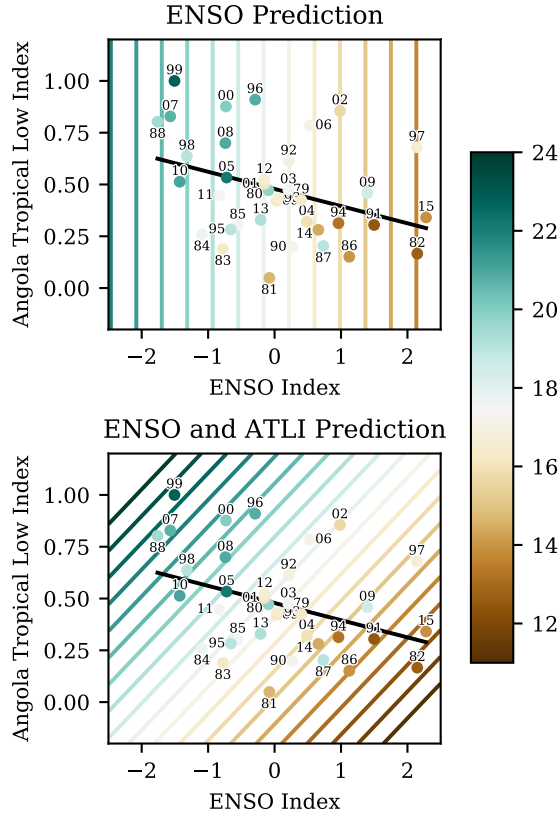


Figure 3.8. Scatter plots of November - March Niño 3.4 SST index and ATLI. Point colours represent GPCC precipitation over African mainland south of 15°. Line colours represent (top) precipitation predicted by the Niño 3.4 SST index only regression (column 1 of Table 3.2) and (bottom) precipitation predicted by the Niño 3.4 SST and ATLI regression (column 2 of Table 3.2). The black line shows the predicted ATLI for each value of the Niño 3.4 SST index (column 1 of Table 3.1). Labels indicate the year in which each season starts, e.g. 99 for 1999 - 2000.

of precipitation variability by the Angola Low should take the separation of the Angola Heat Low and the Angola Tropical Low into account.

3.6. Anchoring processes

Section 3.4 demonstrated that the climatological Angola Low is the combined effect of a series of heat lows and tropical lows. The heat lows tend to form and remain exclusively over the Angola Low region. In contrast, tropical lows track across tropical southern Africa, but linger over east Angola. Therefore, tropical lows are more likely to persist in east Angola than elsewhere. If the tropical lows instead tracked away from Angola as quickly as they track towards it, then the climatological depth of the Angola Low would be diminished. Thus,

the placement of the Angola Low in the late summer climatology originates from the behaviour of these transient synoptic-scale systems. Throughout the tropics, moist convecting lows are generally transient features and do not usually exhibit the stationary behaviour of the tropical lows in the Angola Region. This section therefore aims to discover why southern African tropical lows behave in this manner.

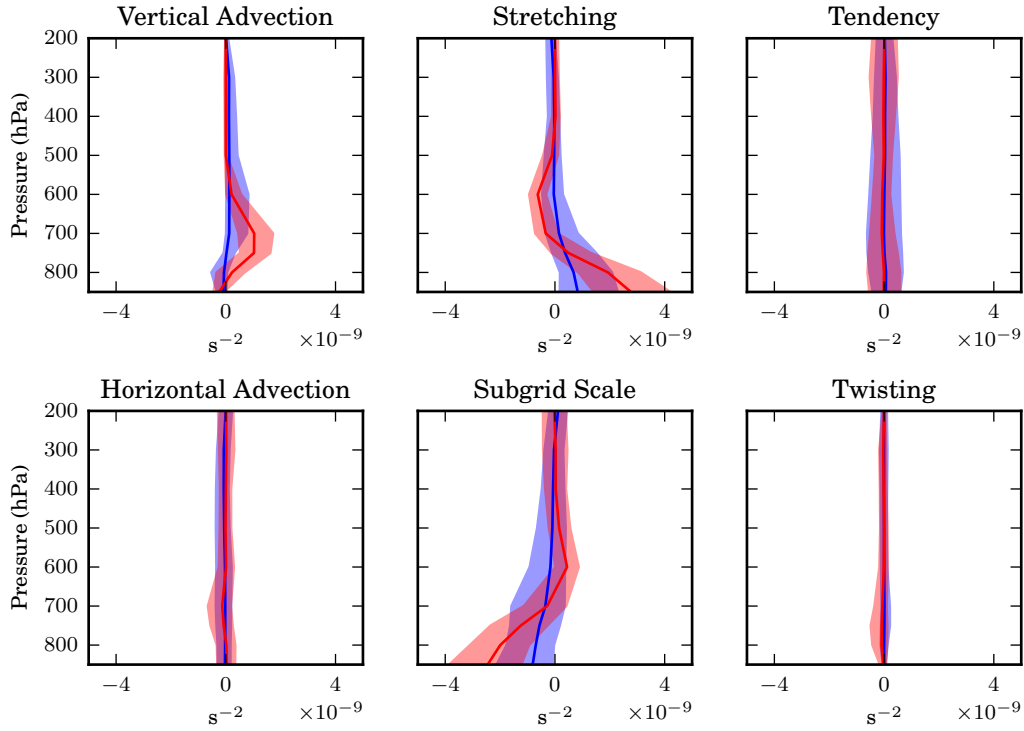


Figure 3.9. Vertical profiles of vorticity budget terms during Angola Low events. Red: heat lows, Blue: tropical lows. The solid line indicates the composite mean, and the coloured regions represent one standard deviation either side of the mean. The vorticity budget terms are labelled as per Equation 1, and the subgrid scale term represents friction.

An analysis of the vorticity budgets of the Angola Low phases was carried out in order to explain the motion and structure of the lows. Equation 1 shows the vorticity budget in the form that has been studied. Here, ζ is relative vorticity,

$\mathbf{v}$ and $\mathbf{v}_h$ are the 3D and 2D velocities respectively, ω is vertical velocity in pressure coordinates and $\mathbf{F}$ is the friction term of the momentum equation. This balance indicates that the possible sources and sinks of vorticity are advection, stretching, twisting and friction. The friction term cannot be directly computed from resolved model variables, and so is represented by a subgrid scale residual term.

$$(1) \quad \underbrace{\frac{\partial}{\partial t}(\zeta)}_{\text{tendency}} + \underbrace{\mathbf{v} \cdot \nabla(\zeta + f)}_{\text{advection}} + \underbrace{(\zeta + f) \nabla_h \cdot \mathbf{v}_h}_{\text{stretching}} - \underbrace{\hat{k} \cdot \left(\frac{\partial \mathbf{v}}{\partial p} \times \nabla \omega \right)}_{\text{twisting}} - \underbrace{\hat{k} \cdot (\nabla \times \mathbf{F})}_{\text{subgrid scale/friction}} = 0$$

Vertical profiles of the terms of the vorticity budget at the centroids of both heat lows (red) and tropical lows (blue) are shown in Figure 3.9. For both phases of the Angola Low, the largest source term in the budget is the stretching term, and the largest sink term is friction. This implies vorticity is created by the amplification of cyclonic absolute vorticity in a convergent airmass. Convergence may amplify either relative or planetary vorticity, and may be decomposed into two terms, $\zeta \nabla_h \cdot \mathbf{v}_h$ and $f \nabla_h \cdot \mathbf{v}_h$ to reflect this. The majority of this cyclonic acceleration is balanced by an opposing frictional force, but some fraction of it contributes to increasing the cyclonic vorticity of the system.

The dominant role of the stretching term is consistent with the general theory of cyclonic vortices on a rotating plane. A low-pressure anomaly is associated with uplift and convergent inflow, which is rotated by the Coriolis force to create cyclonic vorticity. However, a closer analysis of the stretching term indicates that a convergent anabatic sea breeze provides a second order source of stretching which may play a role in anchoring tropical lows to the Angola region. Figure

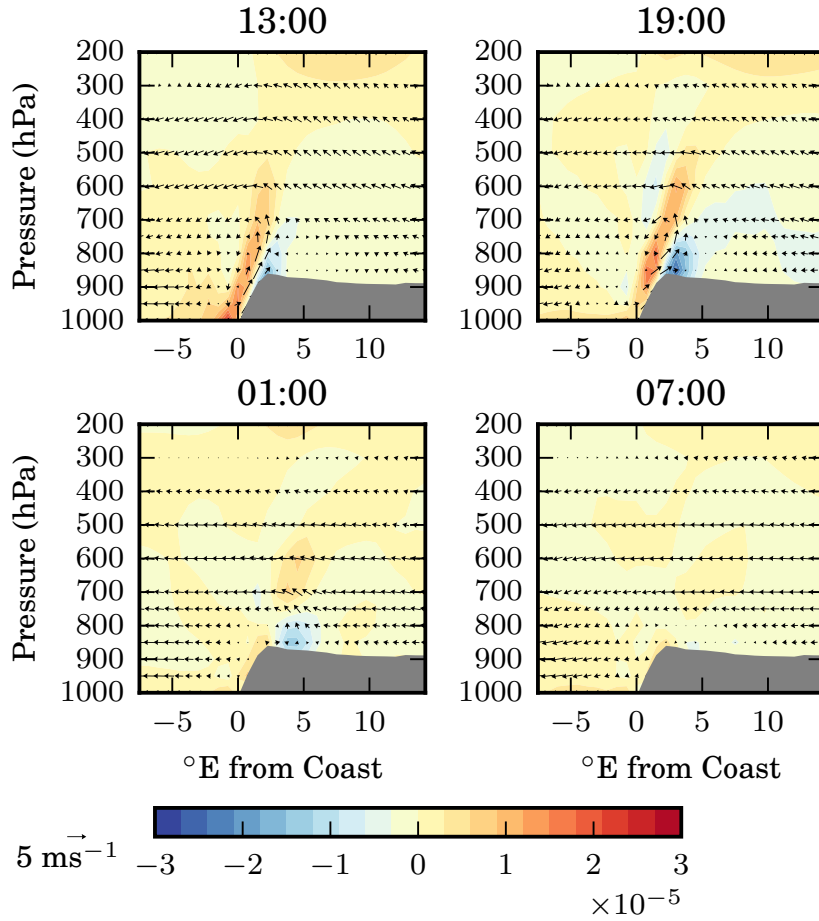


Figure 3.10. November - February climatology diurnal winds and horizontal divergence across the Angola Coast at 11 - 19°S. Vectors show zonal and vertical winds, red colours divergence and blue colours convergence. The times of the day for each panel are: 01:00 (top left), 07:00 (top right), 13:00 (bottom left) and 19:00 (bottom right). The x -axis is degrees of longitude East of the coastline.

3.10 shows the vertical cross-section of the anabatic sea breeze as it crosses the coastline at 11 - 19°S. From this it is apparent that the anabatic sea breeze initiates at midday and then advects inland. As it crosses the coastline, the anabatic sea breeze rises up the escarpment and continues its trajectory upwards such that its presence is apparent up to 600 hPa.

As it approaches the coast and proceeds upwards, the zonal wind strengthens and hence diverges, causing a plume of divergence (coloured red) rising from the ocean. The direction of the wind ensures that this plume is directed upwards and eastwards, and rises up over the plateau. The anabatic sea breeze slows down due to friction directly above the land surface, causing horizontal convergence (coloured blue). This alternating pattern of divergence and convergence, also reflected in vertical and onshore winds, resembles a topographically generated gravity wave. A second trough of convergence is faintly visible at 19:00, centred at around 500 hPa. Throughout the course of the night, the gravity wave is advected inland by its own surface winds, and steadily decays. By 01:00 it is apparent 5° east of the coastline. The anabatic sea breeze is present in the diurnal climatology every month of the year (not shown), although it is strongest in November when the surface heating is greatest. The surface convergence has the capacity to generate vortex stretching, which can invigorate low-pressure systems located in the same region. Meanwhile, the divergence above the boundary layer would generate negative vortex stretching and inhibit the convection.

Figure 3.11 shows the six hourly climatological surface irrotational winds averaged from November to February. The main feature that is apparent is the westerly anabatic sea breeze blowing across the south west African coastline. The blue colours along the west coast in the left column represent the location of the surface convergence maximum associated with the anabatic sea breeze. The red colours in the right column represent the associated divergence maximum higher in the atmosphere. Both the convergence and divergence zones form bands stretching along the western and southern coastlines, which move inland overnight. By 01:00, two regions of strong convergence remain: one at 16°E, 16°S

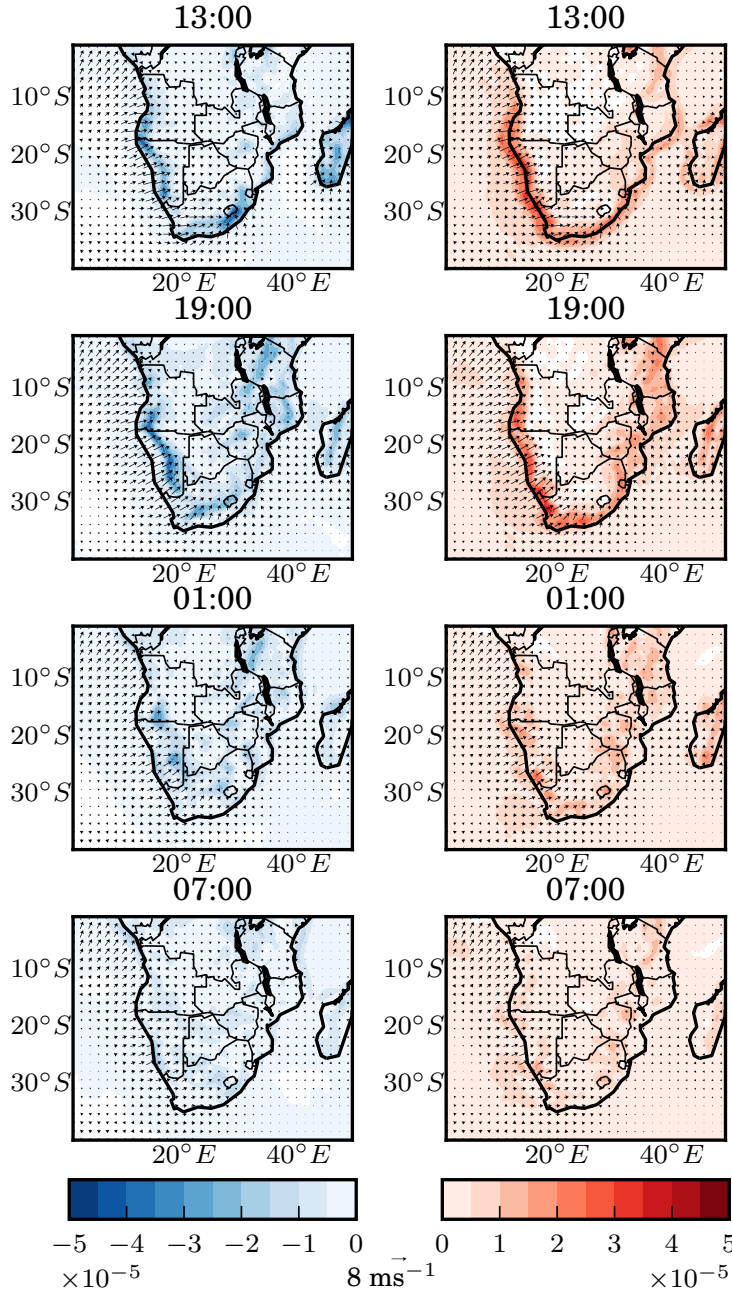


Figure 3.11. Irrotational component of diurnal surface winds in the November to February climatology with column maximum convergence (left) and divergence (right) across southern Africa. The times of the day for each panel are: 13:00 (first row), 19:00 (second row), 01:00 (third row) and 07:00 (fourth row).

and the other at 18.5°E, 24°S. Based on Figure 3.4, the former is a preferred location of both Angola heat lows and tropical lows. The latter is coincident with the Kalahari Heat Low. The divergence zone in Figure 3.11 at 19:00 - directly east of the Angola coast - is completely devoid of heat lows and tropical lows

in Figure 3.4. This suggests that the surface convergence and mid-tropospheric divergence of the anabatic sea breeze does indeed influence the placement of the Angola Low in the climatological average.

Because the centroids of the Angola lows are variable and the location of the coast is fixed, it is difficult to compare the stretching due to convergent inflow and the anabatic sea breeze in the same reference frame. We solve this problem by compositing lows centred at a fixed distance from the coast. Figures 3.12 and 3.13 show cross-sections of longitude against pressure for both stretching terms of the vorticity budget of heat lows and tropical lows respectively at different times of the day. The full vorticity budgets of these composites are shown in the supplementary figures. The first two columns are composited over lows centred 5° of longitude east from the coast, while the lows composited in the second two columns were centred 8° degrees east from the coast. Because the divergence from the anabatic sea breeze travels approximately 6° inland (Figure 3.11), the anabatic sea breeze may be expected to influence the western set of lows, but not the eastern set. When performing significance testing on these composites, we tested the null hypothesis that each vorticity budget term was equal zero. This means that the alternative hypothesis would imply that a vorticity budget term was a significant source or sink of cyclonic vorticity. This was tested using the Student's t-test, with autocorrelation and false discovery rates controlled for as per the other regressions described in Section 3.3.

In order to unpack the influence of the anabatic sea breeze on the Angola Low, it is useful to consider the influence on the Angola Heat Low and Angola Tropical Low separately. As alluded to in section 3.4.2, the anabatic sea breeze is a fundamental component of the Angola Heat Low. The idealised heat low of RÁCZ and Smith (1999) featured convergent low-level sea breezes in the afternoon,

which were rotated by the Coriolis force into a cyclonic vortex overnight. The sea breezes of RÁCZ and Smith (1999) originated from all directions. However on a larger continent and in the presence of easterly trade winds, westerly sea breezes dominated (Spengler and Smith, 2008). The sea breeze is a consequence of the meso-scale temperature gradient between the hot land surface and the cold ocean to the west. It is therefore a mechanism through which direct thermal heating may be converted into vorticity. The heating of the easterly trade winds as they rise over the plateaus of southern Africa is also expected to play a key role in the formation of the heat lows. However, the sea breeze provides a large component of the convergent inflow which creates cyclonic vorticity through stretching.

In our study, the importance of the sea breeze to the heat lows is apparent in Figure 3.12. At 19:00, the cyclonic vorticity of the heat low is very small and the primary source of vorticity for lows near the coast in the eastern composite is planetary vorticity stretching associated with the sea breeze. By 01:00, this vorticity source has intensified the cyclonic vortex, and relative vorticity stretching has become an important term. By 07:00, the cyclonic vortex is still strong but the stretching terms are both greatly reduced, suggesting that the horizontal convergence has dropped (consistent with Figure 3.10). At 13:00, the vortex and both stretching terms are both weakened once again. Heat lows located further from the coast in the western composite experience a similar diurnal cycle, however the initial planetary vorticity stretching originates from a different source. Significantly, 63% of all heat lows occurred within 5 degrees of the coast.

The influence that the anabatic sea breeze has on the Angola Tropical Low is more complicated and requires further investigation. In the eastern composite of Figure 3.13, the main source of stretching comes from the relative vorticity convergence at 01:00 and 07:00. The signature of the anabatic sea breeze can be

seen in the planetary vorticity stretching term, however this influence is limited to within 6 degrees of the coast and does not impact on the cores of the tropical lows. In the western composite of Figure 3.13, both stretching terms contribute more cyclonic vorticity at the core of the tropical lows. Planetary vorticity stretching carries a strong signature of the anabatic sea breeze and contributes to a vorticity source at the cores of the tropical lows about half as strong as that contributed by relative vorticity stretching. Relative vorticity stretching is stronger in western composite than the eastern composite, which could be due to the anabatic sea breeze influence. Together, these results imply that vorticity stretching due to the convergence of the anabatic sea breeze can be a second order vorticity source for tropical lows centred within 6 degrees of the coast.

The vorticity sink associated with the divergent tail of the anabatic sea breeze is of similar order of magnitude as the vorticity source terms in every case. This divergence zone may act as a barrier to eastward propagating tropical lows and prevent them from crossing the coast into the Atlantic Ocean. The full vorticity budget of the western composite (see supplementary figures) also indicates that low-level cyclonic vorticity is advected inland from the western coast at 19:00 by the anabatic sea breeze. Therefore, the anabatic sea breeze may cause the tropical lows to linger in the Angola region, as can be observed in Figure 3.7 in section 3.4. This means that the climatology average contains a larger number of days featuring tropical lows. Therefore, the action of the anabatic sea breeze deepens the Angola Low and intensifies its cyclonic vorticity in the climatological average.

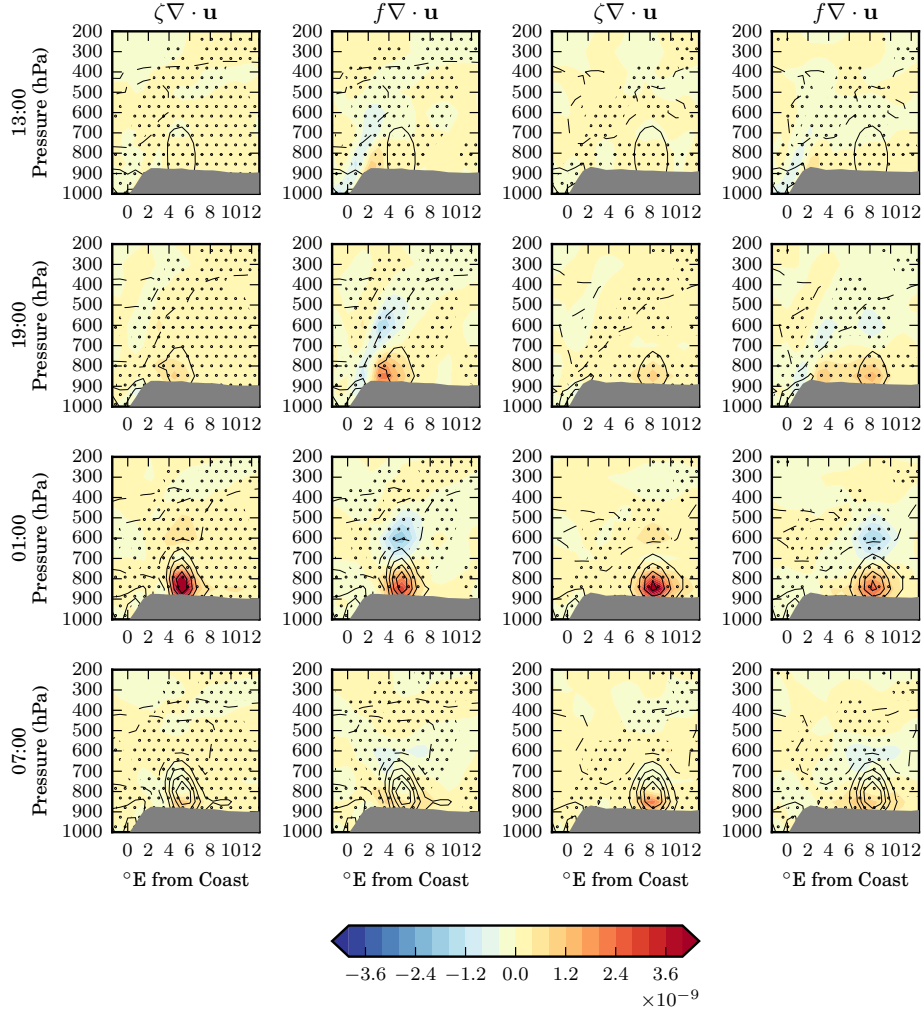


Figure 3.12. Vertical cross-sections of vorticity budget stretching terms at time 01:00 (first row), 07:00 (second row), 13:00 (third row), and 17:00 (fourth row). The first and third columns show stretching of relative vorticity, and the second and fourth columns show stretching of planetary vorticity. The first two columns show composites of heat low events located 5 degrees from the coast, while the second two show heat low events located 8 degrees from the coast. The cross-section is taken across the latitude of the vortex centres. The x -axis is °E of the coast. Black contours indicate the cyclonic vorticity ($2 \times 10^{-5} \text{ s}^{-1}$ contour interval). Stippling shows the statistically significant grid points, determined based on a threshold $p_{FDR}^* = 0.027$.

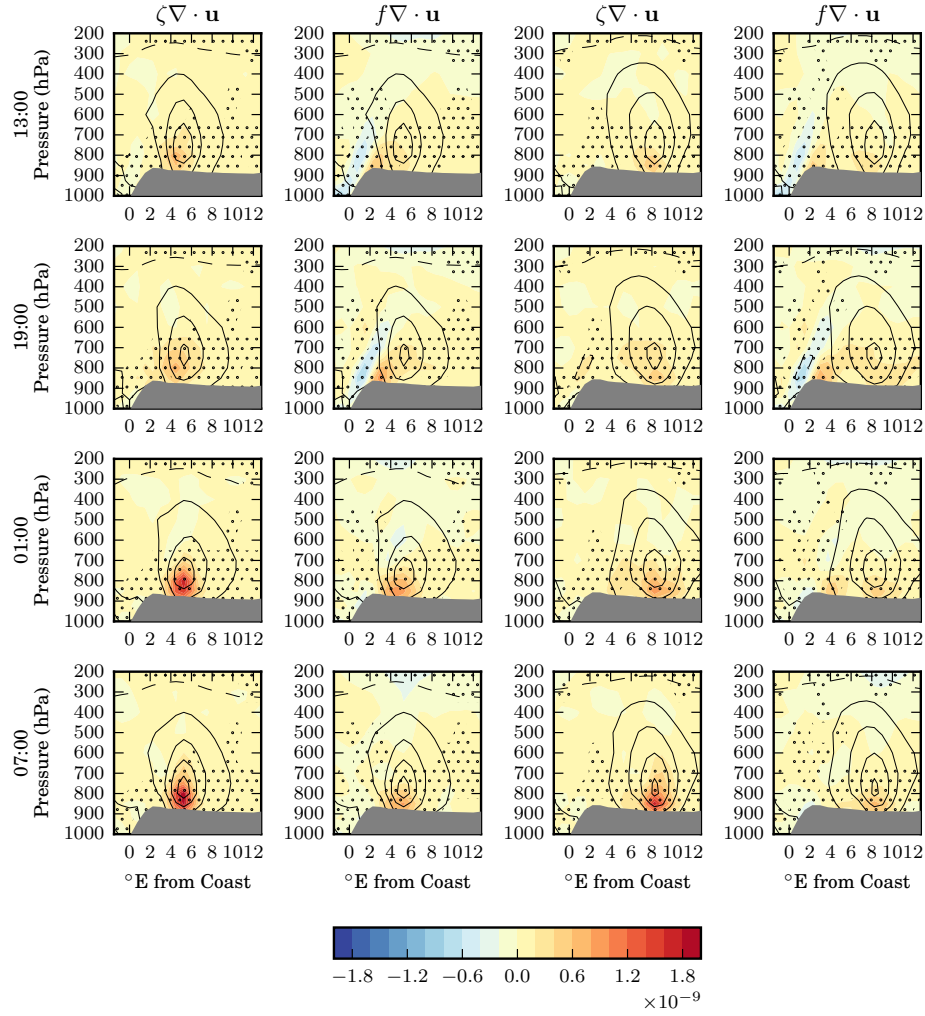


Figure 3.13. Vertical cross-sections of vorticity budget stretching terms at time 01:00 (first row), 07:00 (second row), 13:00 (third row), and 17:00 (fourth row). The first and third columns show stretching of relative vorticity, and the second and fourth columns show stretching of planetary vorticity. The first two columns show composites of tropical low events located 5 degrees from the coast, while the second two show tropical low events located 8 degrees from the coast. The cross-section is taken across the latitude of the vortex centres. The x -axis is °E of the coast. Black contours indicate the cyclonic vorticity ($2 \times 10^{-5} \text{ s}^{-1}$ contour interval). Stippling shows the statistically significant grid points, determined based on a threshold $p_{FDR}^* = 0.020$.

3.7. Discussion and conclusions

This chapter has shown that the Angola Low can be separated on a synoptic-scale into two distinct phases - the Angola Heat Low and the Angola Tropical Low. It was found that this distinction clarifies the link between the Angola Low, the precipitation and ENSO on an interannual timescale. The Angola Tropical Low is stronger during La Niña seasons. However, the relationship between ENSO and the Angola Low Indices was relatively weak and we found that the Angola Low undergoes considerable variability independent of ENSO. A partial linear regression of southern African precipitation on the Niño 3.4 SST index and an Angola Tropical Low Index was found to explain the large variance in precipitation during the three strongest El Niño summers in the study period, two of which were associated with severe droughts while a third experienced average rainfall. This regression also did well at explaining the variance between precipitation during the three strongest La Niña summers in the study period.

These regression results suggest that the Angola Tropical Low is important for southern African rainfall. The tropical low events were found to be dynamically similar to monsoon low-pressure systems which form throughout the tropical landmasses. However, the key difference between the southern African tropical lows and those observed elsewhere was their propensity to linger in the Angola region. This semi-stationary behaviour is fundamental to the impact that the Angola Tropical Low has on the climatological Angola Low. While each transient tropical low spends a 2 - 3 days directly impacting the weather of any given area, a semi-stationary tropical low may impact the weather for several weeks, building up a stronger influence on the seasonal climate.

Vorticity budget analysis has demonstrated that an anabatic sea breeze circulation plays an important role in anchoring the tropical lows to the Angola region. The impact of the sea breeze on the tropical lows is secondary to the processes which create the tropical lows and only acts as an anchoring mechanism, rather than a formation mechanism. The anabatic sea breeze was shown to be divergent in the mid-troposphere near the coast and convergent near the surface and further inland, which enhances the stretching vorticity budget term. This vorticity source strengthens the Angola Low inland and weakens it near the coast, inhibiting eastward tracking tropical lows from crossing the coast. An equivalent point of view is that the uplift of the eastern branch of the anabatic sea breeze enhances the convection of the tropical lows, while the subsidence associated with the westward branch of the anabatic sea breeze overturning inhibits convection. This overturning circulation can be clearly seen in Figure 3.10.

By considering the synoptic expression of the Angola Low, this chapter has revealed the mechanisms which drive it, namely heat lows, tropical lows and the anabatic sea breeze. This work opens up several avenues of future research. The processes that link the Angola Low to southern African precipitation, such as TTCBs and wet spells, should be studied taking into account the two phases of the Angola Low. A process-based analysis of the Angola Low in CMIP and AMIP models should examine how well the models represent the three mechanisms listed above. These findings will therefore support efforts to reduce uncertainty around future projections of southern African precipitation.

3.8. Acknowledgements

This work has been generously supported by the Origin Foundation John Monash Scholarship and by the NERC-DfID funded UMFULA (NE/ M020207/1)

program. The data used in this analysis were provided by the European Centre for Medium range Weather Forecasts (<http://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=pl/>), the Physical Science Division of the Earth System Research Laboratory (<http://www.esrl.noaa.gov/psd/>), the Climate Prediction Centre (<http://origin.cpc.ncep.noaa.gov/>) at the National Oceanic and Atmospheric Administration, and NASA (<https://disc.sci.gsfc.nasa.gov/daac-bin/FTPSubset.pl>). The authors would like to thank Callum Munday, Neil Hart and Sebastian Engelstaedter for their inputs, as well as Kerry Cook and two anonymous reviewers for their helpful comments.

3.9. Supplementary material

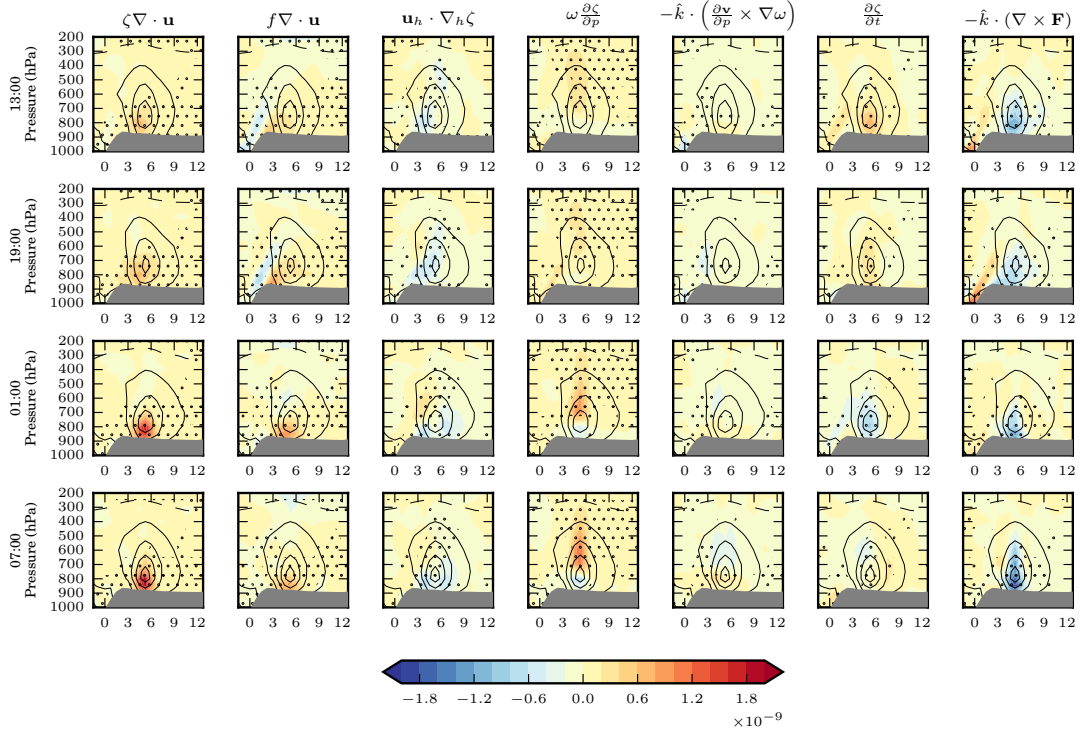


Figure 3.14. Vertical cross sections of vorticity budget terms at times (top row) 01:00, (second row) 07:00, (third row) 13:00, and (final row) 17:00 for tropical low events located 5 degrees from the coast, while the second two show tropical low events located 5 degrees from the coast. Stippling shows the statistically significant grid points, determined based on a threshold $p_{FDR}^* = 0.017$. The composite sample size is 143 days, with the average degrees of freedom per panel ranging from 86 - 121 depending on the autocorrelation of each variable. The x -axis shows distance from the coast in degrees east.

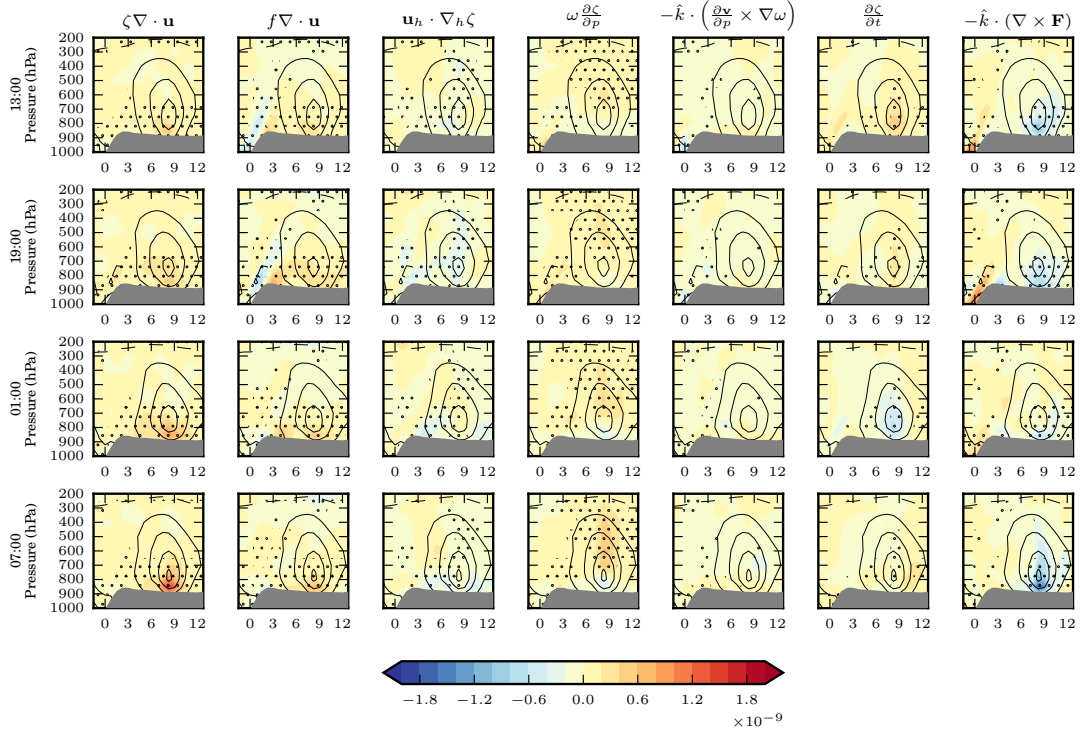


Figure 3.15. Vertical cross sections of vorticity budget terms at times (top row) 01:00, (second row) 07:00, (third row) 13:00, and (final row) 17:00 for tropical low events located 5 degrees from the coast, while the second two show tropical low events located 8 degrees from the coast. Stippling shows the statistically significant grid points, determined based on a threshold $p_{FDR}^* = 0.014$. The composite sample size is 109 days, with the average degrees of freedom per panel ranging from 66 - 113 depending on the autocorrelation of each variable. The x -axis shows distance from the coast in degrees east.

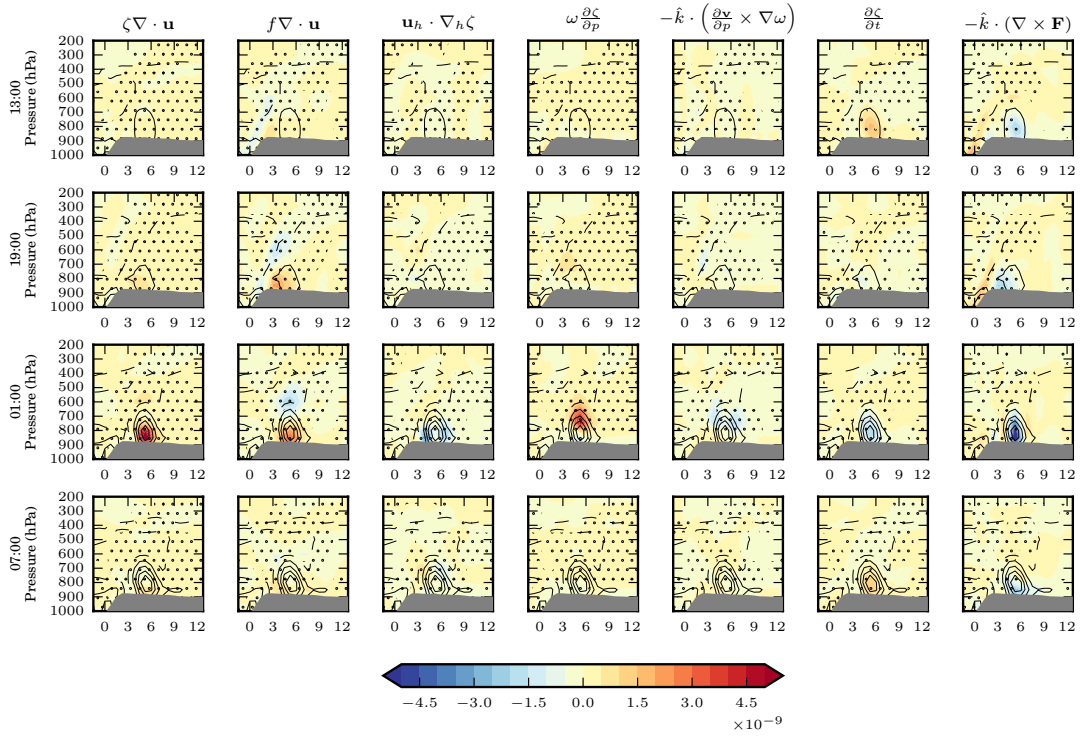


Figure 3.16. Vertical cross sections of vorticity budget terms at times (top row) 01:00, (second row) 07:00, (third row) 13:00, and (final row) 17:00 for heat low events located 5 degrees from the coast, while the second two show heat low events located 5 degrees from the coast. Stippling shows the statistically significant grid points, determined based on a threshold $p_{FDR}^* = 0.028$. The composite sample size is 143 days, with the average degrees of freedom per panel ranging from 86 - 121 depending on the autocorrelation of each variable. The x -axis shows distance from the coast in degrees east.

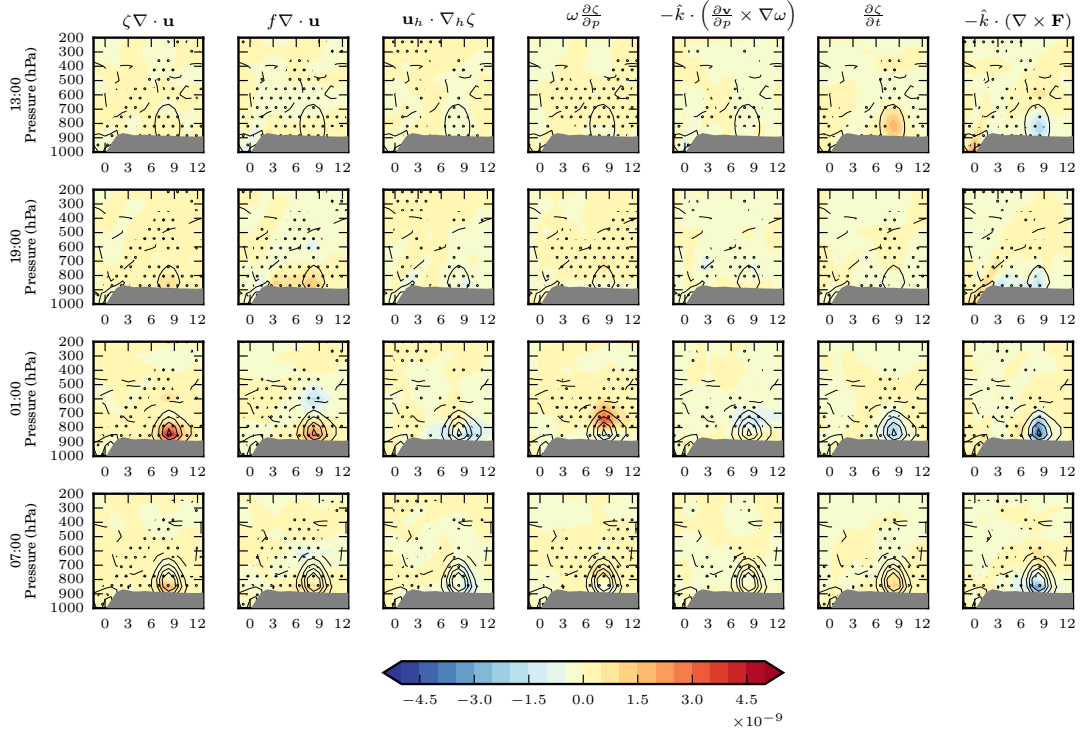


Figure 3.17. Vertical cross sections of vorticity budget terms at times (top row) 01:00, (second row) 07:00, (third row) 13:00, and (final row) 17:00 for heat low events located 5 degrees from the coast, while the second two show heat low events located 8 degrees from the coast. Stippling shows the statistically significant grid points, determined based on a threshold $p_{FDR}^* = 0.018$. The composite sample size is 109 days, with the average degrees of freedom per panel ranging from 66 - 113 depending on the autocorrelation of each variable. The x -axis shows distance from the coast in degrees east.

3.10. Afterword to Chapter 3

This chapter has shown that tropical lows in Angola are important synoptic events which influence interannual rainfall variability and form a component of the climatological Angola Low. Therefore, tropical lows are a form of synoptic weather system that meets the requirements of the overall goal of this thesis. Further analysis has provided an explanation for the semi-stationary behaviour of tropical lows, demonstrated a relationship between tropical low frequency and interannual rainfall variability, and demonstrated the spatial distribution of tropical lows. This addresses Aims 2, 3 and 4 of this thesis.

However, no attempt has been made in this chapter to consider tropical lows as time-evolving systems. Instead, the chapter focussed on centred composites consisting of individual tropical low days. The next step in this analysis is to track the motion and behaviour of tropical low events through time, as will be presented in Chapter 4.

3.10.1. The Angola Heat Low and the Congo Air Boundary

Before proceeding, the role of the AHL in the remainder of this thesis must be addressed. Although the AHL is not addressed directly in the aims and goals of this thesis, it is fundamentally linked to the Congo Air Boundary, which will be introduced in Chapter 5.

The relationship between the CAB and the AHL is demonstrated in Figure 3.18. This figure shows diurnal north-south cross sections of the AHL. This is analogous to Figure 3.5, which shows diurnal east-west cross sections of the AHL. In Figure 3.18, zonal and meridional winds show a similar structure of overnight cyclonic winds and low level inflow as the east-west cross sections showed. Similarly, the lower troposphere is neutrally stratified at midday and in the evening.

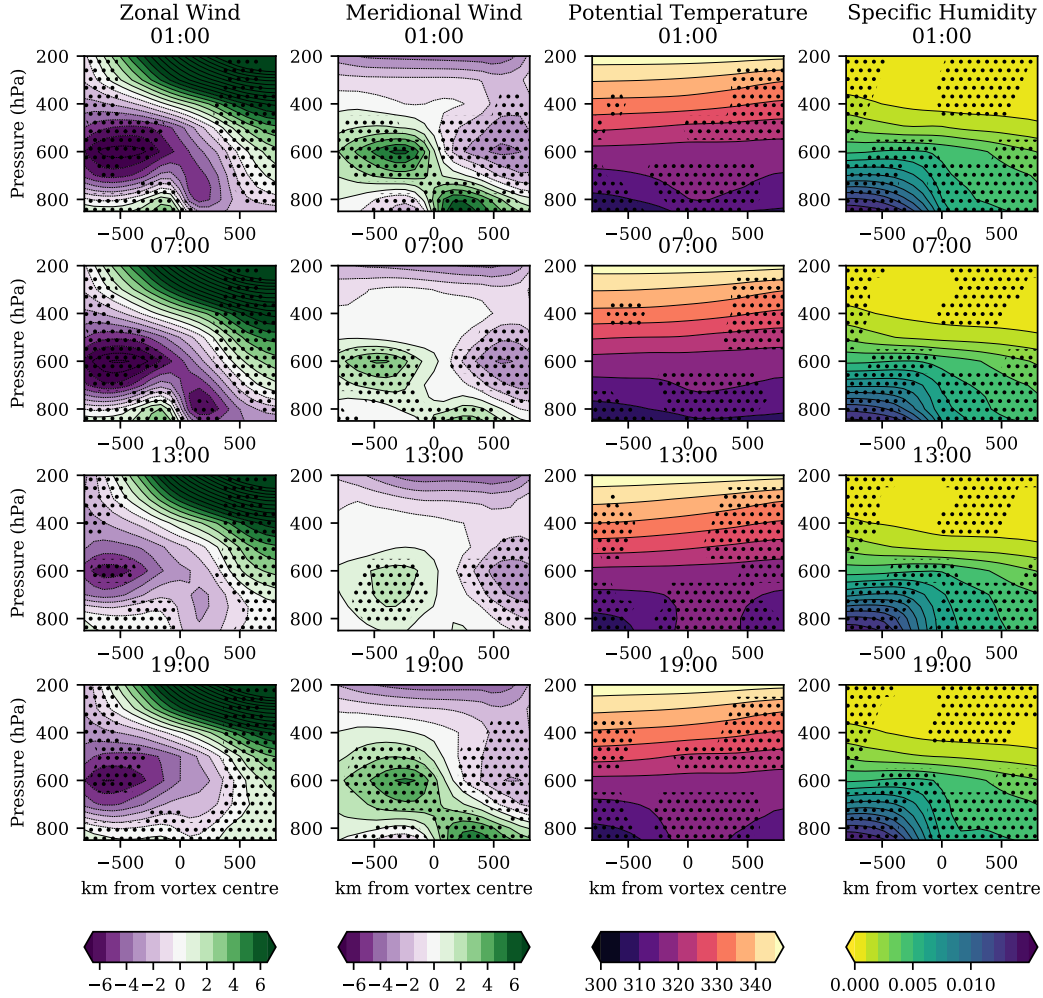


Figure 3.18. Composite north-south cross-sections with height of zonal wind (left column), meridional wind (second column), potential temperature (third column) and specific humidity (right column) for heat low events (see text for definitions) at 01:00 (top row), 07:00 (second row), 13:00 (third row) and 19:00 (fourth row). Stippling shows the statistically significant points using a threshold of $p_{FDR}^* = 0.024$.

Immediately to the north of the heat low, there is a sharp gradient in specific humidity and thermal stability. Subsequent to the publication of this chapter, this gradient was identified as the CAB, which will be addressed in Chapter 5. Therefore, although the heat low has not been directly selected as one of the

synoptic events for consideration in this study, it is indirectly studied when the CAB is considered later on. A similar relationship exists between the Kalahari Heat Low and the KD, also presented in Chapter 5.

CHAPTER 4

Tropical lows in southern Africa: tracks, rainfall contributions and the role of ENSO

Southern African tropical lows are synoptic-scale cyclonic vortices that propagate westward across southern Africa in the Austral summer. They strongly influence local rainfall and aggregate in the climatological DJF mean to form the Angola Low. In this study, tropical lows are identified and tracked using an objective feature tracking method. On average, 35 events are detected between November and March per year. The statistics of tropical low tracks over southern Africa are presented and compared across three reanalysis products. Findings are compared to the literature of tropical low pressure areas elsewhere in the world, where it is found that most tracking statistics compare well, but that the tendency of tropical lows to become semi-stationary over Angola is unique to southern Africa. The hypothesis that tropical lows in Angola have a causal relationship with Tropical Temperate Troughs is tested, and a correlation between occurrence frequencies is found at interannual but not daily timescales. Precipitation is attributed to the tropical lows and it is found that tropical lows are associated with 31% of rainfall across tropical southern Africa, based on gridded precipitation products. The interannual variability of the number of tropical lows that form per year ($\sigma = 6$ events/annum) is linked to ENSO and the tropical easterly jet. The mean latitude of tropical lows is shifted northwards during El Niño and southwards during La Niña. Much of the interannual precipitation variability maximum in Angola is

attributed to rainfall associated with tropical lows. These results provide insights into the southern African response to ENSO and into the mechanisms of rainfall in the southern African tropical edge.

4.1. Foreword to Chapter 4

In this chapter, a fuller description of the characteristics of tropical lows is provided. A tracking algorithm is developed to identify these weather systems in reanalysis datasets, fulfilling Aim 1 with respect to tropical lows. This algorithm is then used to characterise the behaviour, spatial distributions and frequencies of tropical lows in reanalysis datasets (Aim 2). Rainfall is attributed to individual tropical low events by assuming that rain that falls within 5° of a track path is associated with a tropical low. This allows Aim 4 to be precisely addressed, as the total tropical low and non-tropical low rainfall is aggregated over each season and the interannual variability of each component is analysed. Tropical lows are found to have a strong influence over interannual rainfall variability. In the interior of the continent (west of about 27° E in Angola, Namibia and Western Zambia), they are found to deliver the majority of the ENSO signal.

Further analysis linking the results of Chapter 4 to those of Chapter 3 is included in the afterword. This puts the regression presented in Figure 3.8 into the context of the decomposition of rainfall into tropical low and remainder rainfall. We find that the interannual remainder component of rainfall is not correlated to the tropical low index. Thus, the Angola Low modulates rainfall by the direct effect of its rainfall from associated tropical lows, and not indirectly by remote moisture fluxes. This chapter was accepted for publication in the Journal of Geophysical Research (Atmospheres) in September 2019.

4.2. Introduction

The southern ‘tropical edge’ of the African rain belt (16° - 22°S) is a hotspot of rainfall variability. The tropical edge spans 8 countries and its precipitation feeds the Okavango Delta and the Zambezi river basin. The former is a hotspot of biodiversity (Ramberg et al., 2006), while the latter supports an economy of 30 million people and has been extensively dammed for hydroelectric plants (World Bank, 2010). Quantification of the behaviour and rainfall contributions of the local synoptic processes provides important context for understanding interannual variability and future change of regional precipitation.

Tropical lows, mid-level cyclonic vortices that form in the tropical easterlies, are one such process that contribute rainfall in the tropical edge. For the purposes of this study, we define the tropical edge as a zonal band spanning 12° - 22°S , from the west to the east coast of southern Africa. Tropical lows are well studied across the world, including in India (Godbole, 1977; Hunt et al., 2016; Hurley and Boos, 2015), northern Australia (Kilroy et al., 2017; McBride and Keenan, 1982) and the low-latitude oceans (Wallace, 1971; Fine et al., 2016). When conditions are favourable, tropical lows located over oceans may intensify into tropical cyclones, and so much of the research into tropical lows in general has focussed in regions where this occurs, motivated by increasing the predictability of extreme weather risk (eg: (Smith and Montgomery, 2015)). However, while tropical lows that exist over land and do not intensify are much less catastrophic than tropical cyclones, they are also more frequent, and so their primary importance to the regional population shifts from being a class of extreme weather events to a source of precipitation (Lavender and Abbs, 2013; Hunt and Fletcher, 2019).

The track statistics of these weather systems and their local contributions to rainfall have been studied across the tropics. In the south Asian monsoon, monsoon depressions are an important source of rainfall in north and west India (Hunt et al., 2016a). Typically, 2-5 systems form each summer, and tend to track westward along the southern edge of the Himalayas (Hunt et al., 2016). African easterly waves, which form in north Africa, have also been intensely studied (Burpee, 1974; Hsieh and Cook, 2005; Thorncroft and Hoskins, 1995; Gomes et al., 2019; Hopsch et al., 2007), but are significantly larger and more frequent than southern African tropical lows. In northern Australia, closed low pressure systems contribute up to 60% of precipitation in the Pilbara, and the proportion of precipitation that is attributable to tropical lows increased between 1989 and 2009 (Lavender and Abbs, 2013). Berry et al. (2012) identified monsoon weather systems in northern Australia using the 315-K isotherm of isentropic potential vorticity, and found that these systems track westward at 6 ms^{-1} and form on average every 2.5 days in the wet season. Hurley and Boos (2015) conducted a global study of land-based depressions in the tropics, and found similar results over northern Australia and India. Chapter 3 found that the radii and vertical structures of Australian and Indian monsoon lows were qualitatively comparable to those of tropical lows in southern Africa.

The prominence of tropical lows over southern Africa was first described by Taljaard (1953), who noted that these lows are quasi-stationary in the Angola region, and are associated with low level convergence. In the seasonal and climatological means, these systems, along with dry and shallow heat lows, aggregate to create a local pressure minimum which is called the Angola Low (Munday and Washington, 2017, and Chapter 3). In recent years, the Angola Low has become well known due to statistical correlations to rainfall variability, and to Tropical

Temperate Trough (TTT) formation on the interannual timescale. Reason et al. (2006) first suggested that the Angola Low modulates the relationship between the El Niño Southern Oscillation (ENSO) and southern African precipitation. ENSO is the main driver of precipitation variability in southern Africa, with the El Niño phase commonly associated with drought. The contrast between the very strong El Niño of 1998, when the Angola Low was strong and the forecast drought did not occur, and 2015, when the Angola Low was unusually weak and the summer rainfall was the lowest in recorded history, demonstrates this relationship (Blamey et al., 2018). Chapter 3 found that it is the contribution of tropical lows to the Angola Low, and not that of heat lows, which influences the relationship between rainfall and ENSO.

TTTs are elongated cloud-bands that form over southern Africa and provide a large proportion of the subtropical summer rainfall to the south of the tropical edge, where annual rainfall is typically low (Harrison, 1984). TTTs occur when an upper level ridge in the subtropical jet interacts with moist tropical convection from tropical southern Africa and allows convection to spread in a band across the continent (Macron et al., 2014). Since the tropical convection that TTTs are associated with is often in the Angola Low region, it has been hypothesised that this convection may be associated with an Angola tropical low (Todd and Washington, 1999). For example, Hart et al. (2010) identify the importance of tropical lows in three case studies, while Macron et al. (2014) highlight an increase in moisture flux around the Angola Low in lagged TTT composites. However, this relationship has not been studied on a synoptic timescale. Crétat et al. (2018) call for a better understanding of the relationship between the Angola Low and tropical-extra-tropical interactions on a daily timescale.

Despite being studied based on meteorological analyses before the satellite era (Taljaard, 1972), little work has been done since the availability of reanalysis datasets on systematically characterising the behaviour of tropical lows in southern Africa. While tropical lows are precipitating systems, there have been no attempts to determine the proportion of precipitation that may be attributed to tropical lows or their role in interannual rainfall variability, as compared to other systems such as localised convection, thunderstorms and mesoscale convective systems. In the intervening years, various tools have been developed to identify and follow atmospheric features in reanalysis datasets as they evolve and move through space. One such tool is TRACK, which was developed to study extratropical storms tracks (Hodges, 1994, 1999) and has since been applied to tropical cyclones (Hodges et al., 2017) and tropical lows (Hurley and Boos, 2015). Hurley and Boos (2015) found that on average 12.5 tropical lows formed each year in southern Africa, but their globally based approach did not provide detail on the nature of tropical lows over southern Africa. Other tracking algorithms have also been developed for similar purposes and our results may depend on the choice of tracking method (Neu et al., 2013).

This study applies TRACK in the southern African context of region-specific land-based tropical lows in order to determine their fundamental properties and rainfall consequences. We examine the differences and similarities between the representation of tropical lows in three different reanalysis products, and consider how this may reflect the true state of tropical lows in the real atmosphere. The following section discusses the datasets used in this study and elucidates the TRACK methodology. This chapter then proceeds to address the following questions:

- (1) What are the frequencies, lifetimes, intensities and preferred locations of the tropical lows which form and linger in southern Africa, as represented in reanalysis datasets (Section 4.4)?
- (2) What control is exerted by ENSO on the interannual variability of tropical lows (Section 4.5)?
- (3) Is there a link between tropical lows and TTTs on the synoptic timescale (Section 4.6)?
- (4) What is the proportion of southern African rainfall that is associated with tropical lows, and how much interannual rainfall variance does this rain account for (Section 4.7)?

Section 4.8, then examines how the results obtained from derived products such as reanalysis and gridded precipitation may reflect the true atmosphere and discusses implications for southern Africa. A sensitivity analysis of the tropical low detection thresholds is provided in Appendix A.

4.3. Data and methods

This study uses three reanalysis datasets and two gridded precipitation products. The reanalysis products used are MERRA2, ERA-Interim and the high resolution run of ERA-5. The precipitation datasets considered are CHIRPS and TAMSAT. We elected not to use TRMM due to its shorter data availability. Table 4.1 summarises the key features of each product.

4.3.1. Algorithm

The tracking algorithm described by Hodges (1994) was applied using a methodology based on Hodges et al. (2017), who use TRACK to identify tropical cyclones, as well as their antecedent and subsequent phases as tropical lows and

Table 4.1. Reanalysis and Gridded Rainfall Products

Name	Type	Spatial Resolution	Temporal Resolution	Temporal Coverage	Citation
ERA-5	Reanalysis	T639 ($\sim 0.28^\circ \times 0.28^\circ$)	Hourly	1979 - 2018	C3S (2017)
MERRA2	Reanalysis	$\sim 0.5^\circ \times 0.625^\circ$	3-hourly	1980 - 2018	Gelaro et al. (2017)
ERA-Interim	Reanalysis	T255 ($\sim 0.70^\circ \times 0.70^\circ$)	6-hourly	1979 - 2018	Dee et al. (2011)
CHIRPS	Precipitation	$0.05^\circ \times 0.05^\circ$	Daily	1981 - 2018	Funk et al. (2015)
TAMSAT	Precipitation	$4\text{km} \times 4\text{km}$	Daily	1983 - 2018	Maidment et al. (2017)

tropical depressions. This was deemed appropriate since the southern African tropical lows are similar in spatial extent and magnitude to tropical lows which transition into and out from tropical cyclones. The only change made was to increase the lowest pressure level from 850 hPa to 800 hPa, to take into account the high topography of southern Africa. This methodology created a superset of lows in southern Africa and the surrounding oceans, which contained all the land-based tropical lows as well as extra-tropical cyclones, ocean based tropical lows, coastally trapped Kelvin waves, equatorial Rossby waves and potentially spurious events.

More precisely, the tracking algorithm was applied as follows. First, an un-weighted vertical average of six-hourly instantaneous relative vorticity at atmospheric pressure levels 600, 700 and 800 hPa were averaged vertically. The data were then smoothed to a horizontal resolution containing only the spherical harmonic wavelengths between 6 and 63, to remove the spurious effects of noise. A Sardeshmukh and Hoskins (1984) filter was applied to smooth the spectral cut off. This step ensures the same spatial scales are identified and smooths the data to make the tracking more reliable. All grid point vorticity minima within the domain of $0^\circ - 100^\circ \text{ E}$ and $0^\circ - 30^\circ \text{ S}$ with $\zeta < -5 \times 10^{-6} \text{ s}^{-1}$ were identified and then refined using B-spline interpolation and steepest gradient descent minimisation

to identify off-grid minima. In the initial phase of tracking, these minima were linked together using a nearest neighbour method. The accuracy and realism of the tracks are then improved using an optimisation scheme which minimises a cost function of displacement distance and track smoothness based on adaptive constraints following the methodology of Hodges (1999).

To isolate the continental tropical lows, tracks that satisfied the following criteria were selected:

- (1) Tracks must spend at least one time step over the southern African mainland;
- (2) Track longevity must be at least one day;
- (3) The filtered vertical mean relative vorticity must satisfy $\zeta < -3 \times 10^{-5} \text{ s}^{-1}$ in at least one 6-hourly time-step;
- (4) There must be coincident cyclonic vorticity at 600, 700 and 800 hPa for a continuous 24 hour period;
- (5) The genesis location must not be in the Atlantic; and
- (6) The genesis location must be north of 25°S .

The 2nd and 3rd criteria are necessary to eliminate spurious events, and the sensitivity of the results of this chapter to the threshold in the 3rd criteria is considered in the sensitivity analysis (Appendix A). The 4th criterion ensures that the tracked lows are deep, eliminating heat lows, which are capped at mid-levels by an anticyclone. Though coastally trapped Kelvin Waves are shallow features, they do sometimes have weak cyclonic vorticity aloft, thus criterion 5 is introduced to remove those that pass the other criteria. We found that all the systems that formed in the Atlantic and then moved onto the continent were coastally trapped

Kelvin waves, and not tropical lows. The 6th criterion eliminates extra-tropical cyclones.

In Section 4.6, we consider the synoptic relationship between tropical lows and continental TTTs. Continental TTTs are identified as cloud bands with low daily outgoing longwave radiation (OLR) using the MetBot algorithm developed by Hart et al. (2012), with centroids between 0 and 40°E. A TTT was considered to be associated with a tropical low if the tropical low was located within 5° of the TTT axis on the day that the TTT formed.

4.3.2. Validation

No upper bound was placed on the strength of tropical lows, and so any tropical cyclone that made landfall onto the southern African mainland in the requisite time period would be counted as a tropical low. This provides the opportunity to validate our methodology against IBTrACS, the most complete global dataset of tropical cyclones (Knapp et al., 2010), by confirming that the IBTrACS tropical cyclones that make landfall in southern Africa are detected by our algorithm. In the IBTrACS dataset, 25 tropical cyclones met our criteria for landfall between November 1980 and March 2018. Notable cyclones in this dataset included Tropical Cyclone Eline and Tropical Cyclone Bonita (Reason and Keibel, 2004; Mudenda and Mumba, 1996).

The tracked tropical low datasets for each reanalysis was compared to the IBTrACS dataset. Two cyclones from different datasets were deemed to be the same event if their trajectories overlapped in time and the root mean square error between their locations at the overlapping times was less than 2.5 degrees. In ERA-5, 24 cyclones were fully present and one (Severe Tropical Storm Lisette, Feb 1997) was missing. The root mean square error averaged across all present

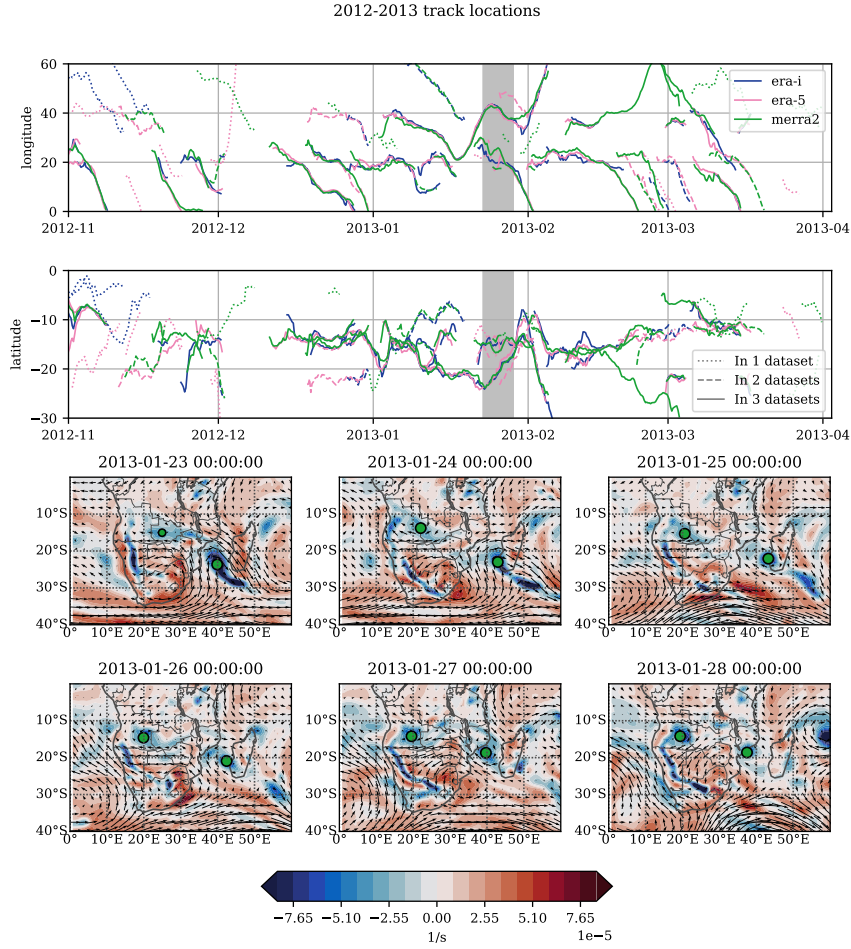


Figure 4.1. Top two rows: Hovmöller plots of track locations for a sample summer season (2012 - 2013). Upper panel: longitude against time, middle panel: latitude against time. Blue: ERA-Interim, Pink: ERA-5 and green: MERRA-2. Solid lines denote tracks that have been identified in all three reanalyses, dashed lines show tracks identified in only two reanalyses and dotted lines show tracks present only in one reanalysis dataset. Black line: bottom row case study date. See text for methodology used to determine whether a track is present in more than one dataset. Bottom two rows: ERA-Interim case study tropical low locations (dots) against 800 hPa vorticity and winds from the 23rd to the 28th of January, 2013. The grey band in the upper panels indicate this time period. Smaller dots indicate weaker tropical lows.

cyclones was 0.9 degrees. In ERA-Interim, 21 cyclones were fully present, 3 had a partial time overlap and one (Severe Tropical Storm Astride, December 1999) was missing. Finally in MERRA-2, 19 were fully present, 4 were partially present and two (unnamed tropical depression of January 1998 and Moderate Tropical Storm Berobia of January 1986) were missing. The root mean square error in overlapping times was 1.2 degrees in both ERA-Interim and MERRA2.

To further validate the tracking algorithm, we compare the tracking results between the three reanalysis products. When tracks whose RMSE error over the period of their overlap is less than 2.5 degrees are deemed to be equivalent, 71% of ERA-Interim tracks, 65% of ERA-5 tracks and 44% of MERRA-2 tracks are equivalent to a track in at least one other dataset. We find that the track locations of low intensity events tend to diverge more than high intensity track events. When the RMSE threshold is increased to 4 degrees, these proportions increase to 83%, 81% and 62%. Note that with this methodology, a track in one reanalysis product may be equivalent to multiple tracks in another reanalysis product. Evidently, MERRA-2 shows less agreement than the other two reanalysis products. The top two rows of Figure 4.1 show Hovmöller plots of track locations in the 2012 - 2013 season, as a sample. Track matching is shown by line styles, using the methodology described above with the 4 degree RMSE threshold. From this, it can be seen that tracks tend to match each other quite well, but sometimes diverge in different directions. 6 of the isolated tracks that are not present in other reanalysis datasets are located over the oceans, only entering southern Africa for short intervals. It is possible that corresponding tracks were present in the other reanalyses but did not move over the southern African land mass.

The lower rows of Figure 4.1 shows the track locations from a sample week (23/1/2013 - 28/1/2013) in ERA-interim, together with the 800 hPa winds and

the non-filtered relative vorticity. Two tropical lows were detected over southern Africa during this time period, one in Angola and the second in the Mozambique channel. Larger circles indicate the track location when the T63 filtered vorticity anomaly is larger than $3 \times 10^{-5} \text{ s}^{-1}$, defined later as a threshold for strong tropical lows. The cyclonic circulations around each strong tropical low centroid is clear, as is the associated vorticity. The wind fields and vorticity anomalies evolve slowly across the time period, and it is evident that the two tropical lows retain their coherence through time. The continental tropical low moves steadily westward from Zambia into Angola and intensifies, while the Mozambique channel event tracks north-westward, making landfall in Mozambique on the final day. A large, elongated cyclonic vorticity anomaly in Namibia is also evident in the vorticity map. This is the trans-Kalahari discontinuity (Chapter 5) and is a shallow feature, disappearing at 600 hPa. Based on the above analysis, we conclude that the track algorithm works acceptably well over southern Africa, and we proceed to examine its results in the following sections.

4.4. Characteristics of tropical lows

Applying the TRACK algorithm to the three reanalysis products lead to the identification of on average 44, 33 and 27 tropical lows per summer from November to March in MERRA-2, ERA-5 and ERA-Interim respectively. These counts differ from that of Hurley and Boos (2015), due to differences in post-TRACK processing: we provide a minimum vorticity threshold rather than mean sea level pressure, and specify a minimum duration of one day rather than two. The months in which tropical lows were most common were December, January and February. The following analysis describes key characteristics of the track

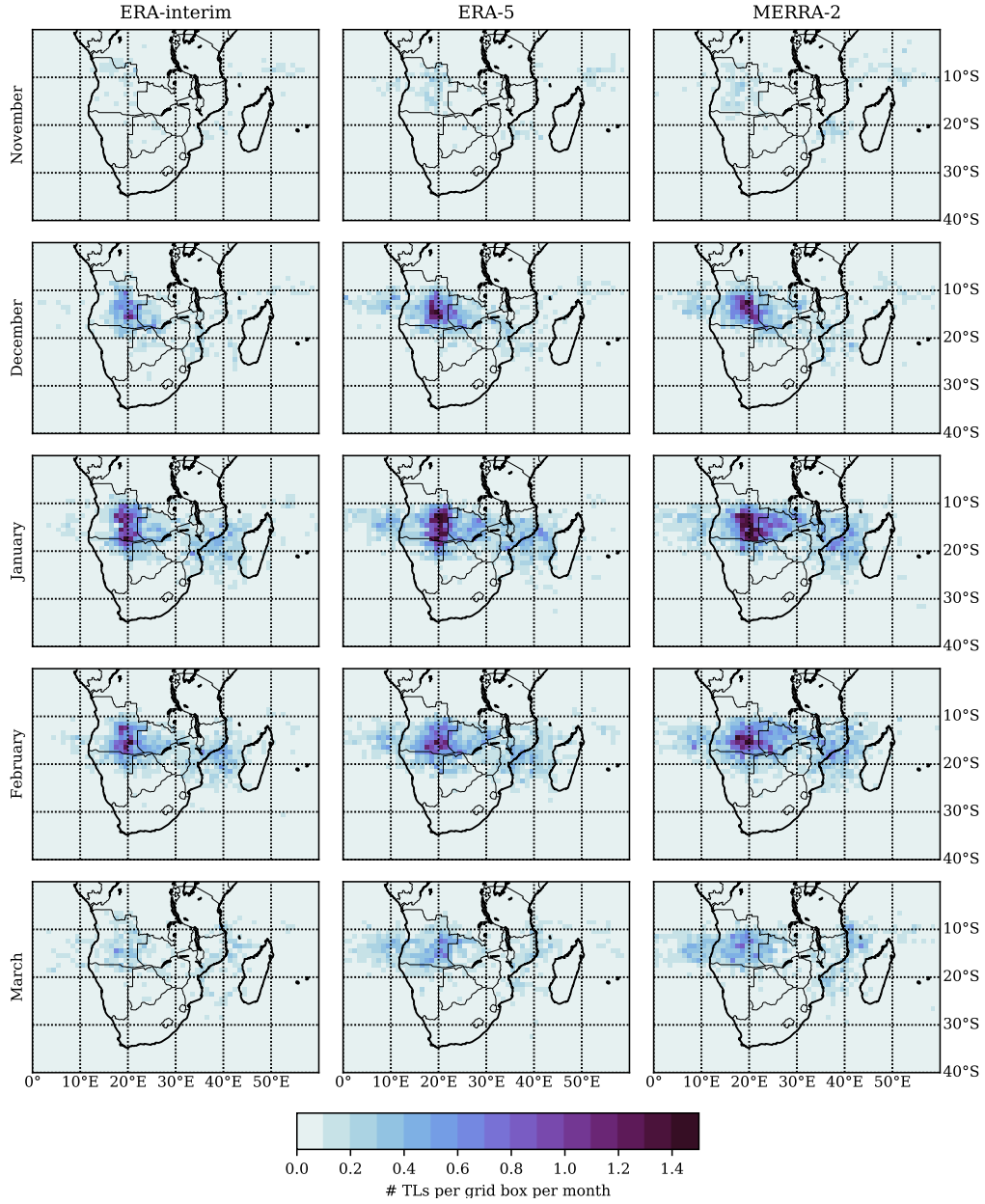


Figure 4.2. Monthly track density in each reanalysis product of tropical lows (number of incidents per $1^\circ \times 1^\circ$ grid box per year). Left column: ERA-Interim, middle column: ERA-5 and right column: MERRA-2. Rows show different months from November through to March, as labelled. The strong vorticity threshold ($-3 \times 10^{-5} \text{ s}^{-1}$) has been applied.

locations and properties, and contrasts the differences between the reanalysis products.

Tropical lows were identified across southern Africa and were most common between the latitudes of 10°S and 20°S , as shown in Figure 4.2, which shows monthly heatmaps of tropical low locations in the three reanalysis products. In all reanalyses from December to February, the maximum number of tropical lows occurred along the 20°E longitude line, the locus of the Angola Low. This maximum was less pronounced in MERRA-2 than the other reanalyses, and the overall number of tropical lows in December and March was lower in ERA-Interim compared to both MERRA-2 and ERA-5. Meanwhile, there is a high degree of agreement evident in Figure 4.2 in the seasonal cycles, spatial patterns and maximum frequencies between MERRA-2 and ERA-5. The mean tropical low latitude was 15.6°S in ERA-5, 15.7°S in ERA-interim and 15.1°S in MERRA-2.

Figure 4.3 shows the genesis and lysis locations of tropical lows across the three reanalysis products, together with the individual genesis locations and track paths from one selected summer season (2006 - 2007). This season was chosen as a fairly typical tropical low year with neutral ENSO, for illustrative purposes only. From this figure it is evident that the vast majority of continental tropical lows form over land, as opposed to the Mozambique channel or the Indian Ocean. The latitude band around 13°S between 18 and 30°E is a particularly strong genesis region in ERA-5 and MERRA-2, although not in ERA-Interim. There is a large lysis zone in the west of the domain at 0°E , however, this simply indicates tropical lows moving outside the tracking domain. In the top row of Figure 4.3, the increased latitudinal spread in ERA-Interim compared to the other reanalyses is again evident. Across the reanalysis products, 33.5% of tropical lows form in the Angola Low region ($11 - 19^{\circ}\text{S}$, $14 - 25^{\circ}\text{E}$, consistent with Chapter 3) and 23.5%

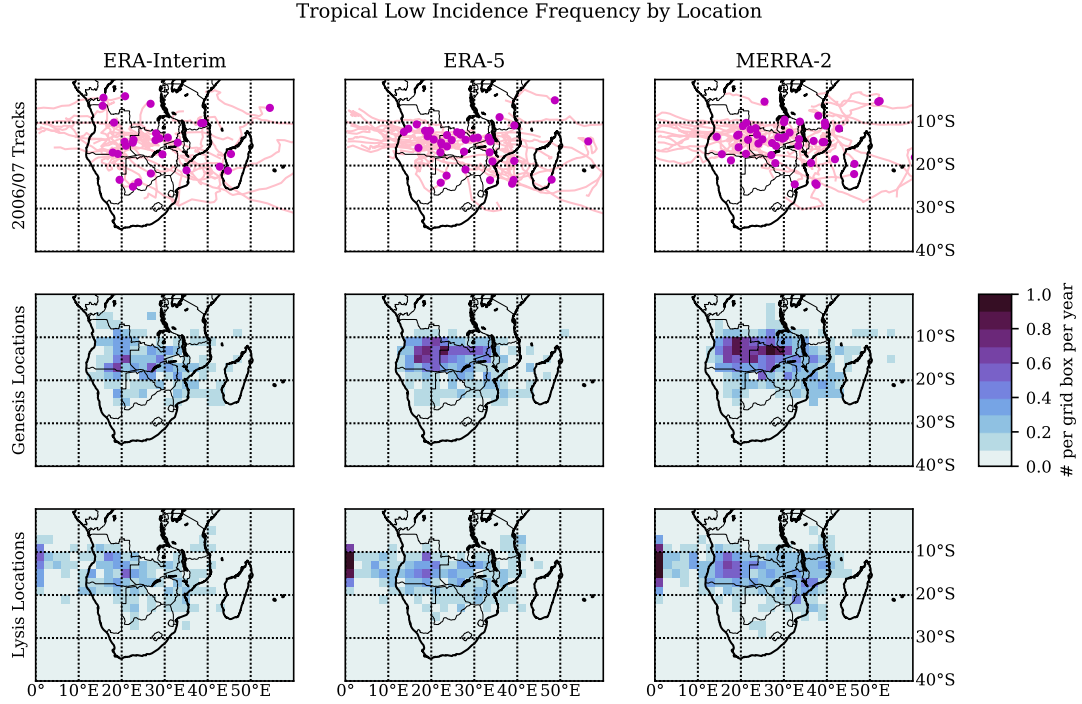


Figure 4.3. Track feature locations for three reanalysis products. Top Row: track paths (pink lines) and genesis points (purple dots) for one selected summer season (2006 - 2007). Middle Row: Genesis Locations (number of incidents per $2^\circ \times 2^\circ$ grid box per year). Bottom Row: Lysis Locations (number of incidents per $2^\circ \times 2^\circ$ grid box per year). Left column: ERA-Interim, middle column: ERA-5 and right column: MERRA-2.

lysis there. Meanwhile, the track density in the Angola Low region comprises of 37% of the total track density.

Other characteristics of the tropical lows are presented in Figure 4.4. It is apparent from the top right panel that the most common duration of a tropical low is 4.5 days, although some last up to a month. Due to the skewness of the distributions, the mean longevity is 7.5 days. In MERRA2, this long tail of the distribution is reduced compared to the other reanalyses, and so although MERRA2 has more tropical lows per year than ERA-5, the total number of tropical low days in the two datasets is roughly equal. This is likely be due

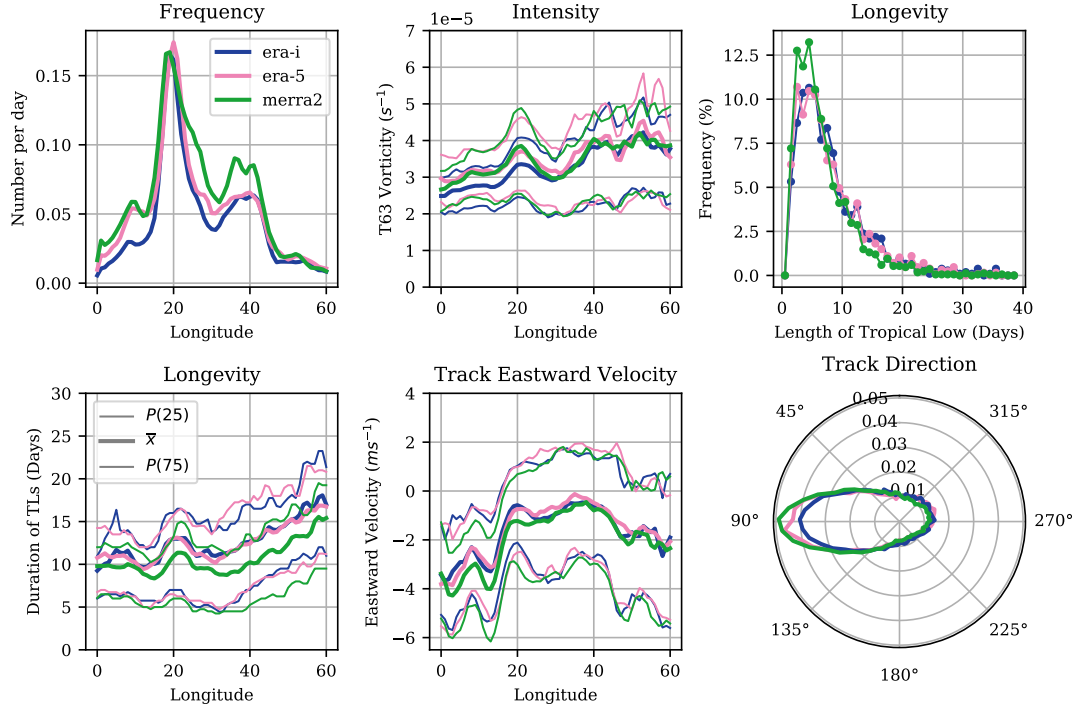


Figure 4.4. Top Left: Probability that a tropical low will be present at a given longitude on a given day over the main tropical low season. Top Centre: Average filtered vorticity at centre of each tracked low at each time-step by longitude. Top Right: Overall distribution of tropical low lifetimes. Bottom Left: average track duration at each time-step as a function of longitude. Bottom Centre: Average zonal velocity of track motion at each time-step by longitude. Bottom Right: Polar distribution plot of track direction in degrees. Thinner lines, when present, indicate the inter-quartile range. Line colours indicate reanalysis product, with dark blue: ERA-Interim, pink: ERA-5 and green: MERRA-2.

to some tropical lows in MERRA-2 decaying and then reforming, and so being counted twice as compared to in ERA-5 where they may have persisted as the one system. The predominant track direction hypothesised from the tracks in Figure 4.3 is confirmed in the bottom right panel, showing that the most common track direction is west. The left and centre panels show frequency, intensity, longevity and eastward velocity as functions of longitude.

The spike in frequency at the Angola Low locus is again evident in the frequency plot in Figure 4.4. This comes with an increase in intensity and longevity at the same location. The increase in longevity explains the differences in the peaks at 20°E between Figures 4.2 and 4.3: tropical lows to the east of 20°E do not last as long as those at 20°E. Note that the longevity in the lower left panel is calculated at each longitude as the average longevity of each track that exists at that longitude at some time. If a track stays at one longitude for more than one time-step, then it is multiply-counted, and if it moves between longitudes then it is included in the average at all these locations. The sudden drop of in frequency to the west of the 20°E meridian is explained by the bottom centre panel: an increase in the mean track velocities of the tropical lows means from 2 to 4 $m s^{-1}$ ensures that tropical lows move swiftly away from the zone between 8 and 18°E, and out into the Atlantic ocean. Over the Indian Ocean between 40 and 60°E, mean tropical low intensity is steady at around $4 \times 10^{-5} s^{-1}$, which then decreases as tracks move inland, before recovering in Angola. The longevity of tracks in the Indian Ocean declines as tracks reach the coast, which may be a selection effect: tracks that exist at 60°E must exist for a longer duration in order to travel westward and meet the criteria of making landfall onto southern Africa.

4.5. Interannual variability

In the previous section, we established that tropical lows occur regularly in southern Africa in the austral summer, and thus are a significant feature of the local climatology. This raises the question of how they vary from year to year, and the potential causes of this variability. This question is particularly pertinent due to the high degree of rainfall variability in southern Africa and the potential contribution of tropical lows to southern African summer rainfall, which will be

addressed in Section 4.7. In this section, we quantify and compare the interannual variability of the incidence frequency of tropical lows in the three reanalysis products, characterise the relationship between tropical lows and ENSO, and identify a link between the tropical easterly jet (TEJ) and seasons with high tropical low activity.

In order to obtain a sensible measure of the interannual variability of tropical low occurrence, we considered the number of days in each year featuring tropical lows centred over land with $\zeta < -3 \times 10^{-5} \text{ s}^{-1}$. This metric creates a good sample of the number of days with strong lows in a season. The top panel of Figure 4.5 shows a time-series of this metric over the years in the study period for each reanalysis product. The lesser number of tropical lows in ERA-Interim compared to the other datasets is apparent from this figure. The correlation coefficients between pairs of reanalyses during their overlaps were 0.72, 0.73 and 0.77 for pairs [MERRA2, ERA-5], [ERA-5, ERA-Interim], and [MERRA2, ERA-Interim] respectively. The statistics of these relationships will be further explored later in this section using composites.

The El Niño Southern Oscillation (ENSO) is an important driver of interannual variability in tropical lows, with more tropical low days in the La Niña phase and fewer in the El Niño phase. The correlation R values were -0.48 , -0.38 and -0.38 between the interannual tropical low index described above and the JFM Niño 3.4 SST index, for ERA-5, ERA-Interim and MERRA2 respectively. All values are significant at the $p = 0.05$ level. The lower panel of Figure 4.5 shows the average number of tropical low days across reanalyses per year compared to the predicted number of tropical low days based on regression onto the Niño 3.4 SST index (DJF). It is evident that there is a strong under-prediction in 1987-88, 1992-93, 1996-97, 1997-98 and 2006-07. This is consistent with the results

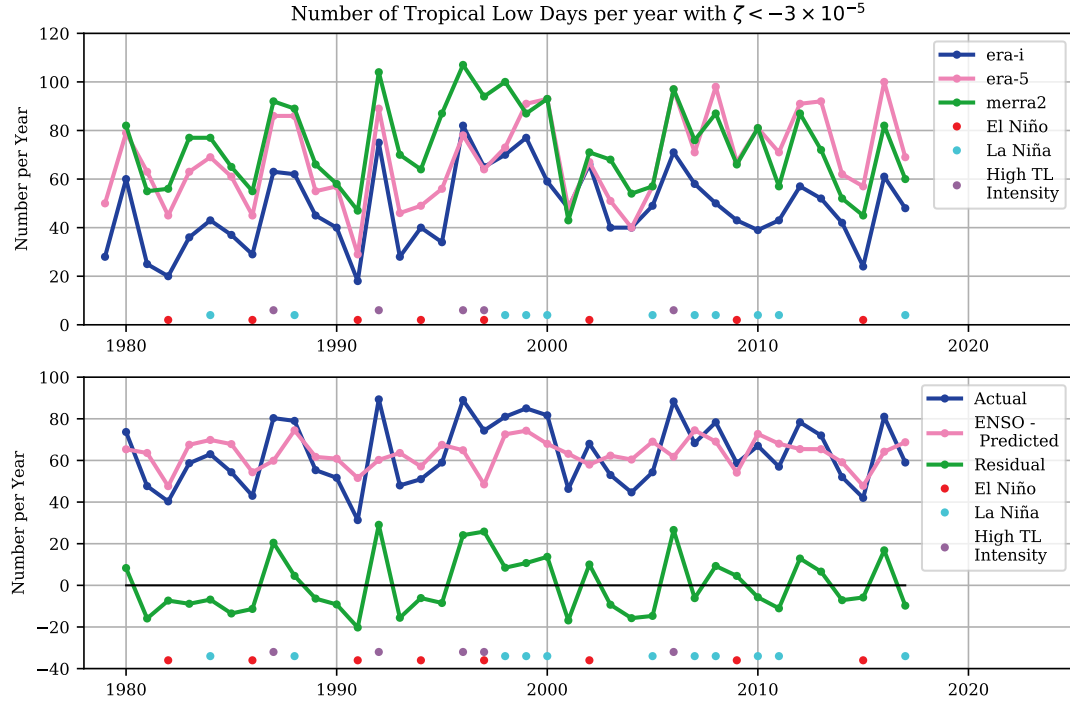


Figure 4.5. Upper panel: interannual variability of number of strong ($\zeta < -3 \times 10^{-5}$) continental tropical low days per year in each reanalysis product. Colours indicate reanalysis product. Lower panel: average number of strong continental tropical low days per year across reanalyses (dark blue), prediction of the blue line based on a regression onto the Niño 3.4 SST index (pink) and the residual of this regression (green). The dots at the bottom of the panels indicate the years of the composites used in Figures 4.7 and 4.12: red for El Niño, light blue for La Niña and purple for the high TL intensity composite.

of Reason and Jagadheesha (2005), who found that the strong Angola Low was present during the 1997-98 El Niño, and linked this to the lack of materialisation of an El Niño linked drought in this year.

For each month between November and March, the interannual monthly average number of tropical lows across the three reanalyses was calculated, and regressed on the Niño 3.4 SST index 3 month average centred on that month. The result of this regression was significant at a $p < 0.05$ level for December,

January, February and March, though not for November. Based on these regressions, two composites of months were created: the first where the residual of the above regressions was above 8.2 tropical low days per month, and the second where the residual was smaller than -8.2 tropical low days per month. The value 8.2 was chosen as $1.5 \times \sigma$, where σ is the standard deviation of the residuals across all 5 regressions. The resultant months in the first composite were: March 1988, February and March 1993, February 1995, February 1997, January 1998, March 2000, March 2003, January 2007, December 2008, January 2013 and January 2015. The second composite contained January 1985, December 1988, February 1992, December 1993, December 1994, January 2000, February 2005, February 2008, February 2011, February 2012 and February 2015. These composites are designed to extract the interannual mode of variability of tropical lows that is independent from ENSO. The full wind vectors at 800 hPa are shown in Figure 4.6. The Angola Low is much stronger in the composite with more tropical lows than that with less tropical lows, by construction.

Figure 4.6 also shows the monthly anomalous zonal winds at 200 hPa for the two composites, with full winds contoured in blue. From this it is apparent that the tropical easterly jet (TEJ) is anomalously strong over central and southern Africa in the first composite. This result is significant at the $p < 0.01$ level. It is slightly weaker than average in the second composite, although this result is not significant at the $p < 0.01$ level. Although these composites were designed to study a mode of variability that was distinct from ENSO, there remained an ENSO signal in the composites, with the strong composite having a slight overall El Niño signal and vice versa. This is in part due to the inclusion of the 1997 - 1998 El Niño in the strong composite. This led to significant wind anomalies over the Western Pacific in Figure 4.6. However, when the Niño 3.4 signal was

linearly removed from the 200 hPa winds, the TEJ signal remained significant and the Western Pacific signal did not (not shown).

Therefore, the strength of the TEJ over Africa appears to be linked with the interannual variability of tropical low days, in a manner that is independent from ENSO. However, this does not imply causation and the direction of any causation that may exist is unknown. It is possible that increased upper level wind speeds may make the formation of tropical lows more likely. However, an increase in tropical low activity would provide a heat source in the upper atmosphere, which may accelerate the TEJ.

To examine the spatial distribution of the interannual variability, three seasonal composites were considered. The first consisted of 5 El Niño years with Niño 3.4 SST anomalies greater than $0.8K$, and are indicated by red dots in Figure 4.5. The second consisted of 5 La Niña years with Niño 3.4 SST anomalies below $-0.8K$, indicated in Figure 4.5 by light blue dots. The final composite, labelled as the High TL Intensity composite, contained the years described above where the annual tropical low strength was significantly under predicted by the ENSO regression, and is shown in purple dots. The strong El Niño of 1997-98 is included in both the first and the third composite because of the strong positive residual in that summer.

Figure 4.7 shows the difference between the climatological heat-maps and the composite heat maps for each case. In the El Niño composite, there is a slight northward shift in the preferred track location, as well as an overall decrease in the number of tropical lows. In the La Niña composite, there is an overall increase and a southward shift. This is consistent with the general behaviour of El Niño reducing precipitation in subtropical southern Africa, and vice versa for La Niña (Lindesay, 1988). The northward and southward shifts of the composite mean

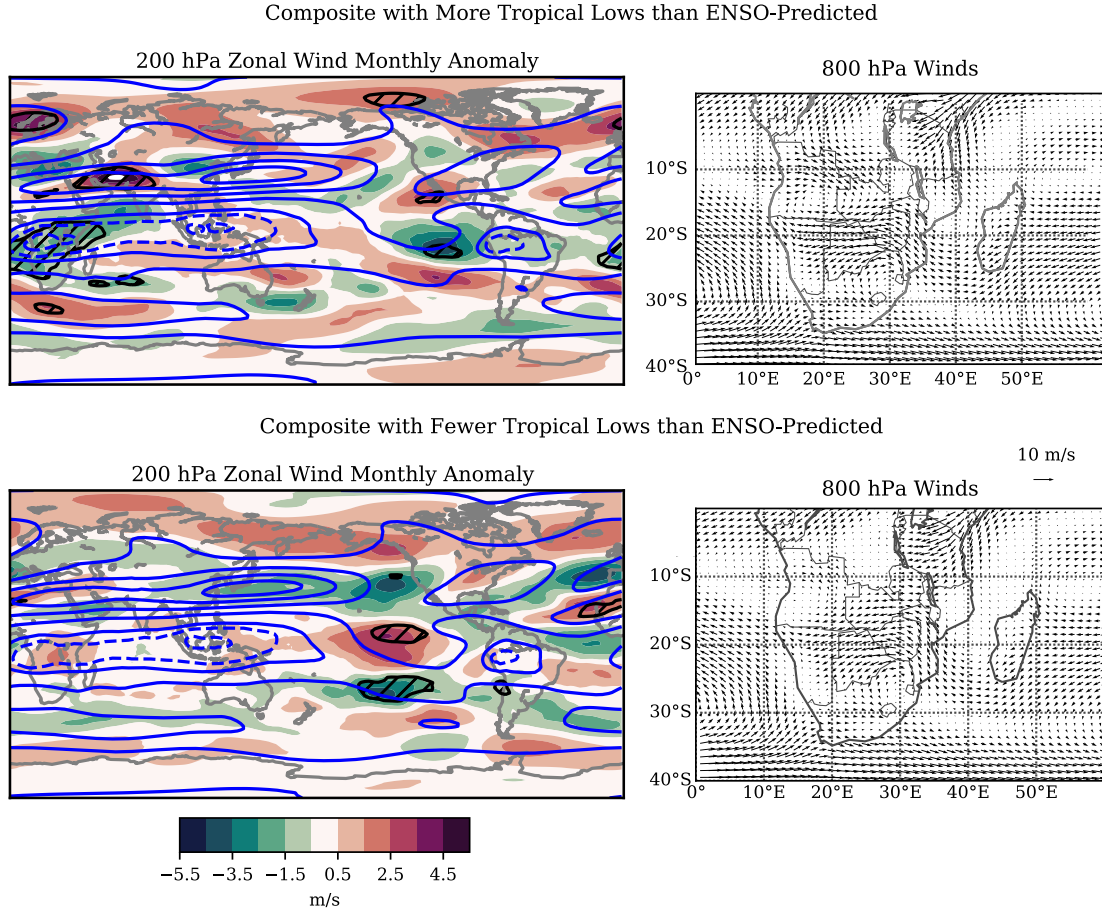


Figure 4.6. Stronger-than-ENSO (top row) and weaker-than-ENSO composites (bottom row). Left: 200 hPa MERRA-2 zonal winds with Colours showing the wind anomaly (m/s), blue contours: full composite wind speed (negative contours dashed with 5 m/s contour intervals, positive contours solid with 15 m/s contour interval). Shaded regions: significant at a $p < 0.01$ level. Right: full 800 hPa wind vectors.

track centroids compared to the climatology were significant at a $p = 0.05$ level, using a one-sided Welsh's t-test. These t-tests were performed using all latitudes of tropical lows over land at time-steps where the filtered vorticity satisfied $\zeta < -3e - 5 \text{ s}^{-1}$, and used the total number of tropical lows (as opposed to the number of tropical low days) as the degrees of freedom. During La Niña, the mean tropical low latitude over land shifted 0.5 degrees further south compared

to the climatological mean in each reanalysis product. During El Niño, the shift was 0.8 degrees north in ERA-5 and MERRA-2 and 1.0 degrees north in ERA-Interim. In the final composite, the increase in tropical low frequency occurs throughout the domain and no clear shift in preferred location is apparent.

A statistically significant overall increase in the number of tropical low days a year is observed at the $p = 0.05$ level during La Niña phase in ERA-5 (12 days/year) and ERA-interim (9 days/year) but not in MERRA-2. A significantly significant overall decrease is observed in the El Niño phase in ERA-5 only (14 days/year). However if 1997-98 is removed from the ENSO composite, the decrease becomes significant for every reanalysis product. The high TL intensity saw a significant increase in all three reanalysis products, with increases of 16, 24 and 27 days a year.

4.6. Links and limitations between Angolan tropical lows and TTTs

It has often been hypothesised that the genesis of Tropical Temperate Troughs (TTTs) is linked to the behaviour of the Angola Low (Todd and Washington, 1999). Since the Angola Low is the climatological result of the synoptic anchoring of tropical lows in the Angola region, it is reasonable to ask whether there is a link on the synoptic timescale between TTTs and Angolan tropical lows (ATLs). In this section, we investigate whether there is a synoptic link between these two classes of events.

If TTT and ATL events were completely unrelated on a daily timescale, then according to the definition of statistical independence, the probability of TTTs that form on an ATL day would be equal to the product of the probability that the day is an ATL day with the probability that a TTT would form on that day. Since ATL and TTTs possess different seasonal cycles, this is equivalent to saying

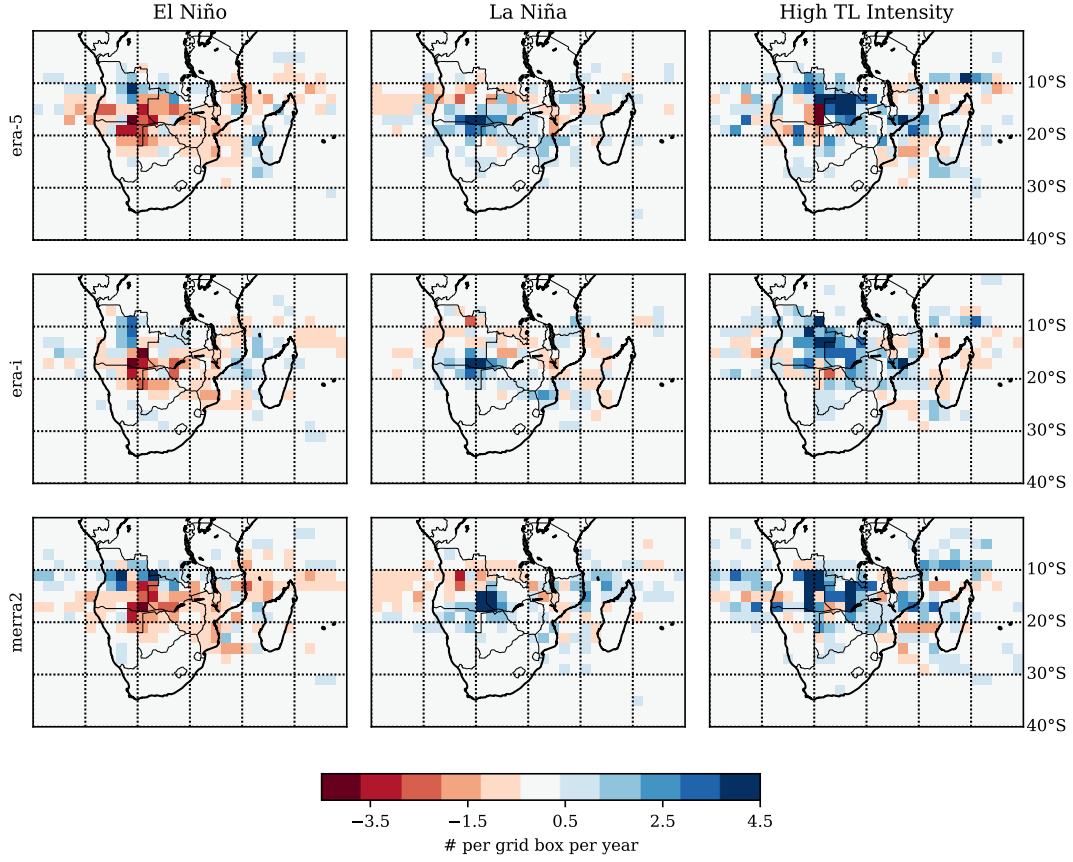


Figure 4.7. Annual anomalous track density for (left) El Niño composite, (centre) La Niña composite and (right): the high TL intensity composite. Rows indicate the reanalysis product used - top row: ERA-Interim, middle row: ERA-5 and bottom row: MERRA2. The strong vorticity threshold ($-3 \times 10^{-5} \text{ s}^{-1}$) has been applied.

that if TTT and ATLs were completely unrelated, then the proportion of days at a given time of year in which of TTTs that form on during an ATL would be equal to the product of the proportion of days that are ATL days with the proportion of days on which a TTT forms, at this time of year. i.e.,

$$P(TTT \cap ATL)_t = P(ATL)_t \times P(TTT)_t \quad \forall t \iff \text{TTT, ATL are independent}$$

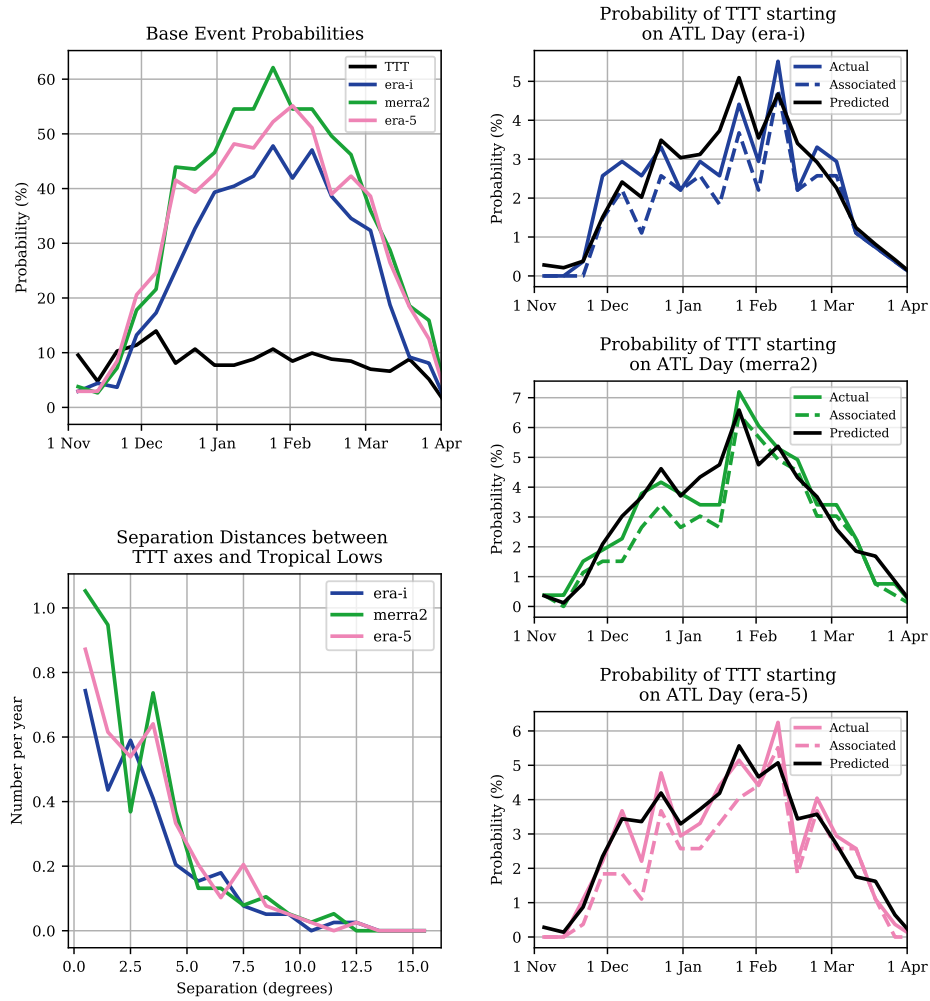


Figure 4.8. Seasonal cycles of probabilities of Angola tropical lows and TTTs, together with separation distances between coincident events. Top left: probability of TTT events beginning (black line) and Angola tropical lows occurring (coloured lines) on a given day. Bottom left: Separation distance between Angola tropical low centroid and TTT axes on days where Angola tropical lows and TTTs both occur. Right: Probability of TTT and Angola tropical lows occurring on the same day. Solid coloured lines: observed coincident events, dashed coloured lines: observed coincident events with separation distance $< 5^\circ$. black: predicted probability of coincident events under the assumption of statistical independence.

Conversely, if ATLs caused TTTS, then $P(TTT \cap ATL)_t$ would be greater than $P(TTT)_t \times P(ATL)_t$ for some time of year t .

To determine the interdependence of ATLs and TTTS, Figure 4.8 compares the predicted probability of TTTS forming on ATLs under the assumption of independence with the observed coincidences of these events. The existence of the ATL has been determined using the three reanalysis products. The top left panel of Figure 4.8 indicates the seasonal cycle of the TTT formation (black), ATL occurrence (coloured). The right panels indicate the seasonal cycle of the proportion of days with formation of TTTS during ATLs (coloured solid lines) and the theoretical prediction of this metric under the assumption that TTTS and ATLs are independent (black line). The dashed line shows the proportion of these events where the axis of the TTT came within 5° of the ATL. In all cases, the probabilities are almost indistinguishable. This implies that the presence of an ATL does not alter the likelihood of TTT formation at any time of the year.

The bottom left panel of Figure 4.8 shows PDFs of the minimum separation distance between the axes of TTTS and the centroids of ATLs when the two events occur on the same day. In all cases the frequency decreases with the distance and 80% of the separation distances are smaller than 5° across reanalysis products. This distance has been chosen as the cut-off distance at which TTTS and ATLs will appear to be synoptically linked (see Appendix A for a sensitivity analysis). Thus, although ATLs do not appear to make TTTS more probable, when TTTS form during ATLs they do appear to interact with the tropical convection associated with the ATL.

The interannual and inter-monthly statistical correlations between the number of TTTS and Angola tropical lows in each year in each reanalysis are given in Table 4.2. The correlation between the number of TTTS and the number of tropical

Table 4.2. Interannual Correlation Statistics between the number of TTTs and Angolan Tropical Lows in each time period. Top section: time periods are aggregated DJFM seasons. Bottom section: time periods are months (D,J,F,M only).

		ERA-5	ERA-Interim	MERRA
Seasonal	R	0.40	0.48	0.38
	N	32	32	32
	S.E.	0.026	0.030	0.030
	P Value	0.025	0.0055	0.030
Monthly	R	-0.12	-0.03	-0.07
	N	128	128	128
	S.E.	0.011	0.012	0.011
	P Value	0.195	0.750	0.410

lows in each summer season (DJFM) from 1979 - 2012 are of order 0.4 and are significant at the $p=0.05$ level. The period of December to March was chosen because it is the intersection of peak TTT season with peak ATL season. To consider the relationship at a monthly time scale, time series of number of ATLs and TTTs respectively in each January, February, March and December between 1979 and 2012 were compared. In this case, the correlation decreased and became insignificant. In order to account for the differing seasonal cycles of TTTs and ATLs, the climatological mean number of events in each month was removed from the time series before correlating. This suggests that the relationship between the Angola Low and TTTs exists on an interannual timescale, but not at higher temporal resolution.

The following three scenarios are all logically consistent with this result:

- (1) There is an independent phenomenon, which operates on an interannual or annual time scale, which modulates the frequency of both TTTs and ATLs. The above interannual variability analysis suggest that ENSO and the TEJ are candidates for this scenario;

- (2) In some years, the unusually persistent presence of ATLs modifies the large-scale atmosphere, for example by enhancing the Hadley Circulation and shifting the southern subtropical jet northwards, causing TTTs to become more frequent; and
- (3) In some years, the southern subtropical jet exists further north than usual, and this leads to more frequent ATL occurrences. However, this result provides more evidence against the existence of a direct causal synoptic link between the ATLs and TTTs.

Although we have established that there is no synoptic causal link between ATLs and TTTs, a quarter of summer TTTs events occur in association with ATLs, with a minimum separation distance smaller than 5° between the TTT axis and the ATL centre. Therefore, we next examine the influence that an ATL along the axis of a TTT has on the precipitation of that TTT. In the following, a TTT is considered to be associated with an ATL if the TTT axis comes within 5° of the ATL centre.

The top row of Figure 4.9 shows composited anomalous precipitation (colours) during TTT events, TTTs associated with ATLs and ATL events. Data have been drawn from MERRA-2 and CHIRPS only, as similar results were obtained when ERA-Interim, ERA-5 and CHIRPS were considered. From this it can be seen that the TTTs associated with ATLs feature more seasonally anomalous rainfall than average TTTs. The bottom row of Figure 4.9 compares this to the linear superposition of anomalies associated with the composites of the TTT only and ATL only days. The linear superposition is not vastly different to the rainfall on combined days, as demonstrated by panel E, which shows the difference between the combined event rainfall and the linear superposition. A Welch's t-test, applied

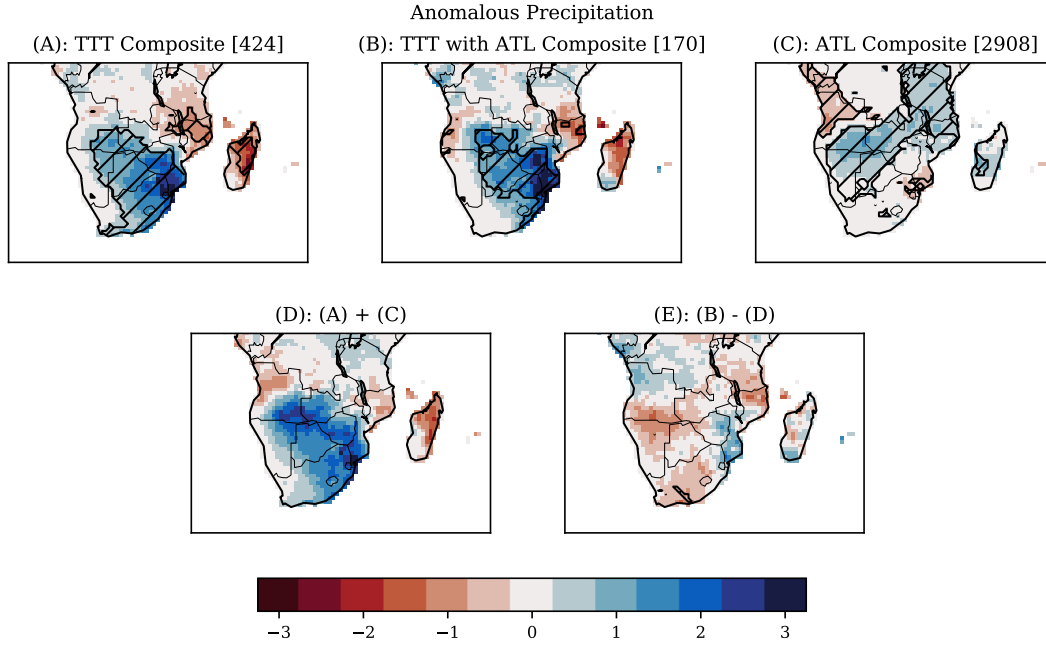


Figure 4.9. Upper row: Precipitation anomalies (mm/day) associated with: (A) TTTs, (B) TTTs coincident with Angola tropical lows with separation distances 5° , and (C) Angola tropical lows. Shading indicates where the result is significantly different from 0 at the $p < 0.05$ level. Numbers in the top right indicate the sample size. Precipitation anomalies are calculated with respect to the climatological daily mean of CHIRPS, smoothed with a 2 week running mean, and composited on days of TTT and/or Angola tropical lows as described in the text. Lower row: linear combinations of the upper panels. (D): the linear superposition of the TTT anomaly and the Angola tropical low anomaly. (E): the difference between the anomaly due to combined events and the linear superposition from (D).

to test whether the data displayed in panel E was significantly different from zero, was unable to reject the null hypothesis at any grid cell after the false discovery rate was controlled for using $\alpha_{\text{FDR}} = 0.05$ (Wilks, 2016). This suggests that the increased precipitation from TTTs associated with tropical lows, as compared to TTT only events, may simply be the rain that would have fallen from the tropical low anyway. Therefore, this analysis has been unable to distinguish combined event rainfall from the superposition of ATL rainfall with TTT rainfall. This

casts doubt further on the existence of a synoptic link between TTTs and ATLs, although the tests applied were not formulated to prove the null hypothesis.

4.7. Precipitation associated with tropical lows

In many regions across the low latitudes, tropical lows are an important feature of local climatologies because of their contribution to annual precipitation. In southern Africa, understanding precipitative processes is particularly pertinent due to the high degree of interannual rainfall variability. In this section, we determine the proportion of rainfall that may be attributed to tropical lows in southern Africa, and the degree to which tropical low variability controls the overall rainfall variability. In order to determine the precipitation contribution of tropical lows, we need a means to attribute rainfall to tropical low events.

Figure 4.10 illustrates the methodology used to attribute rainfall to tropical lows. In the first column, the CHIRPS and TAMSAT rainfall on an example tropical low day is shown. In CHIRPS, a large patch of rainfall exceeding 40 mm/day is located to the north-east of the tropical low is present, and it is highly plausible that this rainfall is associated with the tropical low. The high-intensity blob is contained within 5 degrees of radius of the tropical low centroid. Other rainfall regions are present further afield, but it is unclear whether or not they are associated with the tropical low. The picture in TAMSAT is similar, although the central rain patch beside the tropical low is less intense. Based on this case-study, we consider composites of rainfall centred on the tropical low centroids.

The second column of Figure 4.10 shows the composite anomalous rainfall footprint of tropical lows across datasets, shifted so that the centre of the low is at the plot origin. The climatology used to create the rainfall anomaly is the

smoothed interannual daily rainfall climatology. Since the rainfall datasets were both daily and the tracks were computed every six hours, the composite was constructed by identifying the closest grid cell to the 12pm UTC location of each tropical low on each day, and shifting that grid cell to the origin. The composite mean was calculated using the tracks from the three reanalysis products separately, and then the three resultant composites were averaged together. It is evident that the rainfall is increased at the centre of and to the east of the low. Since tropical lows move predominantly in a westward direction, the rain to the east of the low is most likely a result of the low having been in that location the day before. The positive anomaly is largely contained within a circle of 5° radius around the centre of the low. Therefore, we attribute all rain which falls within 5° of a tropical low as associated with that tropical low. This radius is consistent with many previous studies of tropical lows and tropical cyclones (e.g. Lavender and Abbs (2013), Khouakhi et al. (2017), Dare et al. (2012)). The spatial distribution of the composite anomalies in the three separate reanalysis products were all the same (not shown), but the intensity varied slightly between the reanalyses: for example, the CHIRPS maximum composite anomaly was 5.2 mm/day in ERA-5, 5.0 mm/day in ERA-interim and 3.8 mm/day in MERRA-2.

Based on the attribution methodology described above, between 30 and 70% of precipitation across the tropical edge (16° - 22° S) from November to February was associated with a tropical low. This is demonstrated in the third column of Figure 4.10, which shows a map of the proportion of rainfall attributed to tropical lows. By averaging over the tropical edge latitude band, we found that 31% of rainfall in this region was attributed to tropical lows. In each case, the maximum occurs in south east Angola, and most of the rainfall occurs between 10 and 25 S. Again, the spatial structure was consistent across reanalysis products. Overall, there is

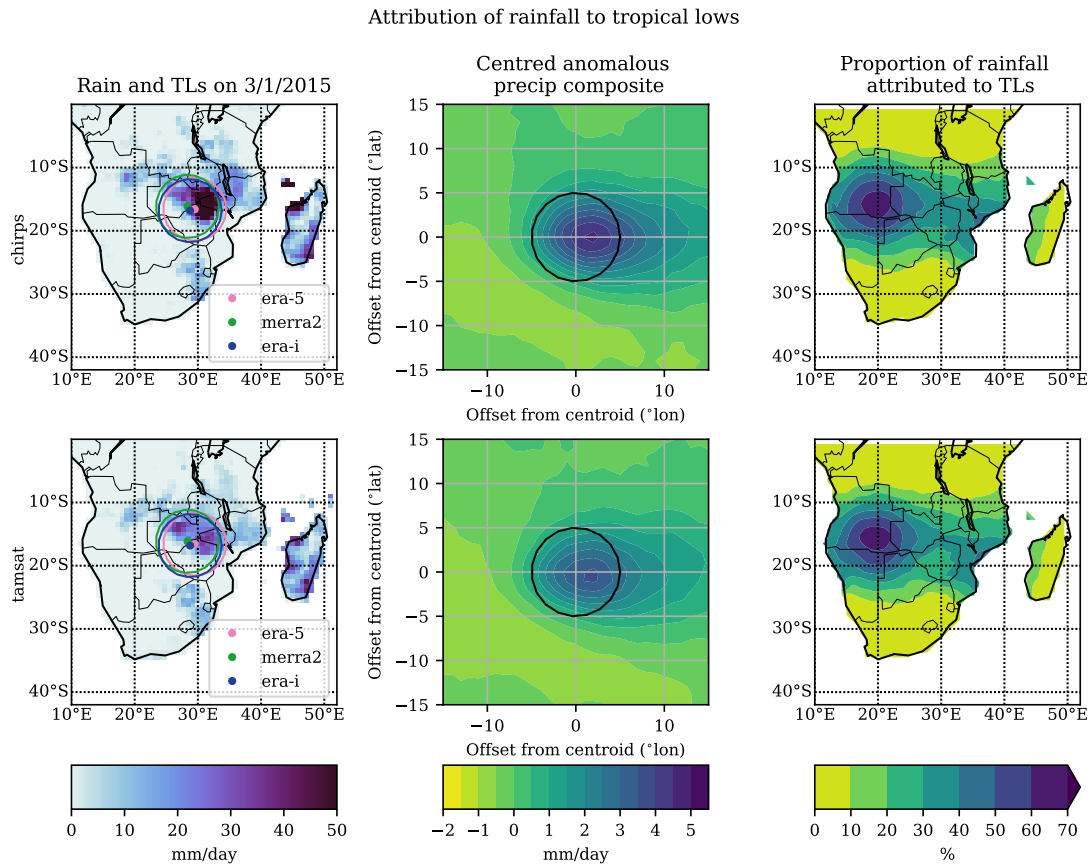


Figure 4.10. Rainfall attribution. Top row: CHIRPS; Bottom row: TAMSAT. The left column shows rainfall on a sample tropical low day (3rd Jan 2015), with the centroid of the tropical low in each reanalysis product shown as a pink, green or blue dot for ERA-5, MERRA-2 and ERA-Interim. A circle of 5 degree radius is shown around each centroid in the same colour. The middle column shows a composite of climatologically anomalous rainfall centred around tropical low centroid (mm/day), averaged across all days and the three reanalysis products. Black circle indicates a radius of 5° around the tropical low centroid. The right column shows the proportion of NDJFM rainfall which falls within 5° from a tropical low centroid. The proportion was calculated independently for each reanalysis product and then averaged.

a lesser fraction of rainfall contributed by tropical lows when ERA-Interim was used to identify the tropical lows, as compared to the other two reanalyses (not shown). The maximum rainfall contributed by tropical lows was approximately 10% lower in ERA-Interim than the other two products. This was likely due to

the smaller number of tropical low days identified in ERA-Interim. There is a good agreement between the two gridded rainfall products.

Interannual rainfall variability is of key concern in southern Africa, and the tropical edge is one of the regions where variability is highest. This raises the question: how much of the variability in this region is variability in tropical low rainfall, and how much derives from other forms of precipitation?

The first column of Figure 4.11 shows the interannual standard deviation of NDJFM precipitation across southern Africa in CHIRPS and TAMSAT respectively. The variability of the tropical edge can be seen very clearly in CHIRPS, as a zonal line of high standard deviation near 18°S. In TAMSAT this is slightly less distinct, but the standard deviation is high between the latitudes between 15 and 25 S. The interannual variability can be decomposed as:

$$\sigma^2 = \sigma_{\text{tl}}^2 + \sigma_{\text{rem}}^2 + 2 \times \text{cov}(P_{\text{tl}}, P_{\text{rem}})$$

In the above equation, σ , σ_{tl} and σ_{rem} are the standard deviations of the seasonal total precipitation, precipitation from tropical lows, and remaining precipitation respectively, and the ‘cov’ term is the covariance of the seasonal precipitation from tropical lows and the remaining precipitation. This decomposition is shown in the remaining columns of Figure 4.11. The green contours on Figure 4.11 show the negative covariance values. Where the covariance is positive or has a relatively low negative value, the interannual variability of tropical low rainfall has a strong control on the total interannual variability.

In the southern tropical edge region, the variability in the rain from tropical lows is high, and the variability of the remaining precipitation is greatly reduced compared to the total variability. Additionally, the covariance term is positive

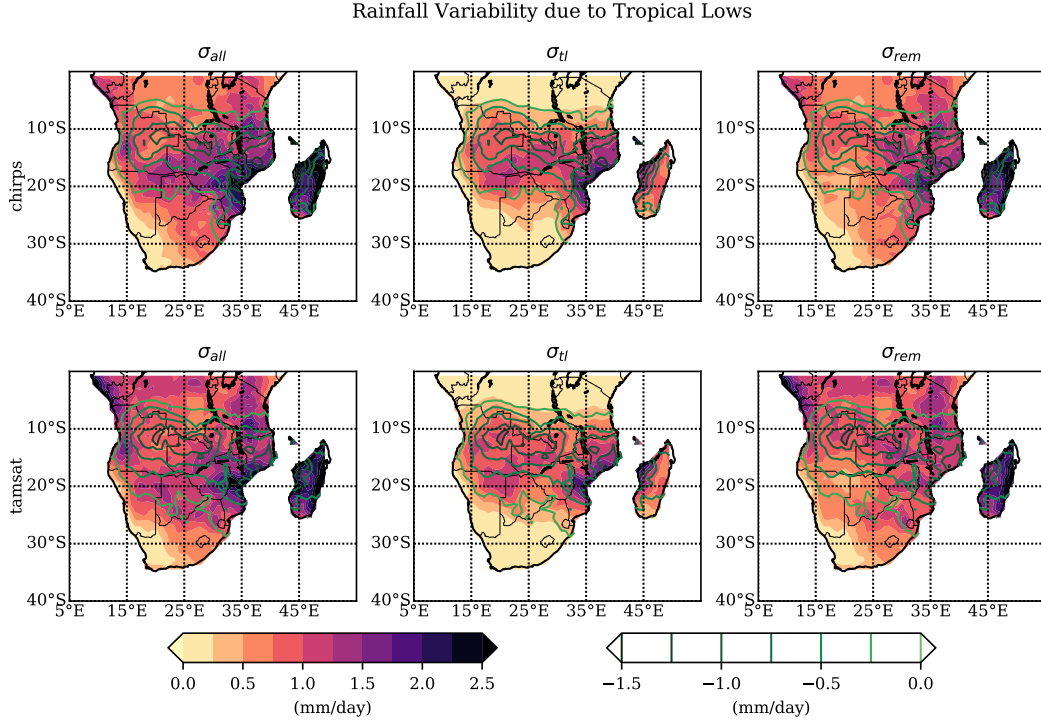


Figure 4.11. Interannual rainfall variability attributed to tropical lows. Top row: CHIRPS, bottom row: TAMSAT. Column 1: interannual standard deviation of total NDJFM precipitation. Column 2: interannual standard deviation of NDJFM precipitation attributed to tropical lows. Column 3: interannual standard deviation of remaining NDJFM precipitation not attributed to tropical lows. The green contours indicate the interannual covariance between tropical low rainfall and other rainfall. For clarity, only non-positive contours of covariance are marked. Each quantity was calculated independently using tracks from ERA-interim, ERA-5 and MERRA-2, and then averaged over the three reanalysis products.

in this region. This implies that a large fraction of the rainfall variability along the tropical edge comes from variability in precipitation that comes from tropical lows, suggesting that an increased understanding of tropical lows may be of value to seasonal forecasting in this region. Notably, this is the region which features more tropical lows during the cool phase of ENSO and less during the warm phase in Figure 4.7.

In northern Angola and Zambia, although the tropical low rainfall variability is high, the variability of the remainder is not reduced. This is because there is a strong negative covariance between interannual tropical low rainfall and the remaining rainfall. In Figure 4.7, this is the region where tropical lows are less numerous during La Niña and slightly more numerous during El Niño. Thus, the reduction in the number of tropical lows that occur in this region during the La Niña phase does not come with a relation in the rainfall in this region because of this negative covariance. This means that in years where there are fewer tropical lows in the northern region, there is more rainfall in the form of meso-scale convective complexes, isolated thunderstorms, etc. that makes up the difference. In the southern tropical edge, however, in years without as many tropical lows, or in years where the tropical lows do not rain as much, precipitation from tropical lows is not replaced and the total rainfall is reduced as a result.

Finally, we use the rainfall attribution methodology to examine the composites developed in Section 4.5. For brevity, we present results from CHIRPS only, though similar conclusions were drawn from TAMSAT. Figure 4.12 shows the anomalous DJF rainfall anomalies for the three composites: El Niño, La Niña and high TL intensity. Rainfall anomalies are shown in total rainfall, rainfall attributed to tropical lows, and the remaining rainfall that fell more than 5° from a tropical low centroid. The majority of the El Niño and La Niña signals over the interior of southern Africa (west of 30°E) are associated with tropical lows. Further east, the ENSO signal over Mozambique and Tanzania is present in the ‘remainder’ panel. Meanwhile, the high TL intensity composite is associated in increased rainfall over Zambia, Tanzania and Mozambique. Only the signal over Zambia is attributable to tropical lows.

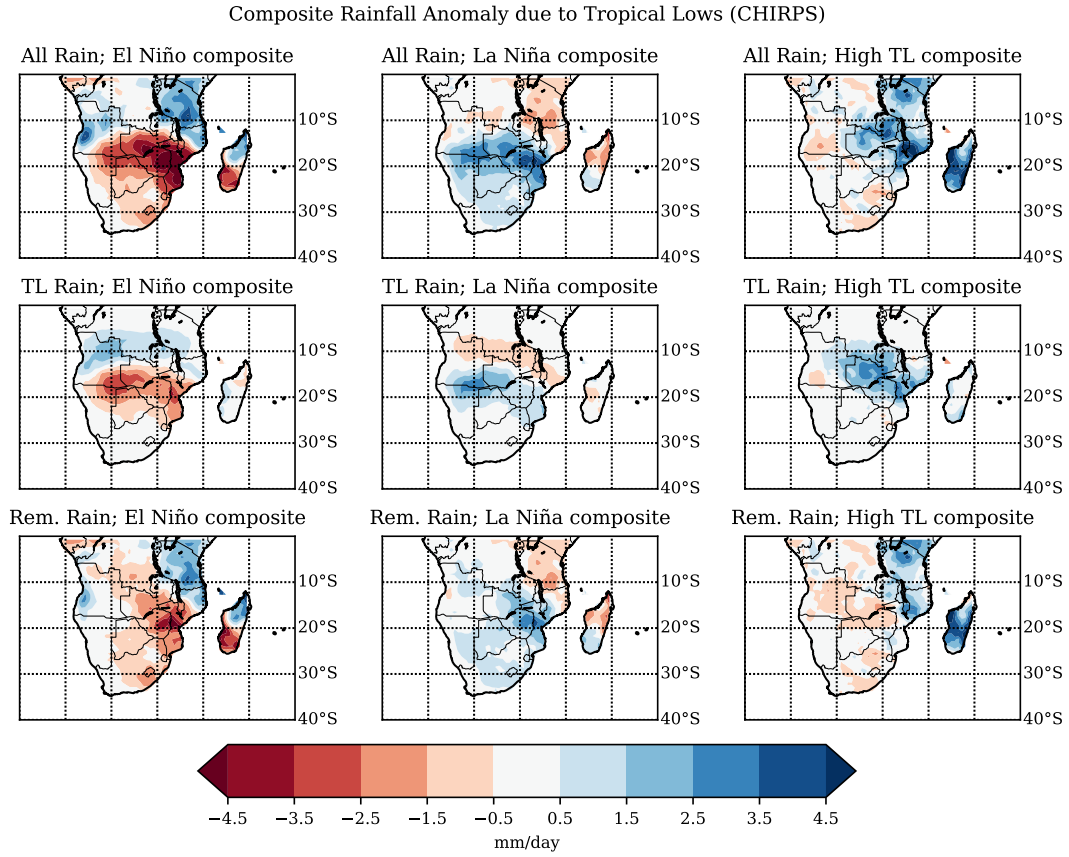


Figure 4.12. Composite rainfall signal explained by Tropical Lows. Left column: El Niño composite, centre column: La Niña composite, right column: High TL Intensity composite. Top row: total DJF rainfall anomaly for composite. Middle row: DJF rainfall anomaly of rainfall attributed to tropical lows during composite. Bottom row: DJF rainfall anomaly of rainfall not attributed to tropical lows during composite. Each quantity was calculated independently using tracks from ERA-interim, ERA-5 and MERRA-2, and then averaged over the three reanalysis products.

4.8. Discussion and conclusions

Comparisons between three reanalysis products and two gridded rainfall datasets have yielded a set of results which are consistent, though with details that differ. These datasets may be viewed as imperfect reflections of the actual atmosphere. By examining the similarities and differences between the datasets,

we are able to assign a qualitative uncertainty estimate to reanalysis features, under the hypothesis that we may be able to ascribe an increased confidence to features that are consistently represented across multiple datasets.

The representation of tropical lows in ERA-5 and MERRA2 are quite similar, with ERA-interim diverging in a number of ways. There are fewer tropical lows that form in ERA-interim than the other two datasets, and their footprint is more spread out. Meanwhile, although there are more separate tropical low events in MERRA-2 than ERA-5, they are typically shorter, possibly because MERRA-2 tropical lows tend to break apart and reform, such that they are double counted with respect to ERA-5. The total number of days with strong tropical lows in MERRA-2 and ERA-5 are remarkably similar (Figure 4.5). Based on this agreement, and with the knowledge that ERA-5 and MERRA2 are both recent and high resolution reanalyses, while ERA-Interim is older and of lower resolution, we may decide to trust the representation of tropical lows in MERRA2 and ERA-5 ahead of ERA-Interim.

Common across all data sources, this research has established that tropical lows are important synoptic events which contribute rainfall along the tropical edge of southern Africa. They tend to form from November to March and peak in January. The distributions of track longevity has a long tail, with some tracks lasting up to 30 days, and with a mean duration of 7.5 days. They move westward and cluster over Angola where they sometimes interact with TTTs, with whom they are not causally linked on a synoptic timescale.

Like the monsoon depressions and tropical lows that have previously been studied in India and northern Australia, southern African tropical lows are an important source of rainfall in the wet season and track predominantly westward. The longevity of southern African tropical lows is longer than in these other

regions, however, at 7.5 days as compared to 4-5 days. Similarly, the semi-stationary nature of the southern African tropical lows appears to be unique, as track density maps in other regions do not show pronounced longitudinal peaks similar to that which occurs over Angola. Chapter 3 found that tropical lows in Angola interacted with the vorticity stretching of the west-coast sea breeze deflected into the mid-troposphere by the steep topography, which caused the tropical lows to be anchored in central Angola. We hypothesise that it is this unique anchoring mechanism that causes the difference between southern African tropical lows and those which form in other regions of the tropics. The mean track speed is 4.2 ms^{-1} , which is comparable to the speed of Australian lows (6 ms^{-1} , (Berry et al., 2012)), where Australian systems were tracked in isentropic 6-hourly ERA-interim PV. Both eastward and westward moving tropical lows were used to calculate this speed.

Regionally, 31% of November-March rainfall was attributed to tropical lows in summer months along the tropical edge, and the maximum attributable proportion was 66% in south west Angola. This compares well to previous studies based in India and Australia. In India 44% of June-September rainfall has been attributed to low pressure areas (Hunt and Fletcher, 2019), with a maximum attributable proportion of 60 - 70% near the north west coast of India. In northern Australia, between 50 and 60% of annual rainfall in the Pilbara has been attributed to closed low pressure systems (Lavender and Abbs, 2013). Taken together, these results imply that in the regions where they form frequently over land, tropical lows form a vital component of the local rainfall climatology.

Generally, tropical lows are more numerous in the cool phase of ENSO than the warm phase, although there is a mode of variability that disrupts this relationship, possibly linked to the Tropical Easterly Jet. Together with the reduction

in tropical low frequency comes a northward shift in tropical low density and rainfall during El Niño and a southward shift during La Niña. This result traces the well known regional ENSO signal to a class of synoptic rain-bearing systems, with the majority of the ENSO rainfall anomaly to the west of 30°E linked to tropical lows (Figure 4.12). These findings are consistent with studies of the dynamical response of the southern African atmosphere to ENSO. Ratna et al. (2013) characterise this response as a shift in the Walker Circulation over equatorial regions, which leads to enhanced upper level convergence (divergence) over southern Africa during the El Niño (La Niña) phase. This response would lead to a less (more) favourable environment for tropical low formation, since tropical lows feature surface convergence and upper level divergence (Chapter 3). Meanwhile, Lindesay et al. (1986) and Vignaud et al. (2009) find that the El Niño phase is associated with a northward shift in the subtropical jet and high-pressure belt, leading to increased subsidence over southern Africa. This subsidence would also inhibit tropical low formation to the south of our region of interest, pushing the locus of tropical lows further north.

This first objective study of tropical lows in southern Africa has identified and tracked cyclonic vortices in southern Africa, and computed the proportion of rainfall that falls within a 5 degree radius of these systems. It has not speculated on the dynamics of these systems or the mechanisms by which they form. Elsewhere in the tropics, particularly over the tropical ocean basins, equatorial wave theory has been successfully used to attribute tropical rainfall to different modes of wave activity, including Kelvin waves, Equatorial Rossby waves, mixed Rossby-Gravity waves, tropical disturbances etc (Wheeler and Kiladis, 2002). This chapter does not claim that these waves are not important over southern Africa. Indeed, many

of these waves, including equatorial Rossby waves and westerly mixed Rossby-Gravity (WMRG) waves, feature vorticity anomalies as key components of their dynamical signals (Kiladis et al., 2009). Therefore we hypothesise that the tropical lows studied in this chapter likely interact with equatorial waves, and in some cases may in fact be superpositions of various equatorial wave events.

Yang et al. (2018) studied the relationship between African Easterly Waves (AEWs) in North Africa and equatorial waves. The equatorial wave methodology rests on the linearization of the dry shallow water equations on a beta plane. The analytic solution to these equations forms a set of basis functions onto which the gridded datasets of observations of the real atmosphere may be projected. Yang et al. (2018) point out that the components of this projection are not necessarily the natural normal modes of an atmospheric structure such as a tropical low or an AEW. They found that, nevertheless, AEWs tend to bear some relation to and are intensified by WMRG waves, although AEWs move more slowly than the WMRGs. A similar study over southern Africa linking equatorial waves (of any flavour) with tropical lows would be a key contribution to the field, however is beyond the scope of this largely descriptive initial survey.

Further research to determine the processes that drive tropical lows and their relationship with ENSO would improve seasonal forecasts of rainfall in this highly variable region. More clarity on the moisture sources, thermodynamics and vertical structure of these tropical lows would provide insight on their formation and maintenance mechanisms. Diagnosing the representation of tropical lows in climate models will assist in judging the ability of these models to simulate the climate of southern Africa. In particular, the relationship between tropical lows and El Niño offers the prospect of understanding the model shortcomings described by Dieppois et al. (2015), who demonstrated that many climate models fail to

deliver the correct teleconnection between southern African precipitation and ENSO. The future change of these systems under anthropogenic climate change must be investigated in order to prepare the tropical edge to adapt to a warmer climate, following the example of Rastogi et al. (2018). This may be achieved by investigating whether they are well represented in CMIP historical simulations and if they are, whether they exhibit coherent change in the corresponding future scenarios.

4.9. Sensitivity analysis

The results of this chapter are sensitive to the vorticity threshold criterion applied in step 3 of the track selection algorithm described in Section 4.3. This criterion was that the filtered vertical mean relative vorticity must satisfy $\zeta < -3 \times 10^{-5} \text{ s}^{-1}$ in at least one 6-hourly time-step. To test the sensitivity to this threshold, all analysis was rerun with threshold values of $2.5 \times 10^{-5} \text{ s}^{-1}$ and $3.5 \times 10^{-5} \text{ s}^{-1}$.

When the vorticity criterion was weakened, 38-50% more tropical low events are identified, and when it was strengthened, about 30% fewer tracks were present. However, this increase impacted the distribution of TTTs uniformly across space and time. Therefore, all spatial features in the figures remained qualitatively similar as the threshold was varied. The change in key aggregated statistics of all tropical lows is given in table 4.3. When a weaker threshold is used, we see more shorter duration, less intense and faster tracks, and a greater proportion of rainfall attributed to tropical lows. The opposite occurs when the threshold is increased. The relationships with ENSO and the TEJ remain significant at a $p < 0.05$ level, although the correlation between ENSO and tropical lows drops in ERA-5 and MERRA-2 when weaker tropical lows are included in the dataset.

Table 4.3. Sensitivity Analysis. The first three columns show how statistics described by the row name change in each reanalysis when the vorticity threshold is weakened, as compared to the original threshold. The second three columns show the same for when the threshold is strengthened.

Threshold (10^{-5} s^{-1})	2.5	2.5	2.5	3.5	3.5	3.5
Reanalysis	era-interim	era-5	merra2	era-i	era-5	merra2
Track Statistics						
Number per year	50%	38%	40%	-33%	-30%	-29%
Intensity	-8%	-7%	-8%	9%	8%	8%
Longevity	-11%	-9%	-10%	12%	10%	11%
Easterly Track Speed	15%	9%	10%	-14%	-9%	-8%
Interannual						
# of threshold days/yr	53%	42%	38%	-32%	-28%	-27%
ENSO Anti-correlation	-3%	-27%	-16%	-3%	6%	-3%
Rainfall						
Chirps Ave. Prop.	11%	7%	9%	-12%	-9%	-8%
Chirps Max. Prop.	6%	3%	5%	-9%	-6%	-4%
TAMSAT Ave. Prop.	10%	8%	9%	-12%	-9%	-10%
TAMSAT Max. Prop.	8%	3%	5%	-10%	-6%	-5%

It is worth noting that the proportion of rainfall attributable to tropical lows is much less sensitive to the threshold than the total number of tropical lows identified is.

4.10. Acknowledgements

This work has been generously supported by the Origin Foundation John Monash Scholarship and by the Future Climate for Africa UMFULA project, with financial support from the U.K. Natural Environment Research Council (NERC) ME/M020207/1, and the UK Government's Department for International Development (DfID). The data used in this analysis were provided by the Copernicus Climate Change Service (<https://cds.climate.copernicus.eu/cdsapp#!/home>), the European Centre for Medium Range Weather Forecasts (<https://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=pl/>), the Climate Hazards Group (<https://chc.ucsb.edu/data/chirps/#%20Data>), NASA

(<https://disc.sci.gsfc.nasa.gov/daac-bin/FTPSubset.pl>) and TAMSAT (<https://www.tamsat.org.uk/data/rfe/index.cgi>). The authors thank Neil Hart for access to the MetBot output data.

4.11. Afterword to Chapter 4

In this chapter, the tropical lows that were identified in Chapter 3 were considered in more detail. The domain of study was no longer restricted to the Angola Low region. It was found that the spatial distribution of tropical lows peaks in the locus of the Angola Low. However, these systems frequently exist outside of this region. Interannual rainfall variability was shown to be sensitive to the number of tropical lows identified per season, with the majority of the ENSO rainfall signal delivered in the direct vicinity tropical lows. This chapter addressed the tropical lows components of Aims 1, 2 and 4.

An algorithm was optimised to track tropical low systems through gridded reanalysis data, and to attribute rainfall to these systems. This algorithm will be utilised in Chapter 6 to track tropical lows in historical and future CMIP5 models. The climatologies presented in this Chapter will be used as a reference for validation of the historical model simulations.

4.11.1. Local/non-local decomposition in the interannual regression

In Chapter 3, it was demonstrated that the ATL modulates interannual rainfall over southern Africa. More precisely, variability in the residual of interannual rainfall after regression onto an ENSO index could be explained by variability in the residual of the ATL index after regression onto the ENSO index. This was demonstrated in Figure 3.8. This raises the question: how does the ATL modulates interannual rainfall variability? In other words, what is the mechanism associated with this relationship? There are two distinct possibilities: a local

mechanism and a non-local mechanism. A local mechanism would consist of increased rainfall in seasons with more tropical lows due to rainfall that occurs near the centroid of tropical lows, associated with the convection and instability of the storm itself. The non-local mechanism would be associated with increased moisture flux into the southern African region caused by circulation associated with tropical lows. This increased moisture flux may then cause rainfall away from the centres of the tropical lows. These proposed mechanisms are not mutually exclusive, since moisture that falls as ‘direct’ rainfall would often also be moved from the ocean to the land via circulation associated with the tropical lows. However, the question boils down to whether the component of rainfall that varies with the ATLI is only associated with rainfall near tropical low centroids, or whether distant locations may be affected as well.

The analysis presented in Section 4.7 is ideally posed for answering this question. This analysis decomposes rainfall into a component that falls 5° from a tropical low centroid, and a component that does not. Figure 4.13 presents the results of Figure 3.8 with respect to this decomposition, using the CHIRPS and ERA-5 datasets. Here, CHIRPS is averaged over the southern African landmass south of 15° S. The ATLI index is replaced by the interannual tropical low index presented in Figure 4.5. Figure 4.13 then shows the regressions of the full, tropical low and remainder rainfall components on the Niño 3.4 SST index and the tropical low index. As in Figure 4.5, the skill of each regression is indicated by the match between the colours of the points and lines in each panel.

The left panels of Figure 4.13 are slightly different from Figure 3.8 since different reanalysis and precipitation datasets have been used, however the statistical significance of the relationship still holds. In the decomposition, the dependence on the tropical low index is significant only for the component of rainfall that

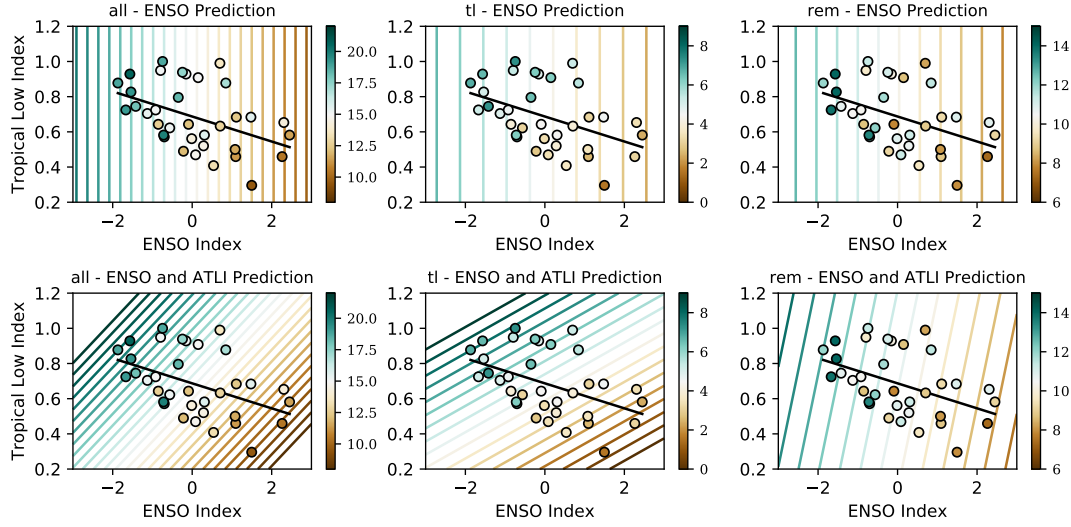


Figure 4.13. Scatter plots of November - March Niño 3.4 SST index and tropical low index from Chapter 4. Point colours represent CHIRPS precipitation (mm/pentad) over African mainland south of 15° with tropical low rainfall decomposition applied. Left column: all rainfall, centre column: tropical low rainfall, and right column: remainder rainfall. Line colours represent (top) precipitation predicted by the Niño 3.4 SST index only regression (column 1 of Table 3.2) and (bottom) precipitation predicted by the Niño 3.4 SST and ATLI regression (column 2 of Table 3.2). The black line shows the predicted tropical low index for each value of the Niño 3.4 SST index.

occurs close to tropical lows and not for the remainder component. This suggests that the relationship between the Angola Low and rainfall is local and direct. Therefore, seasons with a higher ATL index feature more tropical lows, which causes more precipitation to occur near the tropical low centroids. Flux of excess moisture due to the Angola Low into locations that are far away from tropical low centroids does not significantly increase interannual rainfall.

Meanwhile the ENSO signal is evenly spread across the two components of the decomposition. This implies that the ENSO signal is delivered both by tropical low rainfall and the remainder rainfall component. This finding is consistent with the results shown in Figure 4.12, which found that direct tropical low rainfall

Table 4.4. R^2 values for full and partial regressions of interannual indices described in text. Residuals refer to residuals of regression onto the Niño 3.4 SST index. Bold indicates statistical significance in the sign of the regression coefficient at the $p < 0.05$ level.

Reanalysis	Dependent Variable	Independent Variable	All Rainfall	Tropical Low Component	Remainder Component
ERA-5	TL Index	ENSO	0.20	0.20	0.20
	Rainfall	ENSO	0.51	0.43	0.37
	Rain resid.	TL resid.	0.19	0.46	0.01
ERA-Interim	TL Index	ENSO	0.11	0.11	0.11
	Rainfall	ENSO	0.53	0.36	0.41
	Rain resid.	TL resid.	0.21	0.35	0.03
MERRA2	TL Index	ENSO	0.14	0.14	0.14
	Rainfall	ENSO	0.51	0.41	0.40
	Rain resid.	TL resid.	0.23	0.39	0.03

dominates the ENSO signal in the west half of the tropical edge, and not in the east.

Table 4.4 shows that these findings are consistent across reanalysis products. These statistics indicate that the inclusion of the tropical low index in the rainfall regression explains a further 20% of the full interannual precipitation variance after ENSO, an extra 40% of the TL-related variance and none of the remainder precipitation variance across the three reanalysis products. This analysis suggests that the relationship between the Angola Tropical Low and interannual rainfall variability is predominantly local, and non-local effects are not significant. This result is important because a common climatological view of the significance of the Angola Low is that it acts to pull moisture out of the south Indian and south Atlantic maritime atmospheres onto the continent. This conceptual model is based on the climatological mean moisture flux, which resembles the low level winds shown in Figure 3.1. The current analysis has shown that on synoptic time-scales, the behaviour of tropical lows is quite different to this picture.

CHAPTER 5

Drylines in Southern Africa: rediscovering the Congo Air Boundary

Projected rainfall decline in southern Africa is likely to be highly sensitive to subtleties in the local atmospheric circulation. In an effort to understand the regional circulation complexities, a novel algorithm is developed to identify the Congo Air Boundary (CAB) in ERA-5, a high resolution reanalysis dataset. The CAB, a forgotten feature of the circulation, is defined in the Austral spring and early summer, using surface humidity gradients and near-surface wind convergence lines, and is found to be an indicator of the location of the southern edge of the African rain belt. A related combined convergence line - dryline, described in this chapter as the Kalahari Discontinuity (KD), is also identified. It is established that either a dryline CAB or KD is present in southern Africa for over 95% of days between August and December, with arc-lengths typically exceeding 10 degrees. The seasonal and diurnal cycles of the CAB and the KD are presented and their prevalence in station observational data is confirmed. The interannual variability of the CAB latitude and detection frequency is found to explain at least 55% of interannual spring rainfall variability in southern Africa between 15° to 25°S. Links are established with the Angola and Kalahari heat lows, and tropical temperate trough events.

5.1. Foreword to Chapter 5

Chapter 5 introduced the second class of synoptic weather system that this thesis focuses on, namely drylines and convergence lines associated with the CAB and the KD. These features operate as boundaries of the region where tropical rainfall may occur, as they separate saturated, moist convectively unstable air-masses from dry subsiding desert air. It was found that either the dryline CAB or the dryline KD is almost always present in early summer. The latitude and frequency of drylines associated with the CAB was shown to be strongly anti-correlated with interannual rainfall variability, contributing towards Aim 4. An algorithm was developed to identify these near-surface discontinuities, in accordance with Aim 1, using the ERA-5 reanalysis dataset. This algorithm will be modified and applied to CMIP5 models in order to obtain projections of CAB future change in Chapter 6. The climatological distribution of drylines and convergence lines was presented (Aim 2) and a dynamical link between the CAB and the heat lows described in Chapter 3 was established (Aim 3). This chapter has been accepted for publication in the Journal of Climate in August 2019.

5.2. Introduction

Southern Africa is one of the few inhabited regions of the world in which CMIP5 models project a robust decline in annual rainfall under future warming scenarios (Stocker et al., 2013). In this region, the maximum magnitudes of projected rainfall decline in occur in a latitudinal band between 10° and 20°S, primarily in the months of October, November and December. The RCP8.5 multi-model mean places rainfall decline at 6 mm/month per degree of warming (Lazenby et al., 2018). As southern Africa is highly dependent on rainfall for water, energy and food production (Conway et al., 2015), these rainfall changes

have the potential for dire consequences. Further investigations into the mechanisms and uncertainty associated with these projections are required to prepare for life in a warmer world.

The rainfall decline over southern Africa has been attributed to a delayed rainy season onset, causing a shortening in the overall duration in the rain season (Dunning et al., 2018). The onset of the rains results from the seasonal shift in the African rain-belt, which moves slowly south through the continent, with the onset of rainfall occurring earlier further north and later further south (Dunning et al., 2016). A wide, north-south oriented gradient in onset date is present over most of Angola, Zambia, and south east of the Democratic Republic of the Congo (DRC) (Dunning et al., 2016). However, the dynamics of the southern shift in the rain-belt in spring and early summer are not fully understood (Nicholson, 2018), and further investigation of the present day seasonal cycle is required before the mechanisms of future change may be fully characterised.

It is apparent from Figure 5.1 that in the climatological seasonal cycle, the rain-belt (green) is co-located with a broad band of near saturated relative humidity (blue contours). The humidity band moves north and south with the seasonal cycle of the rain-belt. On the northern and southern edges of the rain-belt, the humidity decreases rapidly, particularly from August to November in on the southern edge, and from March to June on the northern edge. These strong humidity gradients are often co-located with strong surface wind convergence (red contours). In this chapter, we show that the southern hemisphere spring wind convergence and humidity gradients are the climatological aggregation of the Congo Air Boundary (CAB). We identify the seasonal migration of the CAB as a key process that heralds the southward shift of the African rain-belt. We

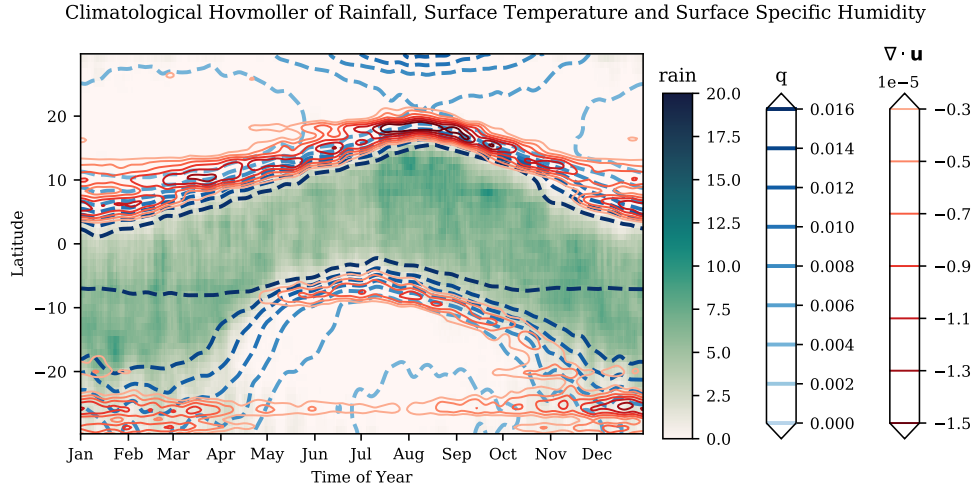


Figure 5.1. Climatology progression of the Southern African Rain Band, averaged between 18°E and 23°E. Green colours: Rainfall (mm/day, GPCP - NOAA (2012)). Dashed blue contours: specific humidity at 2m AGL (kg/kg). Solid red contours: Wind convergence 10m AGL (s^{-1}).

use reanalysis data to establish a baseline against which historical climate models may be compared in future studies.

Spring and summer rainfall in tropical southern Africa, and in particular in Angola and Zambia, has long been understood to be associated with the CAB (e.g. Torrance (1979)). The CAB, also referred to as the Zaire Air Boundary (ZAB), is traditionally defined as the location where the Low Level Westerlies (LLW), which originate as recurved Atlantic south-easterlies, meet with the easterly Indian Ocean trade winds at the surface (Taljaard, 1972; Torrance, 1979). Taljaard (1986) listed the placement of the inter-ocean trough as one of the ten factors that play a key role in the weather over southern Africa. Both Torrance (1979) and Leroux (2001) point out that the converging airmasses differ strongly in humidity, the LLW being close to saturation after their journey over the Congo rainforest, and the easterly trades having been ‘continentalized’ and dried out over the great escarpment.

However, while the CAB was frequently discussed in the 1970s and 1980s when weather forecasters doubled as climate researchers, it has receded in the academic literature of southern African present day climatology in the last two decades while prevailing in East African and paleo-climate literature (e.g. Nicholson (1996); Kizza et al. (2009); Tierney et al. (2011); Junginger et al. (2014)). However, as far as the authors are aware, no attempts have been made to identify the CAB in reanalysis datasets or in global or regional atmospheric models. This may in part be due to an inability of reanalysis and models to fully resolve the CAB, owing to their low resolution. However, the recent publication of the ERA-5 reanalysis dataset, at 30 km resolution, provides a solution to this problem.

This chapter intends to fill this gap by developing an algorithm to detect the CAB in daily ERA-5 reanalysis. Two formulations are presented, one that identifies convergence lines and the other that tracks drylines, and their relative robustness discussed. A second surface convergence line/dry line system, the Kalahari Discontinuity (KD) is also identified and studied. The diurnal and seasonal cycles of various properties of the CAB and the KD are considered, and the CAB's vital importance for regional precipitation is verified. The reanalysis-based CAB is compared favourably to literature of the 20th century. A link to the formation of tropical temperate troughs is discussed.

The remainder of this chapter will proceed as follows. The next section provides a comprehensive review of the 20th century literature relating to the CAB, while section 5.4 compares this literature with climatological means and sample days from ERA-5. In section 5.5, we develop an algorithm to identify the CAB and the KD using techniques adapted from image processing. In section 5.6, we analyse the seasonal cycle and properties of the CAB and the KD in reanalysis and automatic weather station (AWS) data. A moisture budget analysis is

used to study diurnal cycle of the maintenance of the sharp humidity gradient in section 5.7. In section 5.8 we examine the interannual influence of the CAB on rainfall, and its relationship with tropical temperate troughs (TTTs). The final section presents our conclusions.

5.3. Literature review

The concept of the CAB emerged in the scientific literature during attempts to describe the Inter-Tropical Convergence Zone (ITCZ) over the African continent. The ITCZ is traditionally defined as the region in which the trade winds from the northern and southern hemispheres converge (e.g. Van Heerden and Taljaard (1998)). However, the complexity of African climate leads competing processes to create zonally asymmetric wind patterns that do not fit with the ITCZ paradigm (Nicholson, 2018). One key asymmetry is that the LLW winds that enter the continent in the equatorial Congo separate the Harmattan trade winds in the north from the Indian ocean trade winds in the south. The common view in the 1980s was as follows (Leroux, 2001; Taljaard, 1972, 1981, 1986; Torrance, 1979): the LLW was deemed to be a recurvature of the South Atlantic trade winds, and so in the boreal summer, the convergence of the LLW with the Harmattan winds over the Sahel was deemed to be the ITCZ. However, since both the LLW and the Indian Ocean trades originated in the southern hemisphere, their confluence was not considered to be an ITCZ, and instead was labelled the ‘Congo Air Boundary’. Meanwhile, the ITCZ was defined as the convergence of the LLW with the Harmattans in north Africa, of the LLW with the north-easterly monsoon winds in East Africa, and of the south-easterly trades with the north-easterly monsoon in Tanzania and north Mozambique.

Taljaard (1953) first identified both summer convergence zones as possible candidates for the ITCZ, and identified the southernmost (the CAB) to have a more distinct discontinuity in atmospheric properties. However, later works, including Thompson (1965) and Torrance (1979), made a clear delineation between the ITCZ and the CAB. Most sources mark the summer position of the CAB as mostly zonally oriented, and with a slight south westerly - north easterly tilt, stretching from the Angolan coast to eastern Zambia. Summer CAB latitudes range from 20°-15°S. The winter CAB, usually pictured in July and August, is drawn at 10°S over Angola and tilts north-east to the rift valley at various inclinations. It is remarked upon by Torrance (1979) and Taljaard (1986) that the CAB is not a steady state feature, but rather exhibits intra-seasonal fluctuations north and south that can be associated with rainfall variability. Leroux (2001) describes the monthly average position of the CAB as it moves south (north) in Austral spring (autumn), and the role of moving polar highs in controlling the anomalous CAB movement.

There is some ambiguity as to whether the CAB was understood to exist embedded within the tropical rain-belt or along the southern edge of it. Both Leroux (2001) and Torrance (1979) contrast the high humidity Congo Air with the drier, 'continentalized' easterlies, and suggest that precipitation occurs primarily to the north of the CAB. Leroux (2001) claims that the southern African Easterly Jet (AEJ-S) is associated with the trades and overlies the westerlies, inhibiting rainfall immediately to the north of the CAB as well. However, Taljaard (1986) and Van Heerden and Taljaard (1998) describe the CAB as the mean location of the core of a broad zones where active weather dominates, and discusses tropical lows that travel westward along the CAB. However, tropical lows may only form within the moist airmass, and not on its edge (Chapter 3).

5.4. Circulation features

In order to ascertain whether the CAB as described above exists in the ERA-5 reanalysis, Figure 5.2 shows the climatological mean and selected representative days of surface specific humidity and dew point for each month between August and December. From August to October, the picture described by the above sources clearly holds, with convergence between dry easterlies and moist south-westerlies along a band that stretches from the centre axis of Angola to western Zambia. The humidity gradient and confluence zone are both broad in the climatological mean but remarkably sharp in the daily case studies, occurring over 100 km in places.

As the season progresses, the monthly mean specific humidity to the south of the CAB steadily increases. This is likely due to a combination of factors: the increased humidity of the easterly trades due to the warming SSTs in the Indian Ocean, reduced upper level subsidence as the subtropical jet moves south leading to the injection of a lesser amount of dry air into the lower troposphere, and the re-evaporation of the rain that falls due to events such as tropical temperate troughs (Harrison, 1984; Taljaard, 1986). By November, there is no longer a strong humidity gradient between the equatorial LLW and the easterly trades. Nor is the convergence between these two air-masses a bold feature of Figure 5.2. Instead, in November and December the main humidity gradient is oriented north west - south east, from northern Namibia into Botswana. This humidity gradient is also associated with convergence, now between the trades and the southerly Namibian Nocturnal Low Level Jet (Rife et al., 2010). In this chapter, we will call this feature the Kalahari Discontinuity (KD), and will contrast its behaviour and frequency to the CAB.

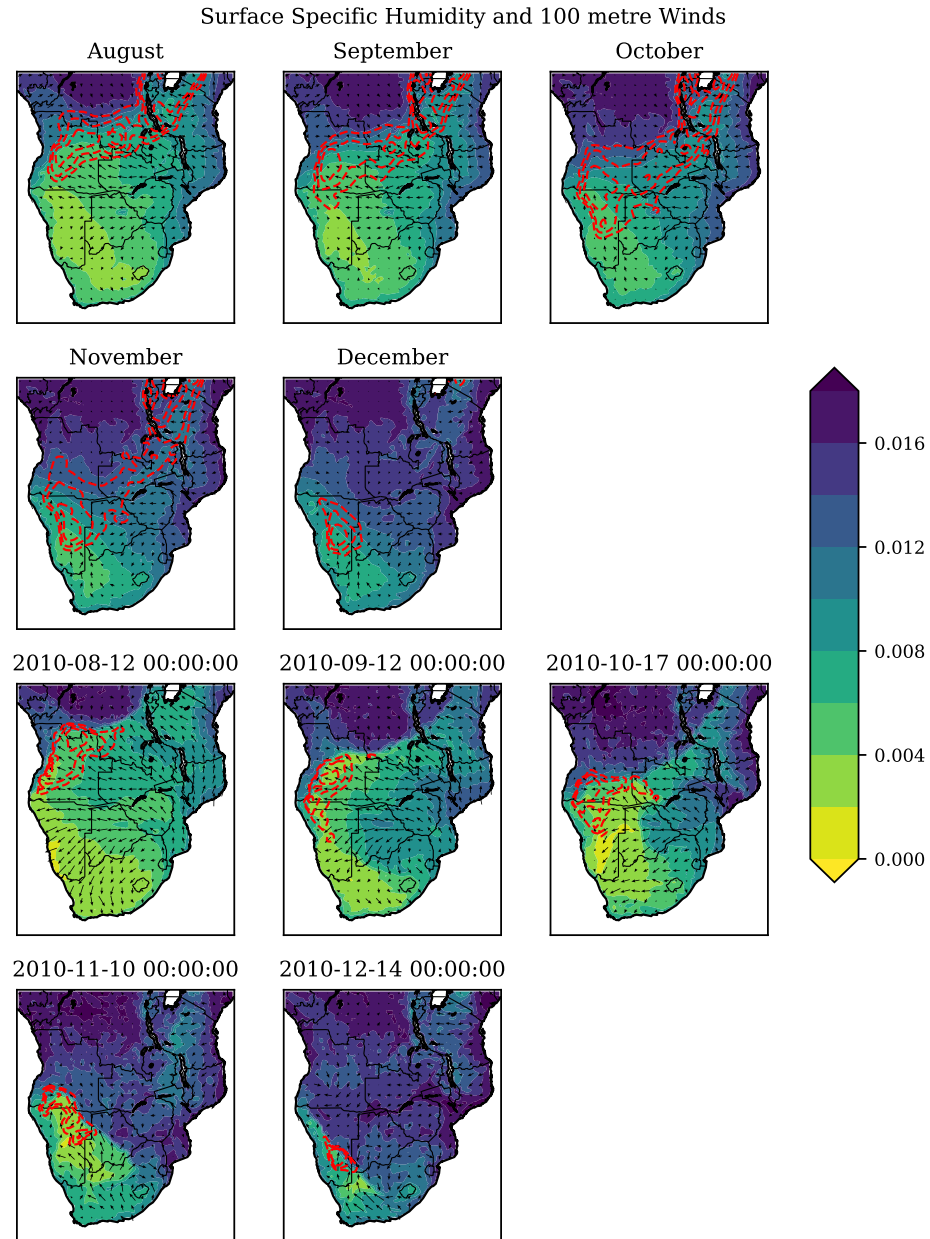


Figure 5.2. Climatological mean (top rows) and instantaneous (bottom rows) surface conditions. Climatological mean 2m specific humidity (kg/kg, colours), 100 m winds (arrows) and 12:00 UTC 2m potential temperature of the $[t_{\text{max-1}}, t_{\text{max-2}}, t_{\text{max-3}}]$ (red contours). Columns indicate months.

Figure 5.2 also shows contours of the locations of the maximum surface temperatures, which are consistently located immediately south of the CAB and to the west of the KD. This indicates a relationship between the surface discontinuities and the Angola and Kalahari heat lows, which will be further investigated in section 5.7.

To summarise, in late winter and early spring, the confluence of the equatorial LLW with the easterly trades over Southern Africa shows up a clear discontinuity in the monthly and daily ERA-5 reanalysis data. In November and December, the humidity of the easterly trades increases and the two air-masses appear to mix, such that the discontinuity is no longer immediately apparent. Thus the convectively active CAB featuring tropical lows as described by Van Heerden and Taljaard (1998) may well exist in this season. However, it does not show up in daily ERA-5 data as an abrupt discontinuity, either in humidity or surface winds. As this study focuses on convergence and drylines, our algorithm will not detect this flavour of CAB. Instead, we focus in November and December on the KD further south.

Criticisms of the use of the ITCZ paradigm over continental Africa are neatly summarised in Nicholson (2018) who argues that as the surface wind convergence maxima do not line up with the locations of the rainfall maxima, and so the traditional model of trade wind convergence setting the location of rainfall does not hold. It is easy to see why this is the case by examining Figure 5.1 - over the tropical rain belt, the relative humidity features very little gradient and is usually in excess of 95%. Thus, little wind convergence is required to reach saturation. However, surface wind convergence is co-located with the edge of the tropical rain belt, and so it is possible that the convergence zones of old are crucial in setting the location of the edge of the African rain-belt. This chapter will demonstrate

that this is indeed the case for the southern edge of the rain-belt in spring and early summer.

5.5. CAB identification algorithms

For the purposes of this study, the CAB is defined as a surface dryline and/or convergence line located at the northern edge of the easterly Indian Ocean trade winds, and the southern edge of the LLW. Similarly, the KD is a dryline or convergence line located at the southern edge of the Indian Ocean trade winds. Gradients in specific humidity were used to identify drylines over gradients in dewpoint temperature or relative humidity, due to the former's conservation properties as per Hoch and Markowski (2005). However, replacing specific humidity with either dew-point temperature or relative humidity gave similar results. 100 metre winds were chosen over winds on any pressure level because of the wide variation of surface pressures across southern Africa due to orography.

This section describes the methodology used to identify the CAB and the KD on a daily timescale in ERA-5 reanalysis (Copernicus Climate Change Service (C3S), 2017). First, strong gradients in relative humidity are identified each day using the Canny algorithm (Canny, 1986). The algorithm is as follows:

- (1) Apply a Gaussian filter with a 0.5 degree radius to smooth the 2D surface specific humidity field q ;
- (2) Calculate the gradient of the field ∇q ;
- (3) Calculate the magnitude M and direction θ of the gradient such that $\nabla q = M \times [\sin(\theta), \cos(\theta)]$;
- (4) Calculate the right and left partial derivatives of M in the θ direction using the finite difference method and using the closest horizontal, vertical and diagonal axes. For example, if $0 \leq \theta \leq 45$, the formulas

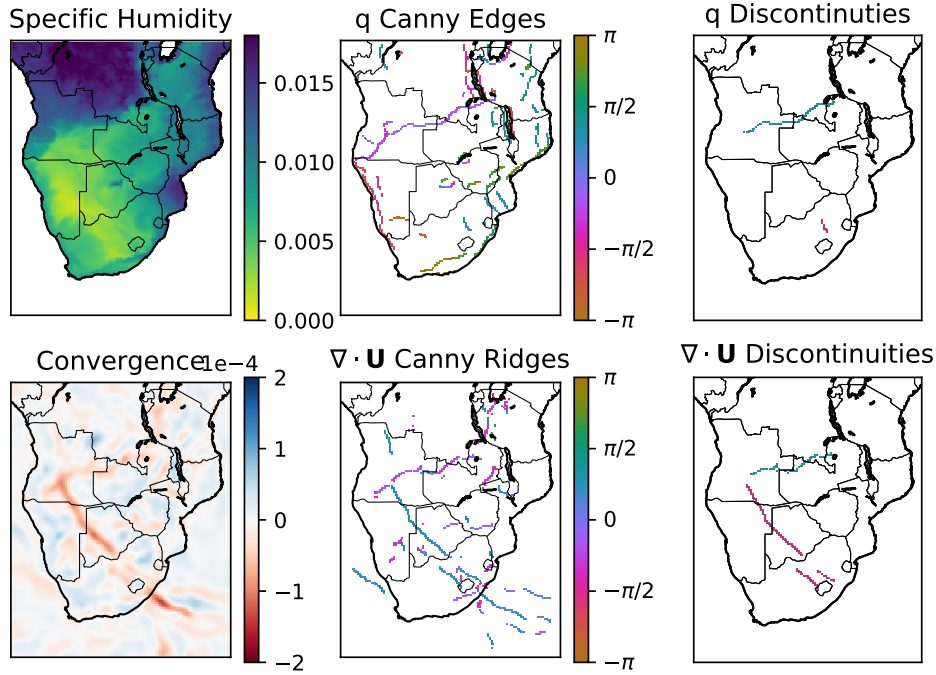


Figure 5.3. Demonstration of the CAB and KD Identification Algorithm. Top left: 2m specific humidity (kg/kg) on the 25th of September 2011. Top centre: corresponding Canny edges with colours indicating edge orientation angle (radians). Top Right: dryline Canny edges that meet the CAB thresholds (green). No dryline KDs were identified this day. Bottom left: 100m wind convergence on the 25th of September 2011. Bottom centre: corresponding Canny ridges with colours indicating edge orientation angle (radians). Bottom Right: convergence line Canny ridges that meet the CAB thresholds (green) and KD thresholds (red).

would be as follows:

$$\frac{\partial M}{\partial \theta^+} = \tan \theta (M(x + \delta, y + \delta) - M(x, y)) + (1 - \tan \theta) (M(x + \delta, y) - M(x, y))$$

$$\frac{\partial M}{\partial \theta^-} = \tan \theta (M(x, y) - M(x - \delta, y - \delta)) + (1 - \tan \theta) (M(x, y) - M(x - \delta, y))$$

- (5) Identify the locations where $\frac{\partial M}{\partial \theta^+} \leq 0$ and $\frac{\partial M}{\partial \theta^-} \geq 0$. In the continuous limit, this would be equivalent to $\frac{\partial M}{\partial \theta} = 0$, implying the presence of a local maximum along the θ axis; and

- (6) Return the locations where the above is satisfied and also $M > 0.003$ kg/kg per grid cell.

The resultant locations are the Canny edges and will, by construction, be lines in the discrete 2D space with widths of 1 grid cell, as shown in the top centre panel of Figure 5.3. The colours in this figure indicate the angle θ . This creates a set of points that are potentially locations of the CAB or the KD. To identify the CAB, these points are then filtered to retain locations where θ is between $-\frac{\pi}{4}$ and $\frac{\pi}{6}$ and restricted to latitudes of 5° - 18°S . To identify the KD, these points are then filtered to retain locations where θ is between $\frac{\pi}{6}$ and $\frac{\pi}{2}$ and restricted to latitudes south of 12°S . Connected segments consisting of less than 10 grid cells are also deemed spurious and removed from both definitions. These latitude and angle thresholds have been chosen carefully to include all instances of the CAB and KD while excluding spurious events associated with passing weather systems and coastal boundaries. These discontinuities will be referred to as the dryline CAB and the dryline KD from now on.

Convergence lines are identified using a similar method, detecting ridges in the convergence field rather than edges in humidity. This methodology combines the Canny (1986) algorithm with the approach of Weller et al. (2017) in identifying convergence lines. The Weller et al. (2017) methodology is used to find the angle of the convergence line, which allows the Canny (1986) methodology to be applied despite the fact that horizontal wind convergence is a scalar and so does not have an inherent direction associated with, unlike the humidity gradient vector. The approach is as follows:

- (1) Apply a Gaussian filter with 0.5 degree radius to smooth the 100 m wind fields u and v , and then calculate the convergence from these smoothed winds: $c = -\nabla \cdot \mathbf{u}$.
- (2) Threshold convergence at a value of $2 \times 10^{-5} s^{-1}$, and split the remaining field into connected regions.
- (3) For each point with convergence above the threshold, compute the convergence inertia tensor using only points inside the enveloping connected region and within a box of 10 by 10 grid points, equivalent to $2.5^\circ \times 2.5^\circ$ (as per Weller et al. (2017)).
- (4) Compute the eigenvalues of the convergence inertial tensor. The angle of the eigenvector with the largest eigenvalue will be the orientation of the major axis of convergence at this point, and that of the smaller eigenvalue will be the orientation of the minor axis of convergence.
- (5) Proceed as per the humidity algorithm, steps 4 - 6, replacing θ with the angle of the minor axis of convergence, and M with the magnitude of convergence.

Again the CAB and KD are identified from the resulting grid cells based on the angle, latitude and minimum size criteria described earlier. We call these delineations the convergence line CAB and the convergence lines KD. This gives two measures of where each discontinuity, as shown in the right column of Figure 5.3. On a day that a CAB or KD exists, we define its centre as the centroid of the largest connected set of flagged grid cells. The frequency of CAB and KD detection has been found to be sensitive to the thresholds applied throughout the algorithm. However, the other key properties of the CAB and KD considered

in this chapter, including seasonal and diurnal cycles, cross-sections and rainfall controls, remained qualitatively unchanged throughout a sensitivity analysis.

5.6. Properties of the CAB and KD

Having identified the CAB and KD using the methodologies described in section 5.5, this section presents statistics on their prevalence, location, intra-seasonal and interannual variability based on both the dryline and convergence line formulations. We confirm that the CAB and the KD are real phenomena and not reanalysis artefacts by considering 17 automatic weather stations, and consider vertical north-south cross sections of composites around the centroid of the CAB.

Figure 5.4 shows monthly heat-maps of the CAB and KD locations using the two identification methodologies described in section 5.5. The seasonal southward shift in the CAB latitude is easily seen as the locus moves from North Angola and the DRC in August through to south Angola in November. The frequency of the CAB also decreases over the season, lessening in October and being almost non-existent by December. The dryline CAB is detected approximately 20% more frequently than the convergence line CAB.

The convergence line KD is evident in each month, and does not move substantially north-south or east-west. However, in August and September it is rarely associated with a dryline KD, and from October onwards the dryline KD becomes more frequent. The dryline KD ramps up as both measures of the CAB ramps down, suggesting that the dryline KD replaces the dryline CAB in the later months.

These results are consistent with the seasonal cycles of latitude, frequency and width shown in Figure 5.5. The top panel of Figure 5.5 demonstrates that the

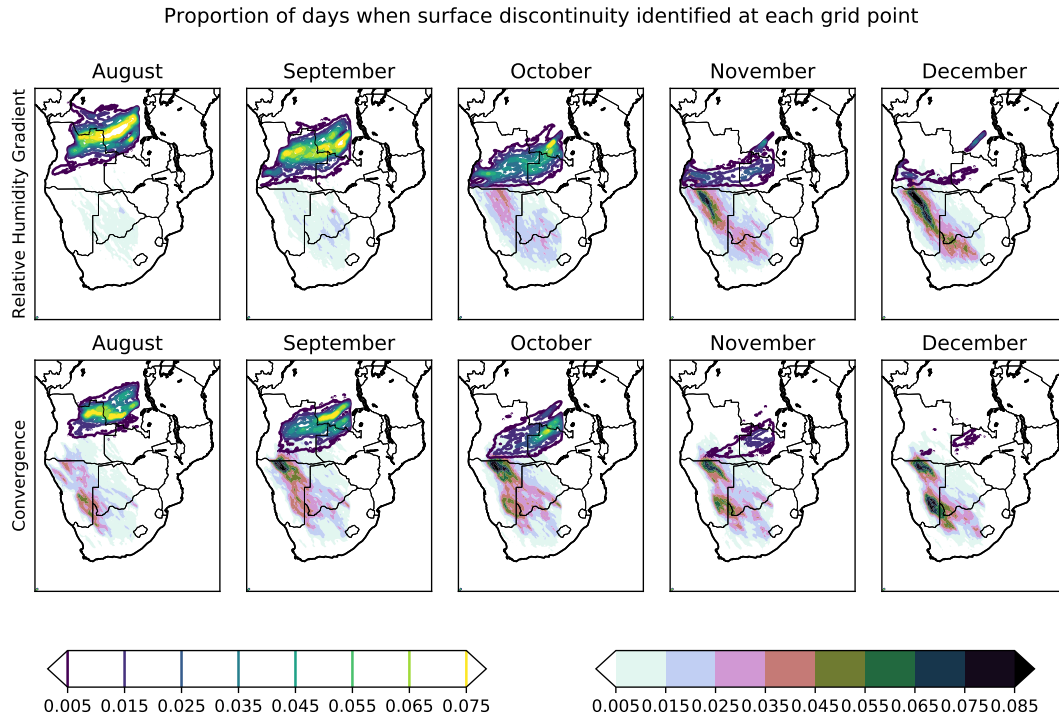


Figure 5.4. Climatological spatial distribution map of CAB and KD locations by month. Top row: dryline CAB and KD. Bottom row: convergence line CAB and KD. Columns indicate months as labelled. Contour lines indicate the CAB, while filled contours show the KD. Units are frequency per grid cell per month in each case.

dryline CAB is usually always present in August and September, and its frequency declines over October and November, until it is rarely detected in December. The convergence line CAB is consistently about 20% less frequent than the dryline CAB from August to November. Meanwhile, the dryline KD is present 20% of the time at the start of August, which increases to 100% by November.

The latitude of both CAB centroids, shown in the centre panel of Figure 5.5, moves steadily south from 7°S at the start of August to the start of November in a linear fashion, with a slope of 2 degrees of latitude per month. At the start of November the latitude of the CAB levels out. There is a large amount of noise in

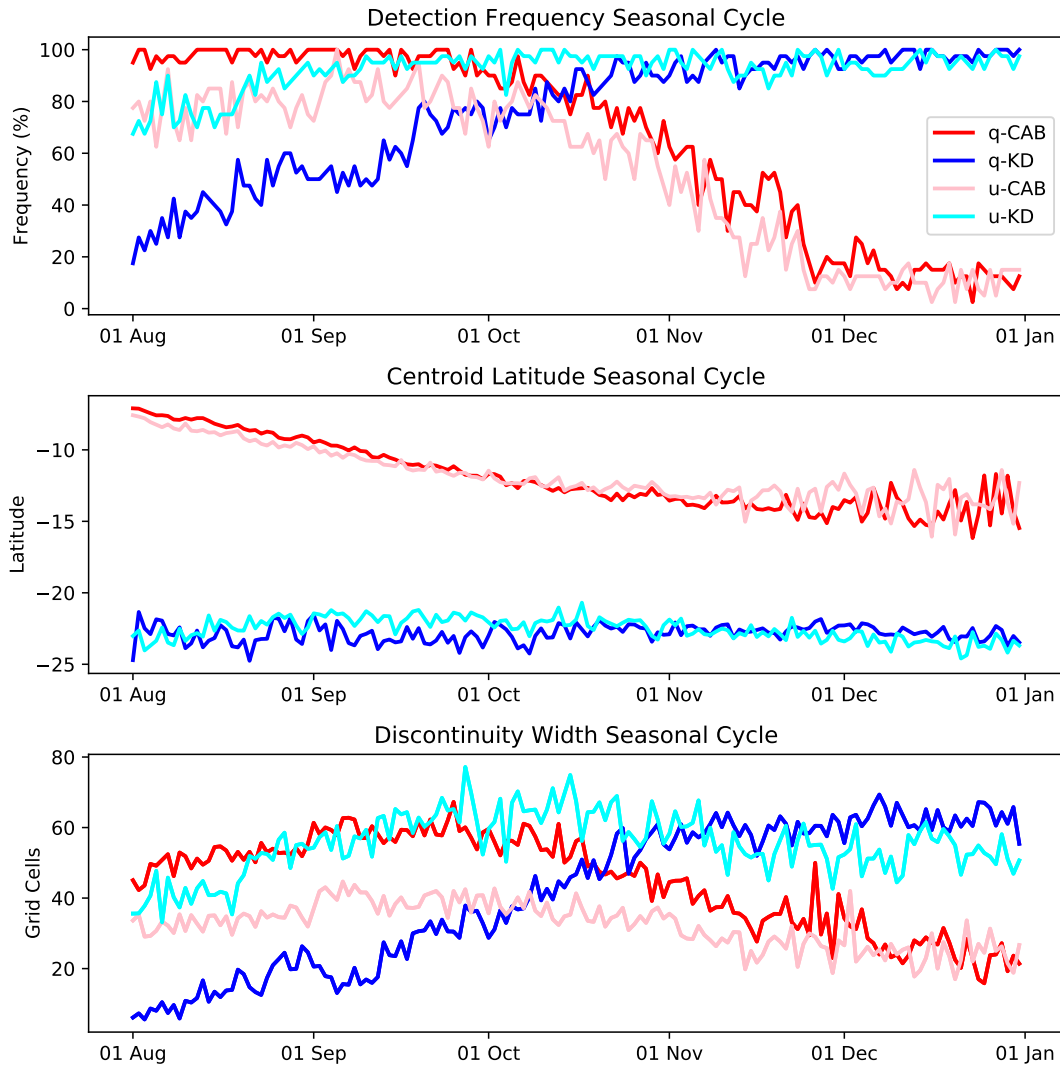


Figure 5.5. Top Panel: Climatological frequency that a CAB or KD is identified, by day of the year. Centre panel: climatological mean latitude of the CAB and KD by day of the year. Bottom panel: climatological mean number of CAB and KD grid cells identified on each day of the year. This is a measure of the length of the CAB. Dark blue: dryline KD; light blue: convergence line KD; red: dryline CAB; and pink: convergence line CAB.

the estimates of the CAB latitudes in December, which is likely due to the low frequency at which either CAB is detected in this month. The bottom panel of Figure 5.5 shows the climatological average number of CAB and KD grid cells

identified each day of the year. This is a proxy for the length of the discontinuity - if the discontinuity was purely horizontal, then 40 grid cells would mean that a discontinuity was 10 degrees long, for example. This measure follows a similar seasonal cycle to the frequency, with the CAB lengths decreasing over the season, and the KD lengths increasing. The convergence line CAB is typically two thirds as long as the dryline CAB from August to October. At all times, there is usually a dryline CAB or KD spanning on average 60 grid cells, equivalent to about 15 degrees.

Thus we have shown that the CAB and the KD are significant features of the southern African spring climatology, with at least one present for more than 80% of days and often spanning half the zonal extent of the continent. Figure 5.6 provides further insight into the vertical and diurnal structure of the CAB, showing vertical north-south composite cross-sections of specific humidity, zonal and meridional winds, potential temperature and vertical velocity (ω) at 8 times of the day for the month of September. Equivalent east-west cross sections were examined for the dryline KD in November, and found to be qualitatively similar. These are not shown for brevity. The composites in Figure 5.6 have been centred around the 00:00 UTC position of the dryline CAB, located at $x = 0$ and indicated by the vertical black line. The region to the north of the CAB is displayed to the left of the black line, with high humidity, cooler temperatures and north-westerly winds. Right of the black line is the region to the south of the CAB, with easterly winds, low humidity and very high midday temperatures. The temperature maximum is located approximately 200 km south of the dryline CAB.

The strong meridional humidity gradient is present at all times of the day, but is strongest at midnight and weakest at midday. At 12:00 UTC, surface wind

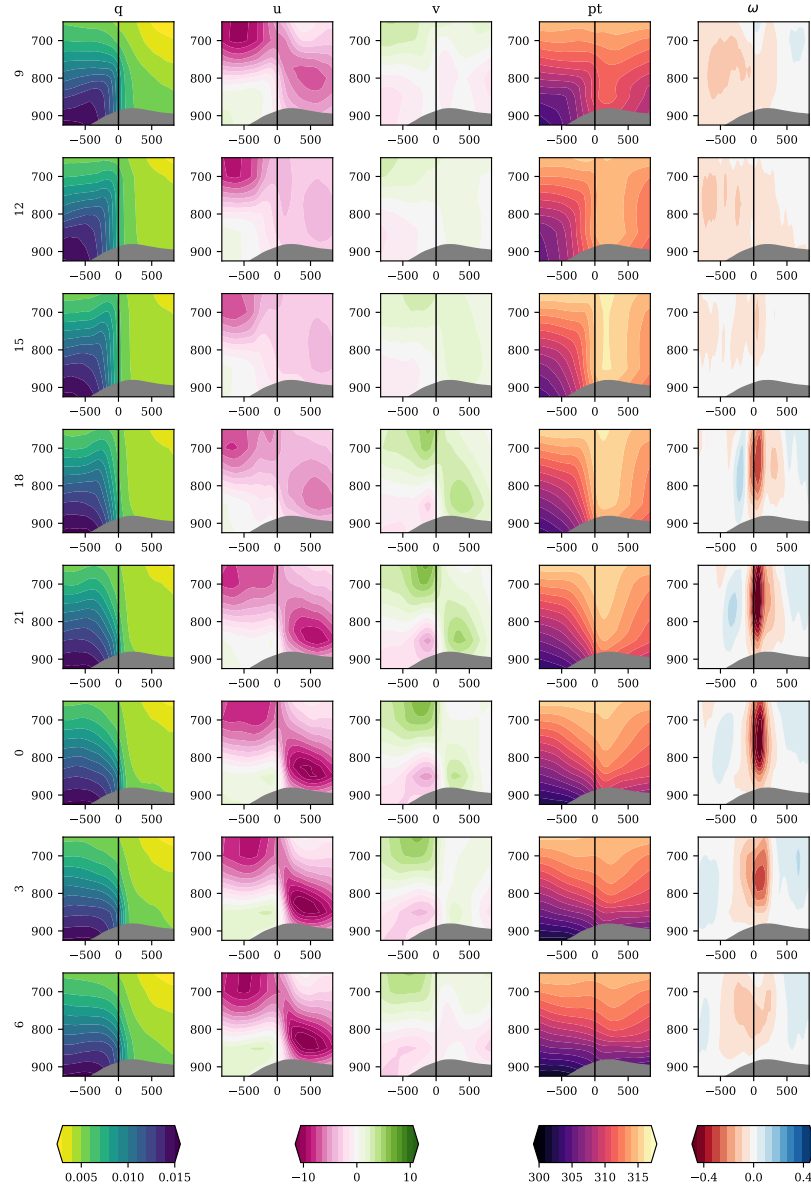


Figure 5.6. Diurnal vertical north south composites centred around the dryline CAB. Columns represent variables: specific humidity (kg/kg), zonal winds (m/s), meridional winds (m/s), potential temperature (K) and uplift (Pa/s). Rows indicate time of day as labelled. The x axis is the displacement south of the CAB in kilometres, and the y axis is pressure.

speeds are low and the potential temperature profile is unstable, implying the presence of dry convection. Surface winds are suppressed by turbulent mixing, consistent with Parker et al. (2005). At 18:00 UTC the atmospheric stability increases and the winds pick up, directed inwards towards the temperature maximum and the centre of the heat low (also apparent in Figure 5.2). As they do so, the humidity maximum near the surface moves southwards. Overnight, the surface winds rotate cyclonically, with winds to the south of the CAB shifting from south-easterly to easterly, and the winds to the north shifting from northerly to north-westerly. This is consistent with these winds being partially driven by the low-level thermally direct inflow of the heat low, which, under the influence of the Coriolis force, rotate overnight to form a cyclonic vortex (Rácz and Smith, 1999, and Chapter 3). In the morning, the surface temperatures increase and so turbulence builds and the winds are inhibited again. In the mid-troposphere, the African Easterly Jet-south is present to the north of the CAB.

Uplift near the CAB is consistently negative (and so, upwards), consistent with the co-location of the CAB and the heat low. However, somewhat counter-intuitively, this uplift is strongest at night, and not in the morning when the atmosphere is most thermally unstable. We hypothesise that this is because the daytime convection is not explicitly resolved by the reanalysis model, and is instead provided by the turbulence scheme. This sub-grid scale dry convection is not included in the ω term, which only represents the explicitly resolved uplift associated with the wind convergence of the resolved horizontal winds, such that the continuity equation is closed. We shall return to this hypothesis in section 5.7.

Before proceeding further, we wish to address the possibility that these dry-lines and convergence lines are reanalysis artefacts as opposed to being features of

Index	Name	Logger ID	Country	Longitude	Latitude	n days (Dry)	n days (wet)
1	Muconda	E12477	Angola	21.3° E	10.6° S	50	51
2	Mwinilunga	856121	Zambia	24.4° E	11.7° S	94	202
3	Copperbelt University	858854	Zambia	28.2° E	12.8° S	31	95
4	Zambezi	856126	Zambia	23.1° E	13.5° S	92	253
5	Cusseque	858129	Angola	17.1° E	13.7° S	55	83
6	Dongwe	361095	Zambia	24.0° E	14.0° S	31	152
7	Kafue National Park - Tatayoyo	858244	Zambia	24.4° E	14.9° S	14	175
8	Kalabo	856135	Zambia	22.7° E	15.0° S	31	311
9	Orongo	31212	Namibia	15.3° E	17.7° S	252	119
10	Erichsfelde	31197	Namibia	16.9° E	21.6° S	463	102
11	Kalahari	112	Namibia	18.5° E	24.2° S	345	45
12	Sonop	31213	Namibia	18.9° E	19.0° S	129	376
13	Sandveld	31198	Namibia	19.1° E	22.0° S	311	254
14	Tsumkwe	31204	Namibia	20.4° E	19.6° S	85	382
15	Ghanzi	68204	Botswana	21.7° E	21.7° S	103	203
16	Tsabong	31204	Botswana	22.4° E	26.0° S	195	113
17	Mogobane	23	Botswana	25.7° E	25.0° S	72	165

Table 5.1. SASSCAL stations used in this study. Number of days column indicates data availability between August and December only

the true atmosphere. The number of observations being fed into reanalysis products in tropical southern Africa is limited, and so the reanalysis products rely heavily on their modelled components and assimilation schemes in these regions. Therefore, we study the representations of the CAB at 8 automatic weather stations (AWS) in the SASSCAL network across Angola and Zambia, and the KD at 9 stations across Namibia and Zambia (Helmschrot et al., 2015). Summaries of these sites are given in Table 5.1, and a map of their locations is shown in Figure 5.7. The 8 CAB sites are labelled in order of decreasing latitude, and the 9 KD sites by increasing longitude.

Like all AWS networks, the SASSCAL dataset is spatially sparse and all the stations are located within their own micro-climates that may not necessarily be comparable. Therefore, we consider the change in the measured weather condition as the reanalysis defined CAB (KD) passes from north to south (east to west) of

each station, rather than trying to identify spatial gradients between stations. For each of the 8 northern SASSCAL sites along the CAB, days were identified during which a dryline CAB grid cell was identified in the reanalysis either directly north or directly south of the station. Similarly for the 9 southern sites along the KD, days were identified when a dryline KD was east or west of the station. The upper two rows of Figure 5.8 then shows a scatter-plot of the midnight recorded specific humidity against dryline CAB latitude at the station longitude. Data-points represent all days when the CAB was present at this longitude and that data was available from the AWS. The marker colours show the 12:00 UTC potential temperature recorded at the station the day before. The lower three rows show similar scatterplots for the KD, replacing the north-south axis with east-west, and transposing the plots so that longitude is shown on the x-axis.

At all the stations, there is a transition as the CAB (KD) moves past the station (black line) from having lower humidity when the CAB (KD) is to the north (east) of the station, to higher humidity when the CAB (KD) is to the south (west). The suddenness of this transition indicates the sharpness of the CAB or KD, and varies across the stations. In most cases, the transition occurs over less than 5 degrees. In some, such as Zambezi, Muconda, Ghanzi and Mogobane, the transition occurs within 1 - 2 degrees.

The point colours at all stations in Figure 5.8 indicate that the warmest temperatures occur when the CAB (KD) is located 0 - 5° to the north (east) of the station. This is consistent with Figures 5.2 and 5.6, that also show a potential temperature maximum occurs immediately south of the CAB and west of the KD. Together with the results from specific humidity, this suggests that the CAB, as identified in ERA-5 reanalysis, is evident with the SASSCAL AWS dataset. Similar analysis of SASSCAL 10 metre winds (not shown) has found that

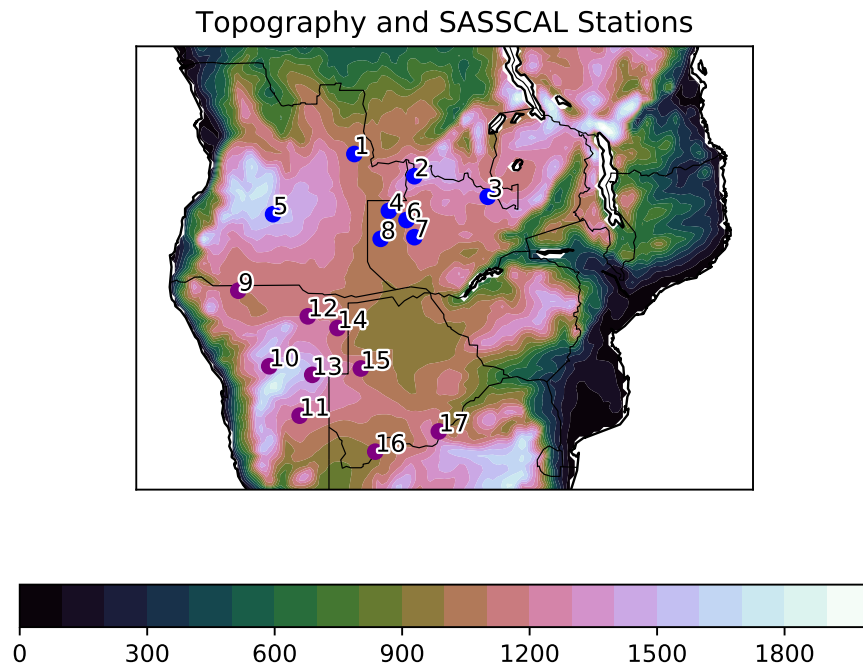


Figure 5.7. SASSCAL site locations, labelled as per Table 5.1. Colours indicate surface altitude (m). Blue: sites used to study the CAB. Purple: sites used to study the KD.

wind direction at 11 out of 17 stations notably changed direction on either side of the CAB. Considering that the measured winds are likely to be particularly susceptible to the micro-climates of the station locations, this suggests that the CAB and KD winds as represented in ERA-5 reanalysis are consistent with the observed atmosphere.

5.7. Moisture budget analysis of the diurnal cycle

The consistent proximity of the CAB and the edge of the rain-belt to the surface potential temperature maximum in Figures 5.1, 5.6 and 5.8 suggests that

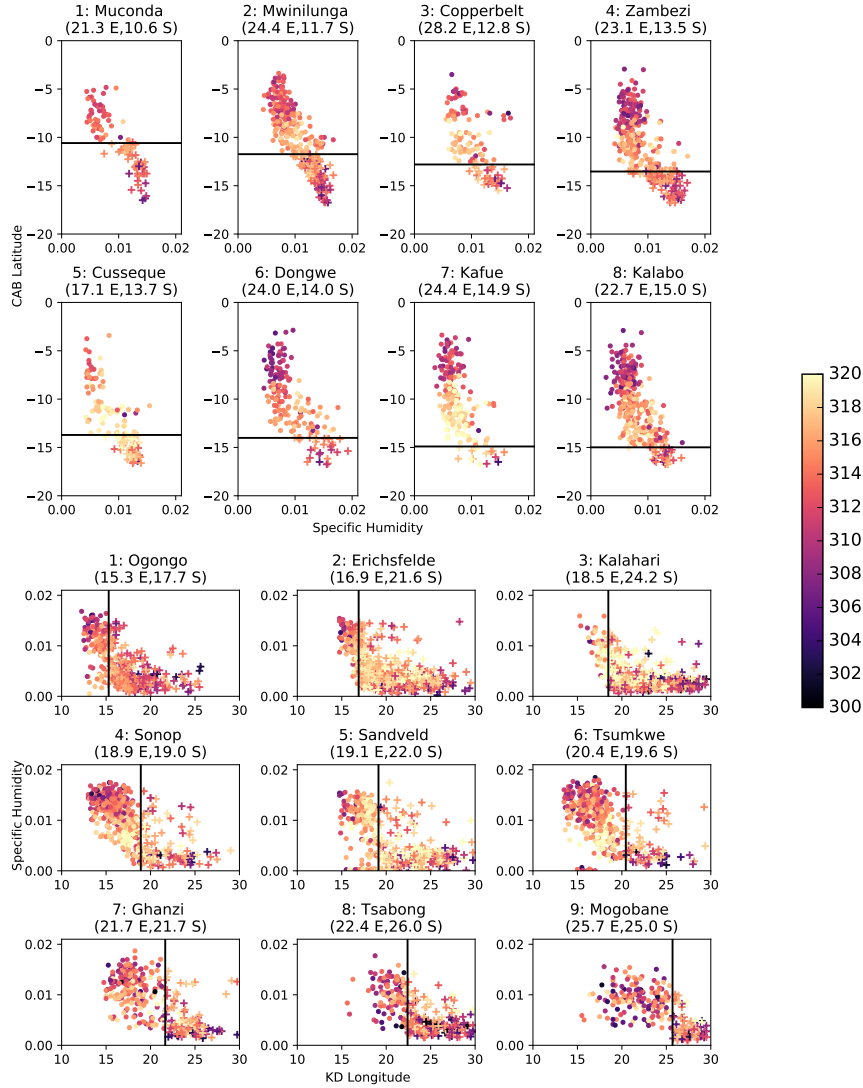


Figure 5.8. Top 2 rows: scatter-plots of SASSCAL specific humidity (kg/kg) against reanalysis dryline CAB latitude (degrees) on days where the dryline CAB exists at the same longitude as the station. Panels indicate stations as labelled in Table 5.1. x -axis: measured 00:00 specific humidity at station. y -axis: latitude of CAB at station longitude. Marker colours: measured potential temperature at station (K). Black line: Station latitude. Stations are presented in order of decreasing latitude. Dots (crosses) show data-points north (south) of the CAB. Bottom 3 rows: as per top rows but for the KD, with x showing reanalysis KD latitude and y axis showing 00:00 measured specific humidity.

there is a link between the location of the Angola Heat Low and the CAB. Such a link would provide a mechanism by which the Angola Heat Low influences the progression of rainfall onset in Southern Africa, as noted by Dunning et al. (2018). In this section, the relationship between the Heat Low and the CAB is explored using a moisture budget analysis.

The formulation of the moisture budget used in this chapter is as follows:

$$(2) \quad \underbrace{\frac{\partial q}{\partial t}}_{\text{Tendency}} = \underbrace{-\mathbf{u}_h \cdot \nabla_h q}_{\text{Horizontal Advection}} - \underbrace{\omega \frac{\partial q}{\partial p}}_{\text{Vertical Advection}} - \underbrace{c - \nabla \cdot q' \mathbf{u}'}_{\text{Residual}}$$

Where q , $\mathbf{u}_h$ and ω are the resolved specific humidity, horizontal and vertical winds respectively, in pressure coordinates, and q' and $\mathbf{u}'$ are the sub-grid scale eddy specific humidity and 3D winds. The left hand term represents the time derivative of specific humidity, the first term on the right its advection by large-scale horizontal winds and the second vertical advection. In the third term, the residual, c represents the addition and removal of moisture from the atmosphere by evaporation and precipitation such that $\int_{p_s}^0 c \, dp = E - P$. The second component of the residual represents the unresolved moisture transport by turbulent processes. This will be dominated by the vertical term, $\frac{\partial q' \omega'}{\partial p}$, representing sub-grid scale convection, turbulent mixing and boundary layer growth.

Figure 5.9 shows composites of the diurnal moisture budget terms, averaged around the centroid of the CAB. Evidently, the residual term dominates the moisture budget during the day. Horizontal advection is minimal since the horizontal winds are suppressed by turbulent mixing (Figure 5.6, Parker et al. (2005)). The spatial distribution of the residual term, as a moisture sink near the surface and a moisture source aloft, is representative of boundary layer growth driven by dry

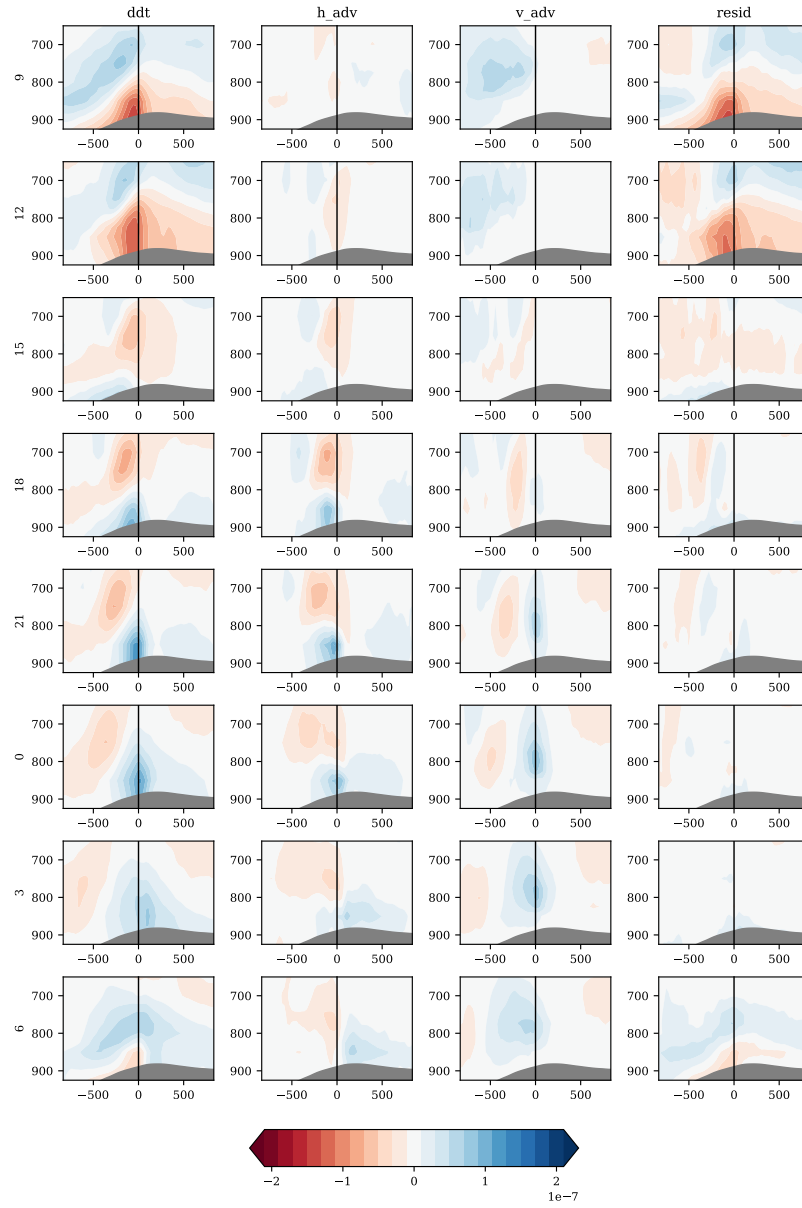


Figure 5.9. Diurnal vertical north south composites centred around the dryline CAB. Columns represent terms in the moisture budget: horizontal advection, vertical advection, time derivative and residual (turbulent transport + rainfall). Rows indicate time of day as labelled. Units are kg/kg/s. The x axis is the displacement south of the CAB in kilometres, and the y axis is pressure.

convection. In the early afternoon, the stratification of potential temperature is unstable (Figure 5.6), meaning that dry sub-grid scale convection is occurring and fuelling boundary layer growth. This residual term largely contains advection of moisture by sub-grid scale uplift and far exceeds the uplift due to resolved ω (third column). This confirms our hypothesis from Section 5.6, that the uplift associated with the heat lows is unresolved by the reanalysis and instead dealt with by the turbulence scheme. Specific humidity decreases in the lower atmosphere and increases aloft as moisture is lifted throughout the day. Boundary layer growth has its highest vertical extent 200 km south of the CAB. Comparing to Figure 5.6, this is the centre of the heat low, the location of maximum surface potential temperature and the maximum thermal instability.

In the evening and overnight, the turbulent fluxes die down and the tendency is balanced by horizontal advection. The impact of this advection is to advect the specific humidity south at the latitude of the CAB. Comparing to Figure 5.6 and remembering that the horizontal advection term is calculated as $u \frac{\partial q}{\partial x} + v \frac{\partial q}{\partial y}$, we conclude that the advection is driven by the southward component of the wind centred at 850 hPa to the north of the CAB (Figure 5.6, 3rd column), as they pass across the humidity gradient. The next morning, the boundary layer grows again, the horizontal winds (and hence horizontal advection) are suppressed, and the process is repeated.

Thus, the potential temperature maximum is intrinsically linked with the diurnal cycle of the CAB. During the day, vertical mixing associated with the unstable temperature profiles at the potential temperature maximum decreases the specific humidity near the land surface by mixing it upwards. Overnight, the north-westerly winds associated with the LLW strengthen and advect moisture on the southern edge of the CAB southwards. As the LLW are driven by differential

heating south of 6°N (Pokam et al., 2014), this process is also associated with the temperature maximum. Thus, by increasing and decreasing the humidity either side of the CAB edge, the heat low maintains the sharp humidity gradient.

A similar moisture budget analysis was performed for the KD in November, and the maintenance mechanism and diurnal cycle were qualitatively similar. In this case, it is the Kalahari Heat Low that maintains the humidity gradient.

5.8. Implications for rainfall in southern Africa

The CAB and the KD both divide the hot dry air to the south from moist, often saturated air to the north, both at the surface and aloft (Figure 5.6). It is clear, therefore, that in the vicinity of these discontinuities, rainfall ought only to occur to the north, as air to the south is far from saturated. Therefore, the location of the CAB and the KD have the potential to exhibit control on interannual rainfall. It is nevertheless well established that tropical rainfall may occur in subtropical southern Africa, in the form of TTTs Harrison (1984); Hart et al. (2013). In this section we determine the degree to which interannual variability of the CAB contributes to interannual variability of southern African spring rainfall, and investigate the relationship between the CAB, the KD and TTTs.

First, we consider the interannual relationship between the dryline CAB latitude and frequency for each month from August to December. The top two rows of Figure 5.10 show regressions of monthly precipitation on the monthly mean frequency and latitude of the CAB. Precipitation is calculated as the CHIRPS monthly mean between 15° E and 25° E, and between the tenth percentile range of the daily CAB location in that month and 25° (Funk et al., 2015). In August and September, when the CAB frequency is always close to 100%, the monthly precipitation is strongly anti-correlated with the CAB latitude ($p < 0.001$, $r = -0.55$

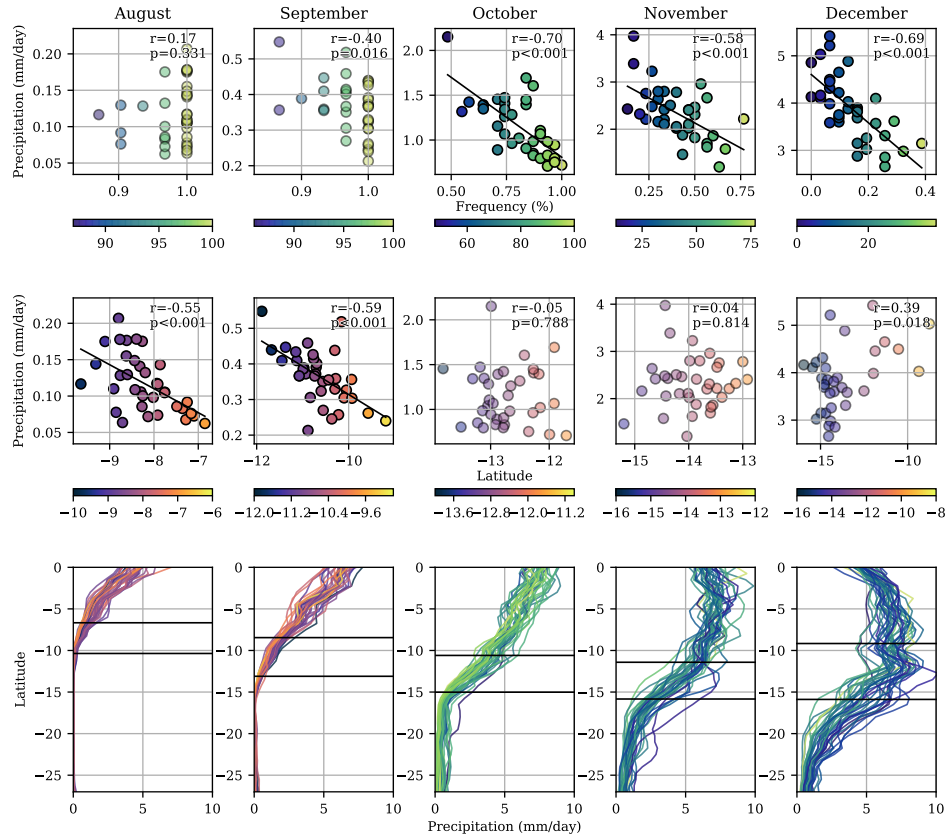


Figure 5.10. Interannual rainfall variability due to the CAB. Columns indicate month. Top row: Regression of rainfall onto CAB frequency. Rainfall is averaged from CHIRPS between longitudes 15°E and 25°E, and between the 10th percentile CAB latitude and 25°S. x -axis: frequency, y -axis: precipitation (mm/day). Black line indicates the linear regression of rainfall onto CAB monthly mean CAB latitude when the regression is significant at a $p < 0.001$ level. Colours also indicate CAB frequency for greater clarity. Centre row: as per top row with precipitation regressed on CAB latitude. Bottom row: latitudinal profiles of monthly rainfall for each year between 1981 and 2015. Line colour indicates CAB latitude (August and September) and CAB frequency (October, November and December) and correspond to the colour-bars in the middle and upper rows respectively. x -axis: precipitation (mm/day), y -axis: latitude. Horizontal black lines represent the 10th and 90th percentile latitude of the dryline CAB based on daily data in a given month. Rainfall averaged between 15°E and 25°E.

and $r = -0.59$). From October to December, the precipitation becomes anti-correlated with the CAB frequency ($p < 0.001$, $r = -0.70$, $r = -0.58$ and $r = -0.69$).

Together, these results imply that the CAB plays a primary control on the spring and early summer rainfall in southern Africa. Early in the season, more rainfall may occur in this region when the CAB is further south. Later, as the CAB begins to break down, there is more rain when the CAB is less frequent, as the moist air is able to move further south into the subtropics. We also found a relationship between the longitude of the dryline KD and rainfall in the same region in October and December ($p < 0.01$, $r = -0.46$ and $r = -0.55$), with less rainfall when the KD is further east (scatterplots not shown). Note that the north-most latitude of the averaging domain varies between climatological months, but not from year to year. This range is 6.7°S in August, 8.5°S in September, 10.6°S , 11.4°S in November and 9.2°S in December.

The relationship between precipitation and the CAB latitude and frequency is evident across all latitudes of the averaging domain, as shown in the lower panels of Figure 5.10. For each months, the edge 10 and 90 percentile range of the daily CAB location (black lines) enclose the southern edge of the tropical rain-belt, where the slope of the precipitation is strongest. In this region, colouring of the profiles changes from light to dark across the figure from left to right. The colouring of the August and September panels corresponds to the CAB latitude, as per the centre row of panels, and that of October - December corresponds to CAB frequency.

In August to September, rain to the south of the P90 CAB line is close to zero in Figure 5.10. From October to December, however, 0 - 1.5 mm of rain falls in this region per day. As the primary source of rainfall in subtropical southern Africa in summer is TTTs, we hypothesise that these weather events are the main source of this rainfall. Since both are key features of the spring climate of

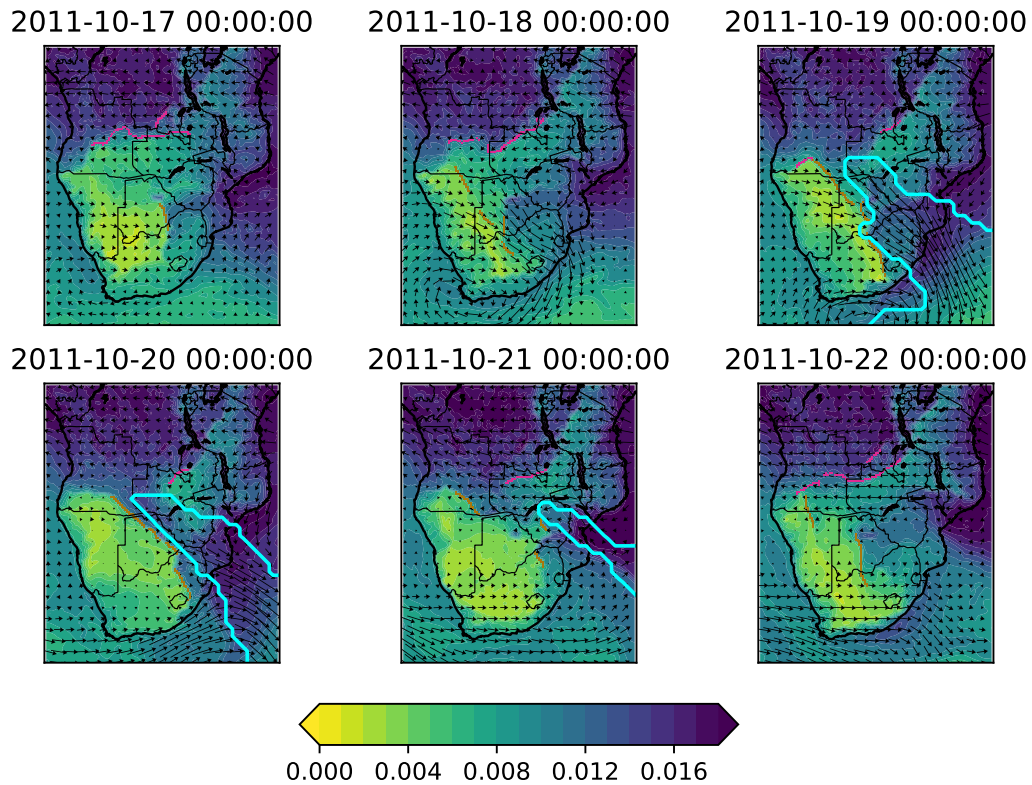


Figure 5.11. Surface specific humidity and integrated moisture flux on the days surrounding a TTT event. Colours: 2m specific humidity (kg/kg). Pink lines: dryline CAB. Brown lines: dryline KD. Blue contours and shading: TTT envelope. Vectors: Column integrated moisture flux.

southern Africa, we wish to investigate the relationship between TTTs and the CAB, along with the KD.

A case-study example of the interaction between the dryline CAB, the dryline KD and a TTT is shown in Figure 5.11. The TTT commenced on the 19th of October, 2011 (as identified by the MetBot, Hart et al. (2012)) and lasted for 3 days. Two days before the TTT was identified, a dryline CAB was present near 11°S, cutting diagonally across Angola and into the DRC from the south-west to the north-east. The day before the TTT formed, a dryline KD began to form along

eastern Namibia and western Botswana, while the dryline CAB was unchanged. A subtropical cyclone was generating northerly moisture flux over Botswana and South Africa. On the day that the TTT formed, the dryline CAB weakened and split into two small components located to the north-east and south-west its previous position. The northern edge of the TTT envelope was located in between the two CAB components. The northerly moisture flux associated with the TTT connected the subtropical cyclone to the moist convective air that is usually trapped to the north of the CAB and allowing it to flow south. The KD formed the south-western edge of the TTT. On the proceeding day, this set-up persisted, with both the TTT and the KD shifted to the east. On the 21st, the TTT moved offshore and by the 22nd, the CAB had reformed along a single axis, and the KD had mostly broken down.

Based on this analysis, we conclude that this TTT was in part the result of the interaction of a cyclone associated with the subtropical easterly jet breaking down the CAB and interacting with the moist tropical air to the north. The KD appeared to form in conjunction with the TTT, and it is possible that all the early season KDs are associated with tropical temperate troughs in this manner. Figure 5.12 indicates that TTTs are associated with an increase in the frequency and length of the KD, and a decrease in the frequency and length of the CAB. Interestingly, the decrease in the CAB length becomes significant at a $\alpha_{FDR} < 0.05$ level 7 days before the initiation of the TTT, suggesting that the CAB may be involved in the set-up of TTT events. It appears that TTTs may be the result of an interaction of the CAB with a wave in the subtropical westerly jet (Macron et al., 2014) with the CAB, that causes the CAB to break down and a KD to form in its place, allowing Congo air to flow south to create the characteristic

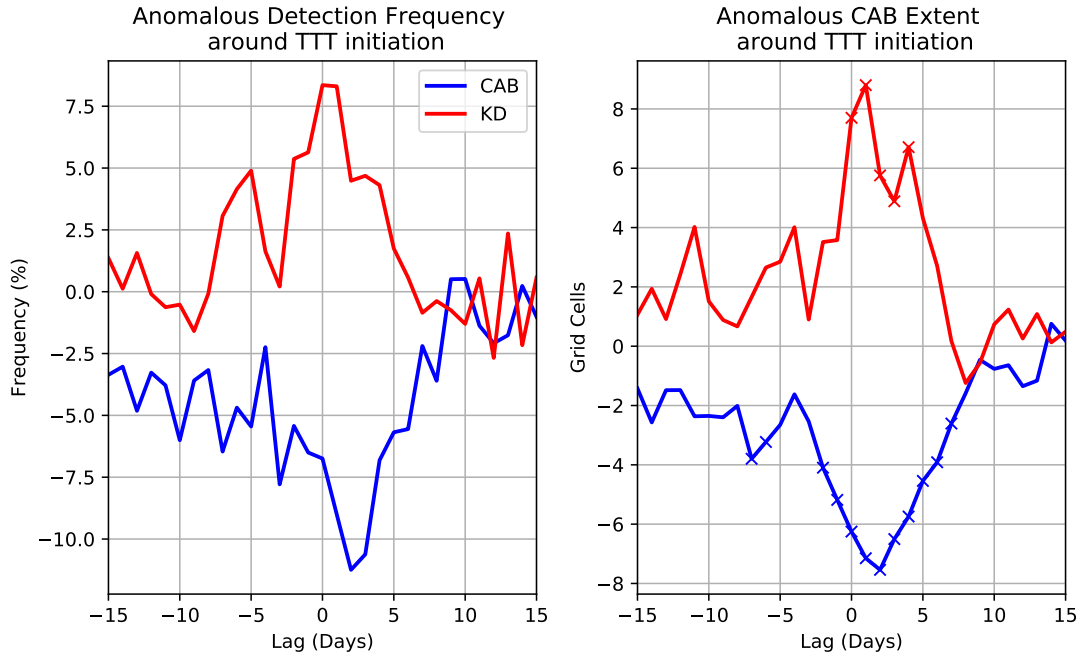


Figure 5.12. Anomalous CAB and KD properties lagged around TTT flag date. Left: anomalous frequency (%). Right: Anomalous number of grid cells identified per day. Anomalies have had the climatological seasonal cycle removed. Blue: dryline CAB, Red: dryline KD. On the right panel, markers are shown where the result is significant at a $\alpha_{FDR} < 0.05$ level, with the false discovery rate controlled for.

TTT southward moisture flux across the subtropical continent. Further work is required to verify this.

5.9. Discussion and conclusions

Surface drylines located along convergence lines at the edges of the easterly trade winds are persistent and important features of the southern African climatology. In spring, the dryline moves south with the trades along the CAB, forming a sharp boundary between the tropics and the subtropics. As the subtropical easterlies become steadily more humid, the CAB dryline breaks down and is replaced with the KD, a south-east to north-west oriented dryline along

the edge of the Kalahari desert in the west of the subcontinent. This dryline aligns with the south-western extent of the easterly trade winds. A schematic of the typical climate features in play before and after the CAB break down is shown in Figure 5.13.

We have found that properties of the CAB have significant and intuitive correlations with interannual rainfall over southern Africa. In the early season, the CAB latitude is strongly anti-correlated with rainfall averaged over the region that the CAB usually exists. This is likely to be a causal relationship, since saturated air may move and rain further south when the CAB stretches further over the continent. In the later seasons, the frequency of the CAB is anti-correlated with rainfall, such that more rainfall occurs when the CAB breaks down earlier. This relationship is more complex: it could be that the breakdown of the CAB allows more moisture to enter southern Africa from the Congo. However, it could also be the case that in seasons in which the low level easterlies become more humid at a faster rate, due to Indian Ocean SSTs, a reduction in subsidence or re-evaporation, the low level easterlies cause both an increase in rainfall and the breakdown of the CAB. Most likely, the correlation is due to a combination of these factors.

Interannual variability in southern Africa has previously been linked to other factors, including ENSO and the Atlantic and Indian Ocean highs. This chapter does not claim that the CAB supersedes these drivers or precipitation. Rather, the CAB may be involved in teleconnection processes, and so may deliver remote signals on a synoptic timescale. Further investigation on the sources of interannual variability in the properties of the CAB and the KD is required, with particular focus on modes of variability deemed important for early wet season rainfall.

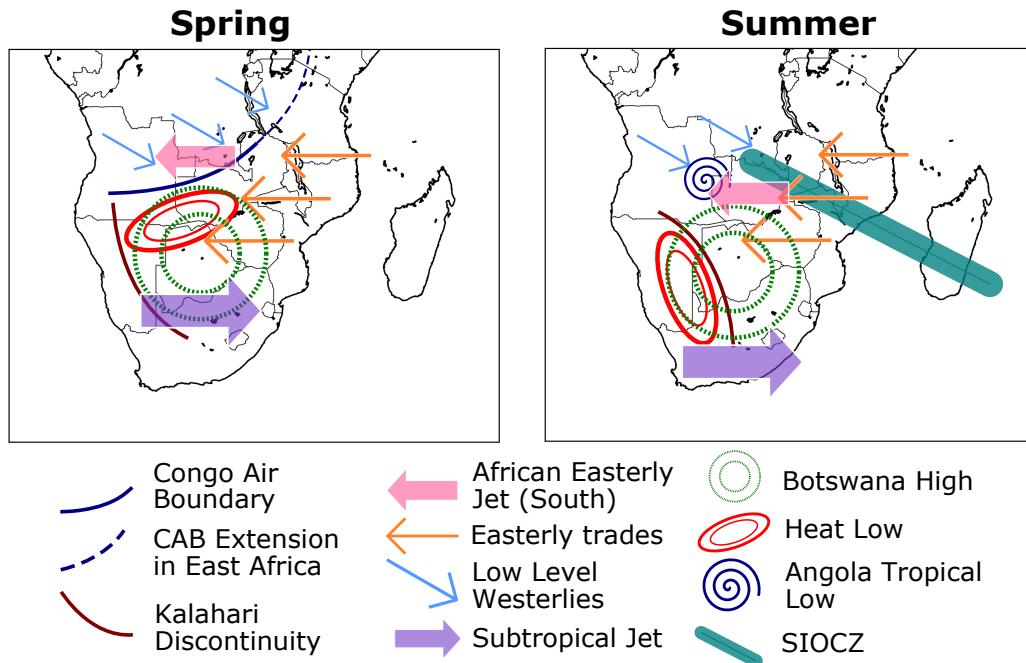


Figure 5.13. Schematic of key features relating to the CAB and KD. Left: before the CAB breaks down (typically August to mid-October), Right, after the CAB breaks down (typically November and December). Features are as given in legend, and not to scale. Locations are approximate. The extension of the CAB into East Africa has not been studied in this chapter. In the left panel, the KD is a convergence line only, and not a dryline.

The KD is consistently located on the slopes of Namibia's Central Plateau, with the humid side down-slope of the dry side. Similarly, the CAB climbs up from the Congo basin into Angola and Zambia over its seasonal cycle, and begins to disintegrating when it reaches the top of the plateau (Figures 5.4 and 5.7). In other regions of the world, notably the Great Plains of the USA (Hoch and Markowski, 2005; Jones and Bannon, 2002), drylines also progress uphill over the season and the topography gradient has been described as the leading cause of the drylines. It is possible that topography also plays a role in maintaining the KD and the CAB.

A consistent co-location of drylines and surface potential temperature maximums has been observed throughout this chapter. This suggests that the dynamics of these drylines may be closely linked with that of heat lows: the CAB with the Angola Heat Low (AHL) and the KD with the Kalahari Heat Low (KHL). The winds that converge into the convergence line CAB are, at least partially, driven by the direct thermal contrast set up by the heat low. By examining Figures 5.6 and 5.9, we found that the overnight northerly winds associated with the AHL's direct thermal inflow (Rácz and Smith, 1999) advected the CAB south overnight, while sub-grid scale turbulent mixing, maximised in the unstably stratified core of the heat low, reduced the surface humidity at and ahead of the CAB during the day. Thus, the heat low maintains the sharp CAB gradient. Conversely, however, the dry and cloudless atmosphere to the south of the CAB and to the south-west of the KD has a relatively low heat capacity and high surface solar irradiation, and so is easy to heat up, meaning that the heat low may preferentially form on the dry sides of the CAB and KD. Thus, it is clear that the heat lows and drylines are intrinsically linked, and operate together.

The definition of the CAB in this chapter was restricted to southern Africa with a near-zonal orientation. This was necessary in order to isolate the CAB from spurious features associated with lake boundaries, coasts and weather events. However, it was noted by the authors during the construction of the algorithm that the CAB drylines and convergence lines frequently extend diagonally eastwards into east Africa. This can also be seen in Figures 5.2, 5.3 and 5.11, and is consistent with early definitions of the CAB. Further analysis is necessary to examine the CAB in East Africa. Similarly, this chapter has only considered the CAB in the Austral spring and early summer, however an investigation into whether it is also present in autumn and winter would be welcomed.

Future projections of climate change in southern Africa indicate that early season drying will be accompanied by an expansion of the Angola and Kalahari heat lows (Dunning et al., 2018; Cook and Vizy, 2013). Given the close relationship between the heat lows and the CAB and the KD, this may cause a change in the behaviour of the CAB and the KD in the future. These changes may be essential for a full understanding of the projected early season drying. The next step is to determine whether the CAB and the KD are represented in future climate models. The fidelity with which these features are simulated likely holds a vital key to confidence in climate model projections.

5.10. Acknowledgments

This work has been generously supported by the Origin Foundation John Monash Scholarship and by the Future Climate for Africa UMFULA project, with financial support from the U.K. Natural Environment Research Council (NERC), NE/M020207/1, and the U.K. Government's Department for International Development (DfID). ERA-5 data was accessed from the Copernicus Climate Change Service (C3S), at <https://cds.climate.copernicus.eu/cdsapp#!/home>. SASSCAL station data was retrieved from SASSCAL WeatherNet at www.sasscalweather.net. Gridded rainfall data was retrieved from the Physical Science Division of the NOAA/ Earth System Research Laboratory (<https://www.esrl.noaa.gov/psd/data/gridded/data.gpcp.html>), and the Climate Hazards Group (http://chg.geog.ucsb.edu/data/chirps/#_Data). The authors thank Neil Hart for access to the MetBot output data.

5.11. Afterword to Chapter 5

This chapter identified and studied the CAB and the KD in ERA-5 reanalysis. After developing an algorithm to detect drylines and convergence lines, the climatologies, dynamical links with the AHL and influence on interannual rainfall variability were studied. This addressed Aims 1 to 4 of this thesis, with respect to the CAB.

Although it differs in character, this work follows from the analysis of the AHL in Chapter 3. In section 5.7, the role that the AHL plays in maintaining the diurnal progression of the CAB was examined. A similar link was shown in Figure 3.18.

The algorithm developed in this section will be modified for use at lower resolutions and then applied to CMIP5 models in Chapter 6. The results presented in this chapter will serve as a reference point for judging the accuracy with which CMIP5 resolve the CAB.

CHAPTER 6

Tracing future early summer drying in southern Africa to tropical lows and the Congo Air Boundary

In southern Africa, models from the 5th Coupled Model Intercomparison Project (CMIP5) predict robust future drying associated with a delayed rainy season onset in the Austral spring and a range of wetting and drying patterns in the Austral summer. This chapter relates these rainfall changes to dynamical shifts in two classes of weather systems: the Congo Air Boundary (CAB) and tropical lows. Objective algorithms are used to track these features in CMIP5 daily and 6-hourly model output. It is then established that the climatological locations and frequencies of these systems are reasonably well represented in the CMIP5 models, based on comparison with reanalysis. RCP8.5 end of 21st century projections are compared with historical end of 20th century simulations. Future projections in tropical low locations and frequencies diverge, but indicate an overall average decrease of 15% and in some cases a northward shift. The projected spatial change in the tropical low frequency distribution is weakly positively correlated to the projected spatial change in the Austral summer rainfall distribution. Meanwhile, future projections indicate a 13% increase in CAB frequency from October to December. This is associated with the gradual climatological CAB breakdown occurring half a month later on average in end of 21st century RCP8.5 projections. A delay in the gradual seasonal decline of the CAB prevents rainfall to the south of the CAB's mean position, most of which is shown to occur on CAB

breakdown days, hence creating the Austral spring drying signal and delayed wet season onset. Inter-model variability in the magnitude of CAB frequency increase is able to explain inter-model variability in the projected drying.

6.1. Foreword to Chapter 6

Chapter 6 solely addresses Aim 5 for both tropical lows and the Congo Air Boundary. Using the algorithms developed in Chapters 4 and 5, the representation of these two features in historical climate models is shown to be sufficient for future projections to be considered. The future change of the frequencies and spatial distributions of these weather systems is then considered. It is found that the CAB becomes more frequent and shifts north, and that tropical lows generally become less frequent and sometimes also shift north. The location of the CAB is shown to control the location of the transition between wetting or drying in OND, and the magnitude of frequency change is correlated with the magnitude of drying. Meanwhile, changes in the spatial distribution of tropical low locations correlates with mid-summer changes in the rainfall distribution along the tropical edge across models. Therefore, the projected changes in the CAB and tropical lows largely characterise future rainfall change on the tropical edge of southern Africa. This chapter has been submitted as a journal article to the Journal of Climate in October 2019.

6.2. Introduction

CMIP5 based rainfall projections exhibit a rainfall decline over southern Africa which is strongest in the October, November and December (OND) season (James and Washington, 2013). This rainfall decline has been linked with a delay in the onset date of the rainy season (Dunning et al., 2018), and in many

models comes with an increase in rainfall in central Africa in the same season (Aloysius et al., 2016; Creese et al., 2019). Southern Africa is highly vulnerable to climate related socio-economic risk, as the regional water resource supports the local food-energy-water nexus (Conway et al., 2015). In particular, rainy season onset and cessation dates are particularly important for farmers and other stakeholders (Hachigonta et al., 2008). Some studies have found consistent delays of southern African wet season onset in satellite and gauge-based observational datasets covering the late 20th and early 21st centuries (e.g. Jiang et al., 2019; Kniveton et al., 2009). Delayed rainfall onset would shorten the growing season, as no corresponding delay in cessation has been projected (Dunning et al., 2018). This will reduce the agricultural viability of the region and have follow-on effects on regional food security and economic growth (Lobell et al., 2008; Schlenker and Lobell, 2010).

At first glance, the southern African drying trend appears to follow that of the broader subtropical drying trend. However, closer analysis reveals that in southern Africa, the drying occurs primarily north of the subtropical rainfall minimum (Scheff and Frierson, 2012*a,b*). Furthermore, while the subtropical drying signal is largely characterised by an increase in the magnitude of the difference between precipitation and evaporation (P-E), in southern Africa evaporation decreases in line with precipitation, so that P-E does not change considerably. Lazenby et al. (2018) demonstrated that the southern African drying trend is primarily a consequence of circulation changes, rather than thermodynamic mechanisms such as dry-get-drier (Held and Soden, 2006) or upped-ante (Chou et al., 2009). Thus, the southern African drying trend is dynamically unique and requires a bespoke approach.

Some work has been done on understanding the dynamics of projected rainfall change in southern Africa. Lazenby et al. (2018) found that this change could be viewed as a northward shift of the African rain-band, while Dunning et al. (2018) linked the delayed onset of OND rains in southern Africa to relative changes in the strengths of the Saharan and Angolan heat lows. Munday and Washington (2019) found a complementary result: models with a deeper future climatological Angola heat low showed a higher intensity of drying. Similarly, Cook and Vizzy (2013) observed a strengthening of the Angola low in regional climate model simulations. However, little work has been done to directly link rainfall change to changes in precipitating weather systems.

Meanwhile, less work has looked at rainfall change in December - February (DJF), the peak of the wet season (Van Heerden and Taljaard, 1998). Model projections disagree in this season, although Lazenby et al. (2018) suggest that inter-model differences in SST trends in adjacent oceans may explain the diverse range of model projections. Once again, the characteristics of the future change patterns of atmospheric features on synoptic timescales has not been studied.

Since atmospheric circulation on climatological timescales is the aggregation of synoptic weather systems, future circulation change hinges on the response of synoptic weather systems to a warmer atmosphere. However, the synoptic weather systems that occur in the southern African tropical margins are understudied. The representation of such synoptic weather systems in CMIP5 models and their future change has not been studied. This knowledge gap ensures that the synoptic context of the future change signal in this region, and whether or not a specific class of weather system may deliver the projected change, is unknown.

This chapter focusses on two local classes of weather systems which occur in the latitudinal band of the advancing spring rains and have previously been associated with rainfall variability. The first is the Congo Air Boundary (CAB), a combined surface dryline and convergence line that marks the boundary between tropical and subtropical weather zones in the Austral spring (Chapter 5). The second class of weather systems considered are tropical lows, cyclonic rotating vortices that form in the Austral summer and tend to cluster near 20°S, creating the climatological Angola tropical low (Chapter 3). By examining the representation of the CAB and tropical lows in historical coupled climate models, this work follows a process based model evaluation framework. After establishing whether these systems are well modelled, we then proceed to studying future projections of these synoptic processes. This involves investigating how the CAB and tropical lows are projected to change according to RCP8.5 end of 21st century simulations.

Both of these weather systems were identified by Taljaard (1986) as two of the ten most important factors that influence the weather over southern Africa. Crucially, of these 10 factors, the CAB and tropical lows are located in the region of maximum rainfall change and both operate on length-scales that are large enough to be resolved in CMIP5 models. The focus of this work is on the Austral spring and summer seasons, since the former is the season with the strongest rainfall change, and the latter is the main rainy season. Our analysis will show that changes in tropical lows and the CAB are vitally important for future projections of southern African precipitation, as they directly influence the delivery of model precipitation. However, result this does not preclude changes in other factors, including the distribution of sea surface temperatures, the subtropical jet and high-pressure belt, and upper-level waves, from also playing a role. Indeed,

it is possible that the circulation changes of tropical lows and the CAB are linked to changes in these other systems.

The CAB has historically been defined as the confluence zone between the Congo Airmass, a moist and convectively active region of air that sits over the Congo rainforest, and the drier trade easterlies that cross southern Africa and originate from the Indian Ocean (Taljaard, 1972). The CAB was originally discussed in the context of identifying the elusive ITCZ over southern and eastern Africa (Taljaard, 1953). It was argued that the CAB could not be the ITCZ itself, as both airmasses involved originate from the Southern Hemisphere (the Congo Airmass being associated with the recurvature of the south Atlantic trade winds, known locally as the low level westerlies, e.g., Leroux (2001)). However, the importance of the CAB for southern African rainfall was undisputed (Torrance, 1979).

Despite its importance to the local climate, no systematic study of the representation of the CAB in reanalysis or model products had been performed before Chapter 5. They optimised an edge-detecting algorithm and a ridge-detecting algorithm to pick out sharp gradients in specific humidity and ridges in wind convergence that were associated with the southern portion of the CAB. They distinguished between the ‘dryline’ CAB and a ‘convergence line’ CAB based on the choice of algorithm and determined that although the two were closely comparable, the ‘dryline’ algorithm was slightly more reliable. Chapter 5 also identified the Kalahari Discontinuity (KD), a similar near-surface dryline/convergence line system located further south and oriented parallel to the west coast of southern Africa, that forms after the CAB breaks down in October and November. They confirmed that the CAB latitude and detection frequency were closely linked to the interannual variability of spring rainfall over southern Africa. This suggests

that the future change of the CAB has potential explanatory power for the OND drying.

Southern African tropical lows are synoptic-scale cyclonic vortices with depths up to 500 hPa that track predominantly westward across the southern African continent (Chapter 3). They are precipitating systems, and cluster in eastern Angola and western Zambia, where they tend to become semi-stationary (Chapter 4). In southern Africa, tropical lows have been most commonly studied in the form of their climatological mean, the Angola tropical low (e.g. Cook et al., 2004; Cr  tat et al., 2018; Pascale et al., 2019). Since the Angola tropical low has been closely linked to interannual rainfall variability (Reason and Jagadheesha, 2005; Cook et al., 2004), considerable attention has been paid to how it is simulated in various models. Lazenby et al. (2016) and Munday and Washington (2017) both found that the Angola low is excessively strong in CMIP5 simulations of historical climate, and linked this to the wet bias over southern Africa. However, the representation of the Angola tropical low on synoptic timescales in CMIP5 models, and its future change, has not yet been assessed.

The goal of this chapter is to express projected rainfall changes from the viewpoint of projected changes in synoptic weather systems. This involves first linking the spring drying signal to projected changes in characteristics of the CAB, and also linking the summer inter-model spread of rainfall projections with the inter-model spread of changes to the spatial distribution and frequencies of tropical lows and the CAB. To achieve this aim, the chapter proceeds as follows. In section 2, the models, datasets and feature identification algorithms are described. In section 3, we consider the historical representation of the CAB, the CAB’s projected future change and the rainfall implications of that projected change. In section 4, we perform a similar analysis on tropical lows. In the final section,

we summarise the importance of this work for understanding rainfall change in southern Africa.

6.3. Methods

6.3.1. Models and reanalysis datasets

A selection of 25 CMIP5 models has been used in this study, based on the availability of the appropriate model output variables. To study the CAB, we require daily surface level specific humidity and temperature data, which 18 of these models had available. To study tropical lows, we required 6 hourly wind data on pressure levels, which was available in a different subset of 18 models. The model names, creators and grid-spacings are shown in table 6.1, together with a list of which models were used to study the CAB and tropical lows. For each model, 30 years of end of 20th century coupled climate model output simulating the period 1970-2000 were examined, and compared to 70 years of end of 21st century model output generated under the RCP8.5 coupled model scenario representing the period 2070-2100. The simulated historical climatology of each feature was compared to the reanalysis climatologies from three reanalysis products: ERA-5, ERA-Interim and MERRA-2. Reanalysis climatologies were taken from the 30 year period 1980-2010.

The CAB has been identified between August and December, and its correlation with rainfall change has been studied between October and December. The former season has been chosen because this is the season in which the CAB is present in southern Africa, and the season in which validation against the results of Chapter 5 was possible. The latter season was chosen because it is the season of maximum rainfall decline in southern Africa (Munday and Washington, 2019). Tropical lows were identified between November and March, and their influence

on future rainfall change was studied between December and February. Again, the former is the season in which tropical lows are present in southern Africa (Chapter 4). The latter is the main wet season in southern Africa, and is also the season in which the contribution of tropical lows towards southern African rainfall is most significant (Chapter 4). The OND precipitation decline across the two 30 year time periods is considerably larger than the 30-year decadal standard deviation as calculated from the corresponding pre-industrial control experiments in over 50% of the models considered, as indicated by Supplementary Figure 6.15. This implies that the drying signal cannot be attributed to the internal decadal variability of the climate models. This was not the case in DJF.

6.3.2. Congo Air Boundary

This chapter adapts the methodology of Chapter 5 to identify the CAB in CMIP5 models. Because they studied the CAB in a high resolution ($\sim 0.25^\circ$) reanalysis dataset, and the models employed here have resolutions ranging from $\sim 1^\circ - 3^\circ$, this methodology needs a few modifications in order to transfer over to lower resolutions. Properties of the CAB are dependent on the resolution of the input data, and so all models and reanalyses are regridded to a $2^\circ \times 2^\circ$ grid.

While in Chapter 5 both wind convergence and humidity were used to detect the CAB, the present study only uses humidity, and focusses solely on what was referred to in Chapter 5 as the ‘dryline CAB’. This choice was made because in Chapter 5 it was found that the dryline CAB was more reliable and easily detected than the convergence line CAB. Near-surface relative humidity gradients, rather than specific humidity gradients, have been used to calculate the CAB location. This choice has been made to allow for easier comparison between historical and RCP8.5 experiments, given that near-surface specific humidity generally increases

Table 6.1. List of CMIP5 models and reanalysis products considered in this study, together with horizontal grid spacings, and an indication of whether the model was used to study the CAB, tropical lows (TL) or both.

Model Name	Institute	Horizontal Spacing	CAB	TL
bcc-csm1-1-m	Beijing Climate Center, China Meteorological Administration College of Global Change and	$1.125^\circ \times 1.125^\circ$	1	1
BNU-ESM	Earth System Science, Beijing Normal University	$2.8^\circ \times 2.8^\circ$	1	1
CanESM2	Canadian Centre for Climate Modelling and Analysis	$2.8^\circ \times 2.8^\circ$	1	1
CNRM-CM5	Centre National de Recherches Météorologiques	$1.4^\circ \times 1.4^\circ$	1	1
ACCESS1-0	Commonwealth Scientific and Industrial Research Organisation and Bureau of Meteorology, Australia	$1.25^\circ \times 1.825^\circ$	1	1
ACCESS1-3	Commonwealth Scientific and Industrial Research Organisation and Bureau of Meteorology, Australia	$1.25^\circ \times 1.825^\circ$	1	1
CSIRO-Mk3-6-0	Commonwealth Scientific and Industrial Research Organisation in collaboration with Queensland Climate Change Centre of Excellence	$1.825^\circ \times 1.825^\circ$	1	
inmcm4	Institute for Numerical Mathematics, Moscow, Russia	$1.5^\circ \times 2.0^\circ$	1	1
IPSL-CM5A-LR	L’Institut Pierre-Simon Laplace	$1.9^\circ \times 3.75^\circ$	1	1
IPSL-CM5A-MR	L’Institut Pierre-Simon Laplace	$1.25^\circ \times 2.5^\circ$	1	1
IPSL-CM5B-LR	L’Institut Pierre-Simon Laplace	$1.9^\circ \times 3.75^\circ$	1	1
HadGEM2-CC	UK Met Office Hadley Centre	$1.25^\circ \times 1.9^\circ$	1	1
HadGEM2-ES	UK Met Office Hadley Centre	$1.25^\circ \times 1.9^\circ$	1	1
MPI-ESM-LR	Max Planck Institute for Meteorology M24	$1.8^\circ \times 1.8^\circ$		1
MPI-ESM-MR	Max Planck Institute for Meteorology M24	$1.8^\circ \times 1.8^\circ$		1
NorESM1-M	Norwegian Climate Centre	$1.875^\circ \times 2.5^\circ$	1	1
GFDL-CM3	NOAA/Geophysical Fluid Dynamics Laboratory	$2.0^\circ \times 2.5^\circ$	1	1
GFDL-ESM2G	NOAA/Geophysical Fluid Dynamics Laboratory	$2.0^\circ \times 2.5^\circ$	1	1
GFDL-ESM2M	NOAA/Geophysical Fluid Dynamics Laboratory	$2.0^\circ \times 2.5^\circ$	1	
CESM1-CAM5	Community Earth System Model contributor	$0.94^\circ \times 1.25^\circ$	1	
Reanalysis Name	Institute	Horizontal Spacing	CAB	TL
ERA-5	European Centre for Medium Range Forecasting	$0.25^\circ \times 0.25^\circ$	1	1
ERA-Interim	European Centre for Medium Range Forecasting	$0.75^\circ \times 0.75^\circ$	1	1
MERRA-2	NASA	$0.5^\circ \times 0.75^\circ$	1	1

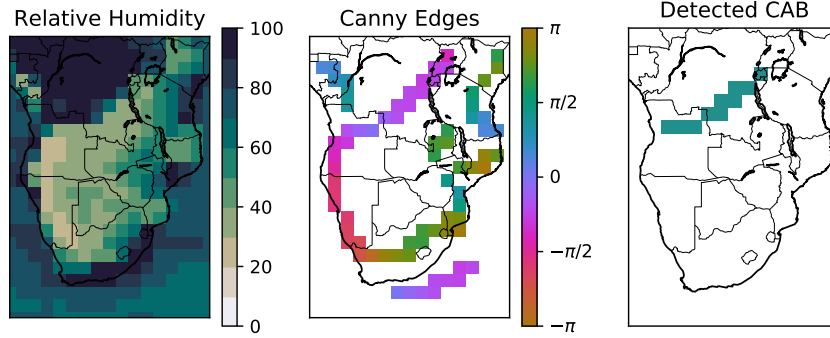


Figure 6.1. Surface relative humidity (left panel), Canny edges (middle panel) and identified CAB locations (right panel) on the 9th of September 1999 in ACCESS1.3.

by a factor of about two across the tropics by the end of the 21st century in the RCP8.5 scenario.

The algorithm used to detect the CAB is as follows. Figure 6.1 shows the algorithm applied to a sample day (9/9/1999) from a sample model (ACCESS1.3).

- (1) The 2m relative humidity is calculated from 2m daily mean air temperature (*tas*), 2m daily mean specific humidity (*huss*) and surface pressure (*sp*), where these fields are available. Where they are not available, the 1000 hPa pressure level relative humidity (*hur*) is used instead.
- (2) Relative humidity is interpolated to a $2^\circ \times 2^\circ$ grid using a nearest neighbour interpolation scheme. This scheme is chosen in order to avoid differentially smoothing humidity gradients in the higher resolution datasets during the regridding process. The left panel of Figure 6.1 shows the regridded field.
- (3) In keeping with the Canny (Canny, 1986) algorithm, a Gaussian filter with a 2° radius is used to smooth the relative humidity field. The

magnitude (M) and direction (θ) of the gradient were then calculated using finite differences.

- (4) Canny edges were calculated as per Chapter 5, thresholding the magnitude of the humidity gradient such that it must undergo an absolute change of $\Delta RH = 40\%$ between grid cells across a Canny edge in order for a grid cell to qualify as a CAB. centre panel of Figure 6.1 shows the Canny edges identified on the sample model day, coloured by the orientation angle of the edge.
- (5) Canny edges were filtered to retain instances with $-\frac{\pi}{4} < \theta < \frac{\pi}{6}$ and restricted to latitudes between 5° - 18°S .
- (6) At least 3 qualifying dryline grid cells were required to be detected at the same time in order for a CAB to be registered. The resultant CAB grid-cells shown in the right panel of Figure 6.1 on the sample model day.

The KD is also extracted from the calculated set of Canny edges. It is restricted to latitudes below 12°S and angles between $\frac{\pi}{6} < \theta < \frac{\pi}{2}$, consistent with Chapter 5. No minimum grid cell thresholds are applied in this case. As described above, three reanalysis products, ERA-5, ERA-Interim and MERRA2, have been included in this study. The thresholds described above were manually optimised so that the seasonal cycle of the relative humidity based CAB in the coarsened ERA-5 using the methodology of this chapter was qualitatively similar to the high resolution specific humidity ERA-5 results of chapter 5.

Figure 6.2 shows the near-surface relative humidity and identified CAB points for a case study day in early September in historical models and reanalyses. The atmospheric states on this chosen day will be in different synoptic setups.

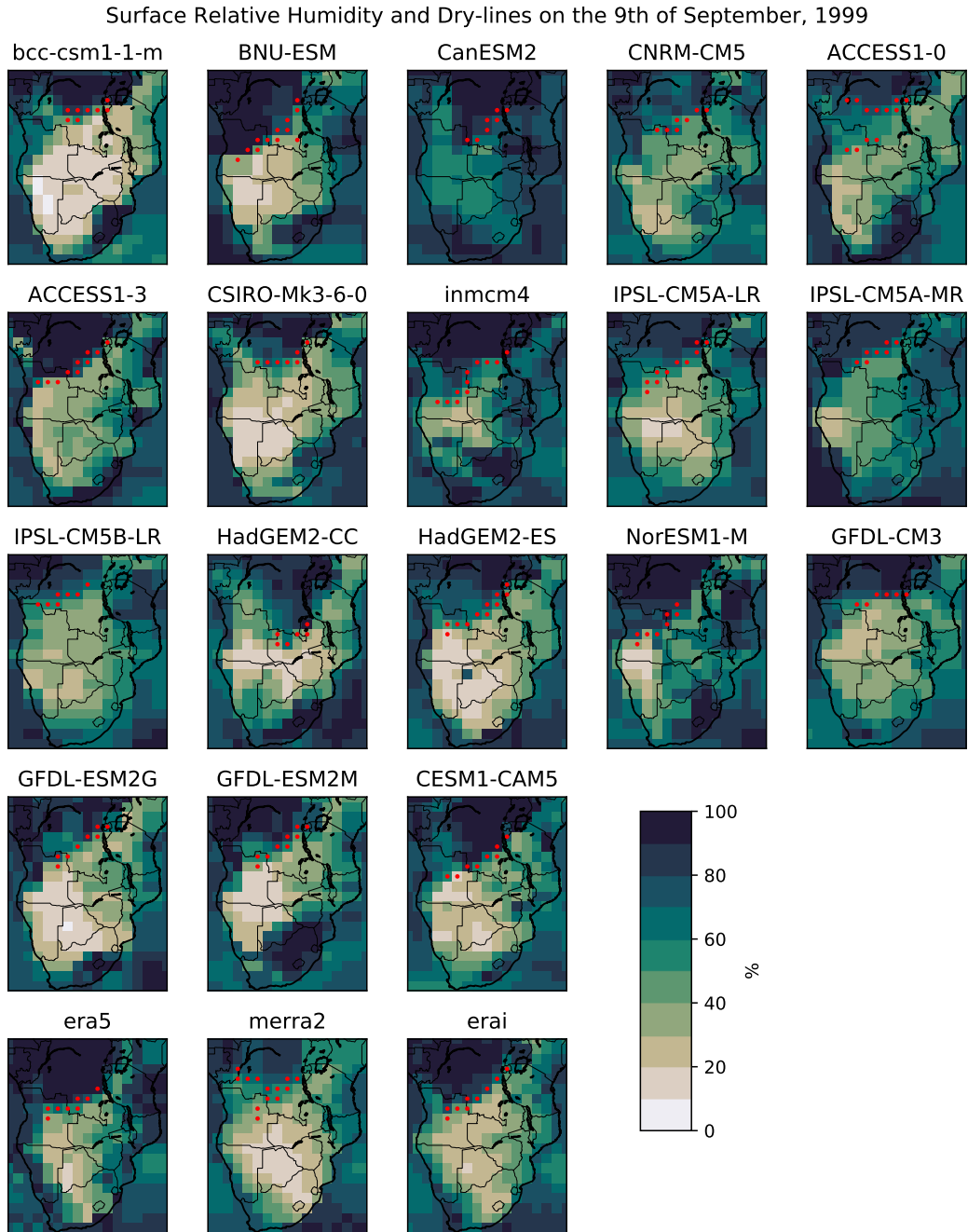


Figure 6.2. Surface relative humidity (colours) and identified CAB locations (red dots) on the 9th of September 1999 in each model and reanalysis product. Surface humidity has been regridded to a $2^\circ \times 2^\circ$ grid using the nearest neighbour method.

Nevertheless, a well-defined CAB is identified on this day in all of the CMIP5 models.

With the daily CAB positions calculated, climatologies of properties such as latitude, frequencies and extent can be computed and compared. Here, the extent is calculated by counting the total number of CAB grid cells on a given day. This gives an approximate measure of the lateral extent of the CAB.

6.3.3. Tropical lows

The methodology used to identify tropical low events follows Chapter 4. We apply the TRACK algorithm Hodges (1994, 1999). The application of TRACK to CMIP5 model experiments has previously been documented by Rastogi et al. (2018) and Bengtsson et al. (2007). Chapter 4 identified southern African tropical lows using 6-hourly vertical mean vorticity averaged across pressure levels at 600, 700 and 800 hPa. However, 6-hourly CMIP5 pressure level model data was only available at 850 and 500 hPa, and so these model levels have been used instead. A comparison between the results of using this set-up and using 600, 700 and 800 hPa daily vorticity was considered for a subset of 9 models and no significant difference was found (not shown).

The data preparation algorithm was as follows. At each vertical level and for each 6-hourly time-step, vertical vorticity was calculated from the zonal and meridional wind components at a T63 resolution, using the python package windspharm. The vertical average was taken, and then a Sardeshmukh and Hoskins (1988) filter was applied to smooth the spectral cut-off. Cyclonic vorticity extrema with $\zeta < -5 \times 10^{-6} s^{-1}$ were then identified and linked using the TRACK algorithm, as detailed in Chapter 4 and Hodges et al. (2017). Once these vortex

tracks were identified, they were filtered for southern African tropical lows using the following criteria:

- (1) Tracks must spend at least one time step over land;
- (2) Track longevity must be at least one day;
- (3) The filtered vertical mean relative vorticity must satisfy $\zeta < -3 \times 10^{-5} s^{-1}$ in at least one 6-hourly time-step;
- (4) There must be coincident cyclonic vorticity at 500, 850 hPa for a continuous 24 hour period;
- (5) The genesis location must not be in the Atlantic; and
- (6) The genesis location must be north of 25° S.

Justification for these criteria are given in Chapter 4: briefly, they exclude extra-tropical cyclones, coastally trapped Kelvin waves, heat lows, and spurious weak events. A sample track longitude Hovmöller plot for each of the models and the reanalysis over one historical year is shown in Figure 6.3. The models reveal a mixture of track behaviours.

Following the identification of tropical lows, rainfall is attributed to tropical lows by making the assumption that all daily rainfall that fell within a 5° radius of a tropical low centroid was attributable to that tropical low. This radius was shown to be appropriate for southern African tropical lows in Chapter 4 and is consistent with many previous studies of tropical lows and tropical cyclones (Baray et al., 2003; Dare et al., 2012; Khouakhi et al., 2017; Lavender and Abbs, 2013).

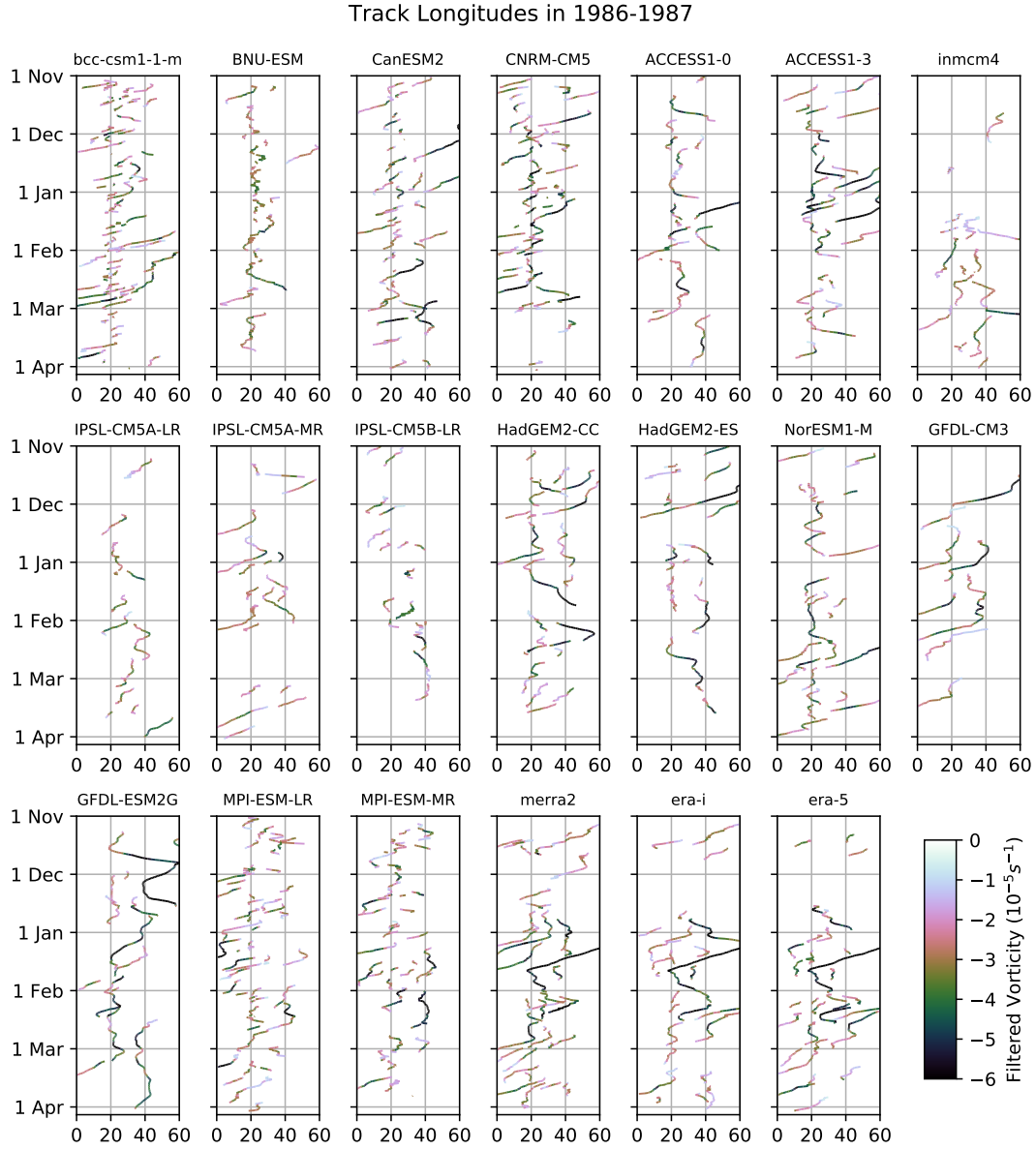


Figure 6.3. Tropical low track Hovmöller plots for each model and reanalysis product. x -axis: longitude, y -axis: time. Colours indicate the T63 filtered vorticity of the tropical low event, a measure of the event intensity.

6.4. The Congo Air Boundary

6.4.1. Representation in climate model historical simulations

The aim of this section is to determine whether or not the CAB is well represented in the CMIP5 historical simulations under consideration for the period from 1970-2000. Figures are designed to be comparable with the ERA-5 based study of the CAB presented in Chapter 5, and modelled CAB properties are compared with the coarsened ERA-5 reanalysis. The KD is also briefly considered.

Figure 6.4 indicates the climatological location of the CAB and KD in each climate model and reanalysis product, based on the frequency the CAB is detected at each interpolated $2^\circ \times 2^\circ$ grid-box. It is evident that the CAB is detected similar frequencies and locations in most of the models as in the reanalysis products. There is a range of variation across the models, however. CMRM-CM5, GFDL-CM3 and the ACCESS models show concentrated, high intensity CAB locations with a diagonal orientation from west of Lake Victoria through to south west Angola. NorESM1-M and bcc-csm1-1m show heat-maps with a distinct north and south peak. All three IPSL models show low CAB detection frequencies in the east. This is particularly interesting since the IPSL-CM5A models have previously been shown to be outliers in East Africa, lacking the moisture-rainfall relationship present in most other models (Rowell and Chadwick, 2018). CanESM2 exhibits lower frequency CAB detection rates than the other models, suggesting that it may struggle to represent the feature. The biases of the spatial distribution of the CAB in these climate models relative to era-5 reanalysis, and their significance relative to natural variability over the 30 year time period, are shown in Supplementary Figure 6.16.

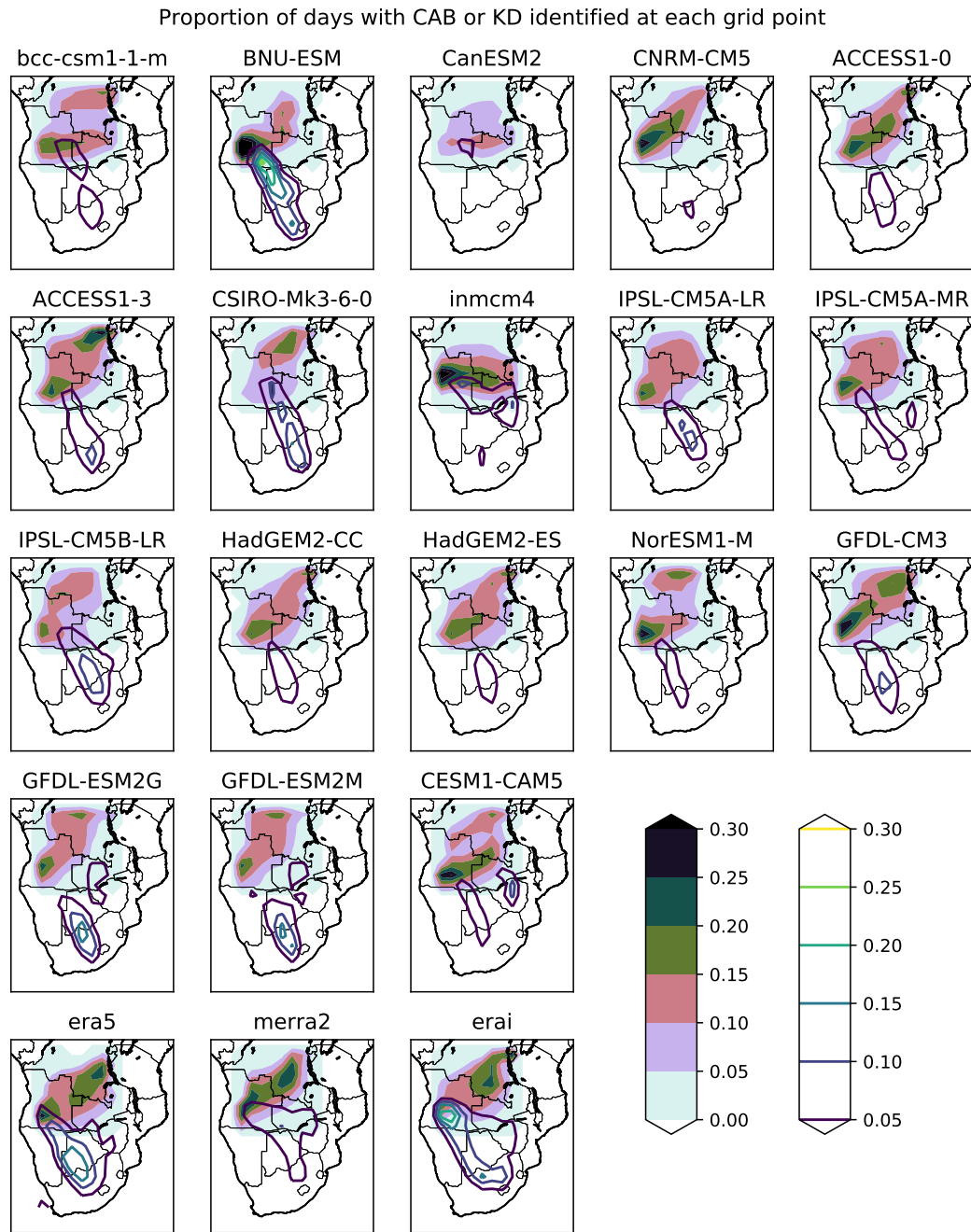


Figure 6.4. Spatial distributions of CAB and KD frequency (number of events per interpolated grid cell per year). Filled contours: CAB. Line contours: KD. Panels indicate different models and reanalysis products.

There is much more variability in the representation of the KD in CMIP5 models. BNU-ESM shows a very persistent KD, while CSIRO-Mk3.6 and GFDL models represent it reasonably well and some other models, such as CNRM-CM5 and CanESM2 miss it completely. We conclude that the KD is not as well represented in CMIP5 models as the CAB, and postpone further analysis for a future study.

We next study the seasonal cycle of key CAB features in the CMIP5 models. The seasonal cycles of the CAB latitude, frequency and extent in CMIP5 models and low resolution reanalysis products are shown in Figure 6.5. The CAB latitude and extent are both calendar day climatological means between 1970 and 2000, and all variables have been smoothed by a 2-week running mean. The extent is calculated by counting the total number of CAB grid cells on a given climatological day of the year. Only the days when a CAB was identified were used in the average. This gives an approximate measure of the lateral extent of the CAB, though it is not precisely comparable across CABs of different orientations. The latitude shown is the mean latitude of all CAB points on a given day, and the frequency is the proportion of days at a given time of year where a CAB is present.

Chapter 5 found that based on ERA-5 reanalysis, the CAB moves steadily southward between the start of August and the end of November, and that its detection frequency drops from near 100% at the start of October through to 10-20% at the beginning of December.

In the present study, similar behaviour is present in the latitude and frequency (the top and centre panels respectively of Figure 6.5). The rate at which the CAB moves south is close to constant across models and consistent with ERA-5, at roughly 2 degrees per month. Results from ERA-5 are located at the

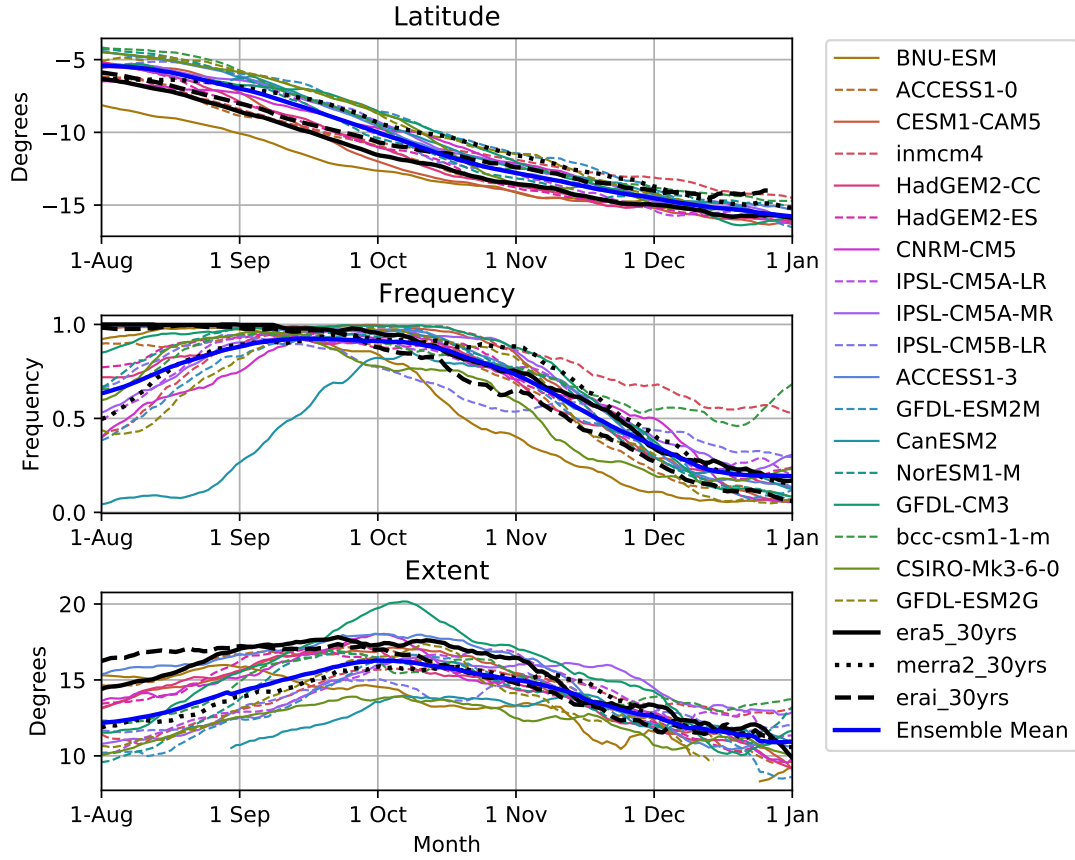


Figure 6.5. Seasonal cycles of CAB properties. Top: mean CAB latitude, second row: CAB frequency, and third row: CAB extent (number of grid cells identified per day). Thin coloured lines indicate models, and are ordered by the mean CAB latitude averaged from August to November. Black lines show re-analysis products, thick blue lines show the ensemble mean. All quantities are smoothed by a 2-week running mean.

southern edge of the CMIP5 simulated range. At any given time of the year, the climatological CAB latitude has a range of 5° . Since this range is of a similar order of magnitude to the Nyquist frequency ($2\Delta x = 4^\circ$), we surmise that the CAB latitude is represented in these models as well as can be expected. Based on the centre panel of Figure 6.5, it is evident most models show a decreased CAB

frequency in August, however all models except CAN-ESM2 recover by early September and the vast majority possess CABs that are present 80% of the time. The CAB breaks down between October and December in all models, with ERA-5 breaking down later than the ensemble mean but being located well within the model range. The ensemble mean CAB extent peaks at the beginning of October, consistent with the discontinuity width presented in Chapter 5 (Figure 5.5).

Based on the above analysis we conclude that the CAB is well represented in most of the climate models considered, with the exception being CanESM2. We therefore proceed to study the change in the CAB between present day and future climate model simulations, and to assess the impact of this change on rainfall in southern Africa under the RCP8.5 scenario.

6.4.2. Future change and rainfall implications

In this section, we explore how the CAB changes in future simulations and whether these changes are linked to projected southern African drying. Figure 6.6 shows the difference, in each model, between the CAB latitude, frequency and extent in the RCP8.5 end of 21st century simulation and in the historical end of 20th century simulation. The main change is an increase in the CAB frequency from October to December, peaking at approximately 25% in GFDL-CM3 and averaging to 13% in the ensemble mean. This frequency increase is accompanied by a northward shift in CAB latitude on average of 0.7°. These changes are significant relative to natural variability in between 11 and 15 of the 18 models across the three months, as shown in the bottom panel of Figure 6.6. This indicates that the gradual southward progression and seasonal frequency decline of the CAB is delayed in most climate models, and so the CAB becomes more frequent towards the end of the season. Further analysis (not shown) indicates

that the delay in CAB breakdown is approximately half a month. During this period, there is an increase in the CAB extent, which is significant different from 0 as compared to model spread at the $p < 0.05$ level in November and December, and compared to natural variability in 12 models.

CAB frequency is also decreased in mid-August in most models and the ensemble mean. This is the same period when the CAB frequency was not well represented in Figure 6.5. This frequency decline may be related to the anomalous representation of the CAB in August. Most models return to displaying historical CAB frequencies by September, with the exception of CanESM2, whose CAB was shown earlier to be poorly represented in historical simulations.

Chapter 5 found that the interannual CAB frequency was anti-correlated with precipitation between 10° and 15°S in October, November and December. Since the October-December CAB frequency increases in most CMIP5 models considered in this study, it seems plausible that this increase may be linked to the projected rainfall decline present in most models. In order to explore this further, we decompose daily OND rainfall at each grid-point into three components:

- (1) North of CAB rain: rain that fell at a grid-point that was in the same longitude band as an identified CAB dryline, with the grid-point located to the north of the dryline;
- (2) South of CAB rain: rain that fell at a grid-point that was in the same longitude band as an identified CAB dryline, with the grid-point located to the south of the dryline; and
- (3) CAB breakdown rain: rain that fell at a grid-cell for which no CAB drylines were detected in the same longitude band.

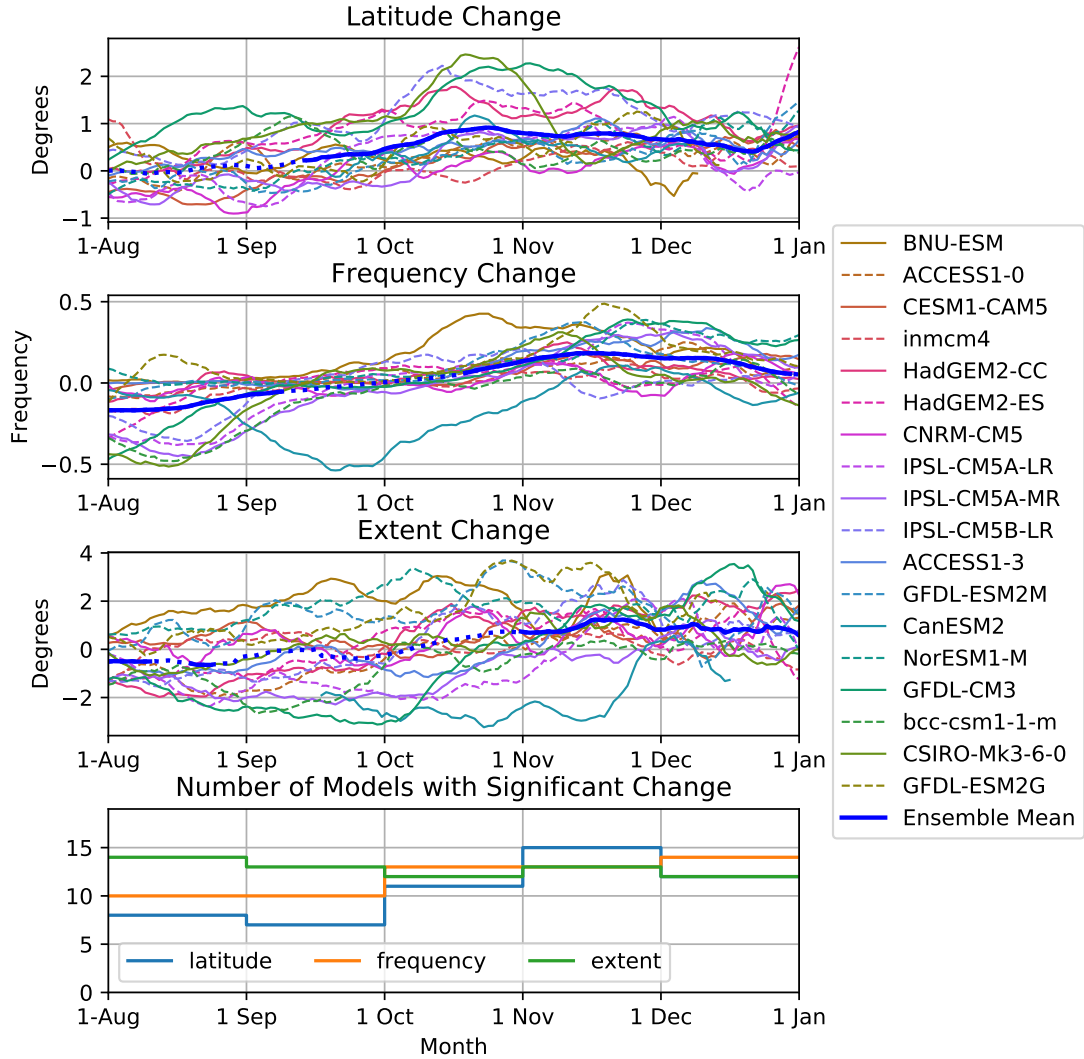


Figure 6.6. Seasonal cycles of future change of CAB properties. As per Figure 6.5, but showing the average of each property for the RCP8.5 end of 21st century scenario, minus that for the historical end of 20th century scenario. The thick blue line indicates the ensemble mean and is shown as a solid line when the ensemble mean is significantly different from zero at the $p < 0.05$ level using a paired t-test, and a dotted line otherwise. All quantities are smoothed by a 2-week running mean. The bottom panel indicates the number of models for which the future change signal is significance against internal variability at a $p < 0.05$ level for each month, using a Welch's t-test. Colours are ordered as per Figure 6.5. Colours in the top 3 panels are ordered as per Figure 6.5.

The top row of Figure 6.7 shows the decomposition of the historical climatological mean OND rainfall into these three categories, as a function of latitude and averaged over longitudes between 20° and 30°S . As predicted, the rainfall to the south of the CAB is small, and is only comparable to the total rainfall at subtropical latitudes (30° - 40°S). Rainfall at tropical latitudes, between 0° and 15°S , is evenly distributed between the remaining two categories. North of CAB rainfall is identically zero south of 18°S , as the CAB does not extend south of this point.

The division of rainfall into these three components is based on the hypothesis, proposed in Chapter 5, that the CAB largely prevents tropical rain to its south and that the primary means by which spring rainfall occurs in southern Africa is a full or partial breach in the CAB. Thus component (2) - rainfall south of the CAB - is expected to be small and primarily of extra-tropical origin. The other components represent: (1) rainfall associated with ‘Congo Air’ in the deep tropics, and (3) rainfall associated with tropical temperate troughs (TTTs) and other CAB breakdown events. This decomposition does not account for the fact that a grid cell associated with an elongated TTT may exist to the south of a CAB grid cell due to the TTT’s diagonal structure. However, this limitation does not appear to be significant, based on the calculated low magnitude of rain to the south of the CAB in Figure 6.7.

Meanwhile, the middle panel of Figure 6.7 shows the future change of each category of rainfall. The change in the total rainfall shows the familiar dipole structure, with most models showing drying south of 10°S and either wetting or a comparatively low magnitude of drying north of 5°S . The decomposition into rainfall classifications is enlightening: there is an increase in rainfall to the north of the CAB, and a decrease in rainfall coming from the CAB breakdown events.

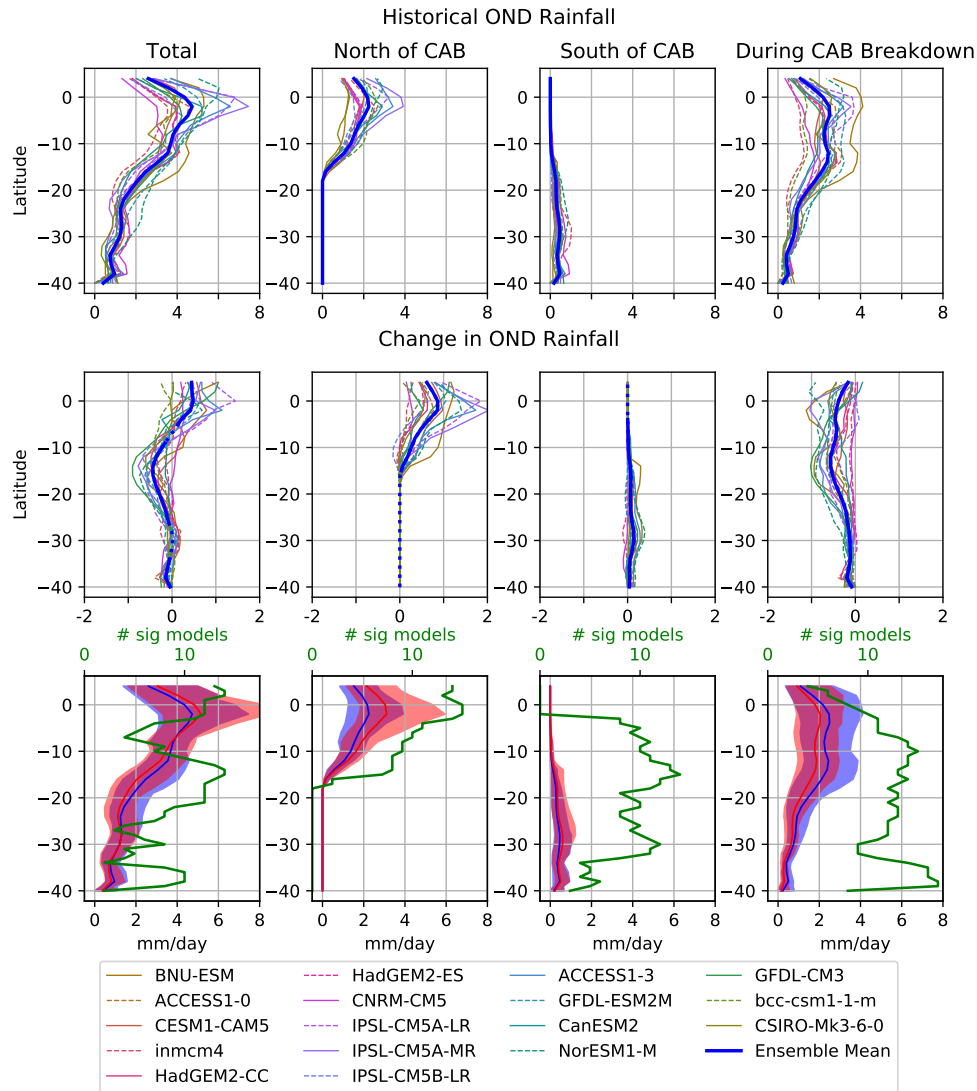


Figure 6.7. CAB rainfall decomposition. Top row: historical rainfall, middle row: RCP8.5 minus historical. The bottom row compares the inter-model spread in historical (blue) and RCP8.5 (red) simulations, and shows the number of models which exhibit a significant change relative to interannual variability at the $p < 0.05$ level, based on a paired t-test (green line, top axis labels). First column: total rain. Following columns show rainfall that falls: north of the CAB (second column), south of the CAB (third column) and during CAB break down (last column). All panel show October - December mean. Colours are ordered as per Figure 6.5.

In each case the direction of change is remarkably robust between models and across latitudes.

To provide further visualisation of the projected change, the lower panel of Figure 6.7 shows the envelopes indicating the model spread of rainfall projections in historical (blue) and RCP8.5 (red) simulations. Change in the rainfall decomposition terms is more pronounced than change in the total rainfall. The green line in these figures, indicates the number of models which show a significant change relative to natural variability. At least 13 of the 18 models show a significant decline in CAB breakdown rainfall between 8° and 28° S.

This implies that the ensemble mean OND drying and rainfall change dipole is associated with the change in the CAB frequency. The drying is fully contained within the component of the rainfall that falls on CAB breakdown days, while the wetting occurs to the north of the CAB. The rainfall rate per CAB breakdown day and or north of CAB rainfall per CAB day was also considered, but no consensus on the sign of change was apparent (not shown). The projected decrease in rainfall on non-CAB days and increase in rainfall to the north of the CAB is therefore directly linked to the projected increase in the frequency of CAB days. The spatial patterns of the ensemble mean change of OND rainfall under this decomposition, shown in the top row of Figure 6.8, are consistent with this conclusion.

Furthermore, inter-model spread of the projected CAB frequency increase explains a large proportion of the projected southern African drying. This is shown in the lower panels of Figure 6.8, which show the inter-model regression of OND model rainfall change (averaged over 15° - 30° E and 5° - 25° S) against the modelled CAB frequency change. CAB frequency change is averaged over November and December, the months in which the ensemble mean change is

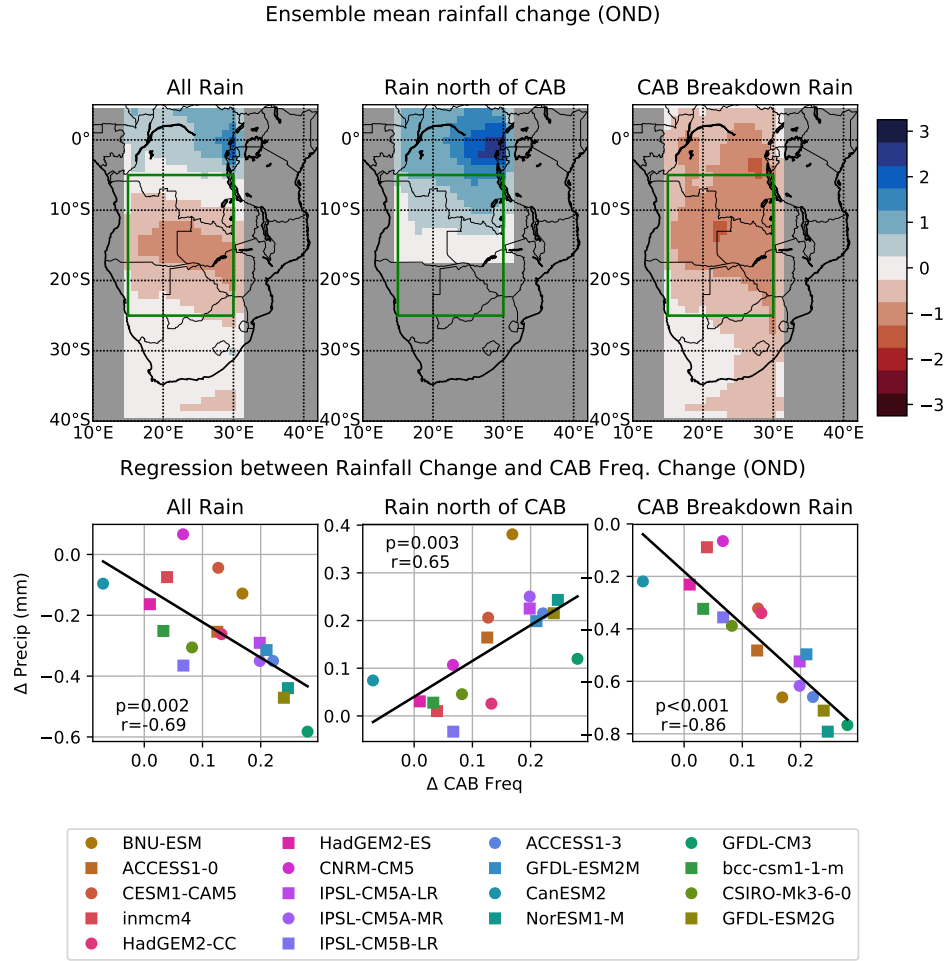


Figure 6.8. Top row: Ensemble mean rainfall change OND based on CAB decomposition. Bottom row: Linear regression between rainfall OND change in the region 15° - 30° E, 5° - 25° S and the November-December CAB frequency change. Black line: least squares regression, text: p-value for the test that the slope of the regression is equal to zero, and Pearson's correlation coefficient. Left column: total rainfall, centre column: rain that occurs north of the CAB, right column: rain that occurs during CAB breakdown events. Colours are ordered as per Figure 6.5.

significant. Before the CAB rainfall decomposition is applied, 46% of the inter-model variation of total OND rainfall change is explained by inter-model variation in the CAB frequency change. Variation in the rainfall decline during CAB

breakdown events is more strongly predicted by variation in the CAB latitude change ($R^2=0.71$). While the CanESM2 data point (blue-green circle) appears to have a large degree of leverage in these regressions, its exclusion did not strongly impact the significance of the results (not shown).

Taken together, these results imply that the OND rainfall decline signal in southern Africa is largely explained by the increased frequency of the CAB, which prevents CAB breakdown associated rainfall in the regions that are located to the south of the CAB. Between 5° and 15°S , this comes with an increase in rainfall on CAB days, as parts of this region are often located to the north of the CAB.

6.5. Tropical lows

We now shift focus to tropical lows, cyclonic vortices that form in the Austral summer and have been found to deliver 31% of summer rainfall to the tropical edge region (16° - 22°S) of southern Africa (Chapter 4). Tropical lows tend to cluster in Angola and western Zambia, where they form the synoptic expression of the late-summer tropical low phase of the climatological Angola low (Chapter 3). This section first examines the representation of tropical lows in 18 CMIP5 models. We then consider their contribution to precipitation and future change.

6.5.1. Representation in historical climate models

In order to evaluate the representation of tropical lows in CMIP5 models, we first consider climatological spatial distributions of tropical low locations, shown in Figure 6.9. From this figure, it is evident that most models get the broad shape of the distribution of tropical lows correct, with a maximum occurring in eastern Angola, the locus of the Angola tropical low. There is a wide range in the number of strong tropical low events per year. Most models show a lower count of tropical low days per year than the MERRA2 and ERA-5 reanalyses, while ERA-Interim

is roughly in the middle of the model distribution. This stands in contrast to the findings of Munday and Washington (2017), who report that the geopotential height anomaly associated with the Angola low is over-represented in CMIP5 models, although they did not consider ERA-5 or MERRA-2 for comparison. However, we do find that two models that have the most prevalent tropical lows (ACCESS1.3 and GFDL-ESM2G) also had strong geopotential height anomalies according to Munday and Washington (2017).

The tropical low latitudes are shifted overly southwards towards the western edge of the Namibian Caprivi strip (18°S , 20°E) in most models, notably BNU-ESM, CNRM-CM5, inmcm4, NorESM1-M and HadGEM2-CC. The number of tropical lows per day is notably low in IPSL-CM5A-LR, inmncm4 and MPI-ESM-LR. BNU-ESM and HadGEM2-ES show an overly strong peak at 20°E , with very few tropical low occurring outside the location of the Angola peak. For BNU-ESM, this is also clear from Figure 6.3, where the track longitudes are largely confined to 18° - 23°E . The biases of the spatial distribution of tropical lows in these climate models relative to era-5 reanalysis, and their significance relative to natural variability over the 30 year time period, are shown in Supplementary Figure 6.17.

Figure 6.10 shows the normalised distributions of four key properties of tropical lows: their longevities, latitudes, zonal velocities and T63 filtered vorticity. The southward shift of the tropical lows in CMIP5 models as compared to reanalysis is more clear in this histogram. Based on the distributions of longevity and zonal velocity, models may be divided into two groups: those with relatively more short-lived (<8 days) and relatively fewer long-lived tropical lows (10-20 days) than the reanalysis products, and those with fewer short-lived and more long-lived tropical lows. Those models in the first category tend to have a wider

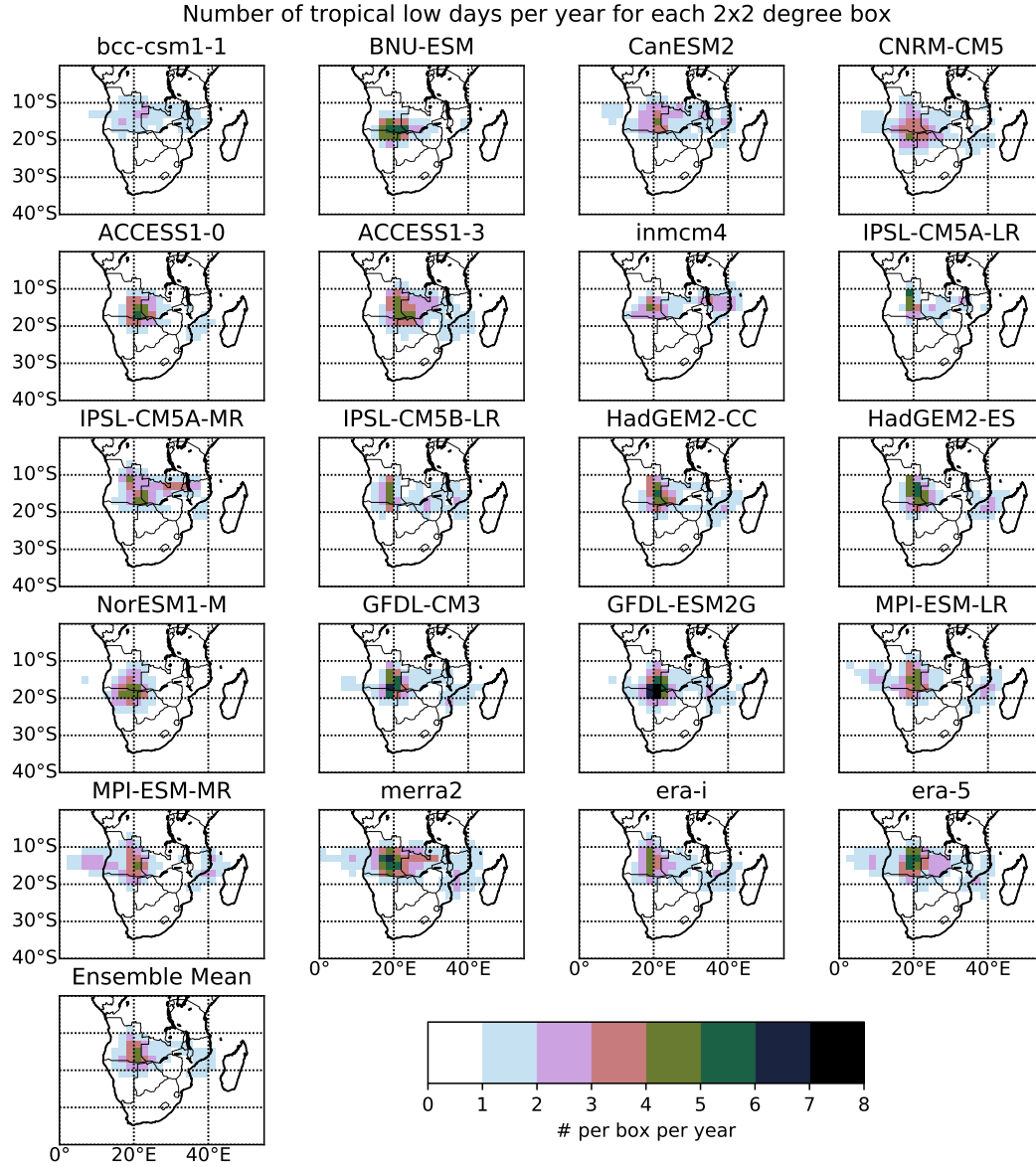


Figure 6.9. Spatial distributions of tropical lows per $2 \times 2^\circ$ grid cell. Panels indicate different models, ensemble mean and reanalysis products.

distribution of zonal velocities, with faster track speeds, while those in the second category have a greater proportion of stationary lows concentrated around 0 m/s. The first category therefore contains a greater proportion of transient events

that move off into the Atlantic Ocean, and the second contains a greater proportion of stationary Angola tropical lows. Typical extreme cases for each category are CNRM-CM5 and GFDL-ESM2G. Examining Figure 6.3 reveals a consistent story: for the case study year CNRM-CM5 is dominated by transient events (diagonal lines) while GFDL-ESM2G contains 4-5 long-lived events that meander across the continent and are frequency stationary, as well as some smaller events that are both stationary and transient. The distributions of the filtered vorticity largely follow that of reanalysis. Biases in all these quantities are significant relative to natural variability in at least 12 out of the 17 models considered.

The top right panel of Figure 6.10 shows the number of tropical low days per year. This metric reflects the overall magnitude of the signal in Figure 6.3 discussed earlier. The models with the lowest values (IPSL-CM5A-LR and inmcm4) also show a higher proportion of tropical lows just below the vorticity cut-off in the lower left panel of Figure 6.10.

Rainfall is attributed to tropical lows under the assumption that all rainfall that falls within 5 degrees of the centroid of a tropical low is associated with that tropical low. Rainfall is decomposed into a tropical low portion and a remainder portion. Chapter 4 found that 70% of rainfall in south west Angola and 31% of rainfall across the tropical edge region (16° - 22° S) was attributable to tropical lows. The spatial pattern of rainfall attribution for CMIP5 models is shown in Figure 6.11, and the overall proportion of rainfall attributed to tropical lows in each model between 15° - 30° E and 10° - 25° S is shown in the lower right panel of Figure 6.10. We find that approximately 30% - 60% of rainfall over southern Africa is associated with tropical lows between these latitudes. The spatial patterns of rainfall attribution match well with those found in Chapter 4 (reproduced here in lower panels of Figure 6.11). The biases of these spatial

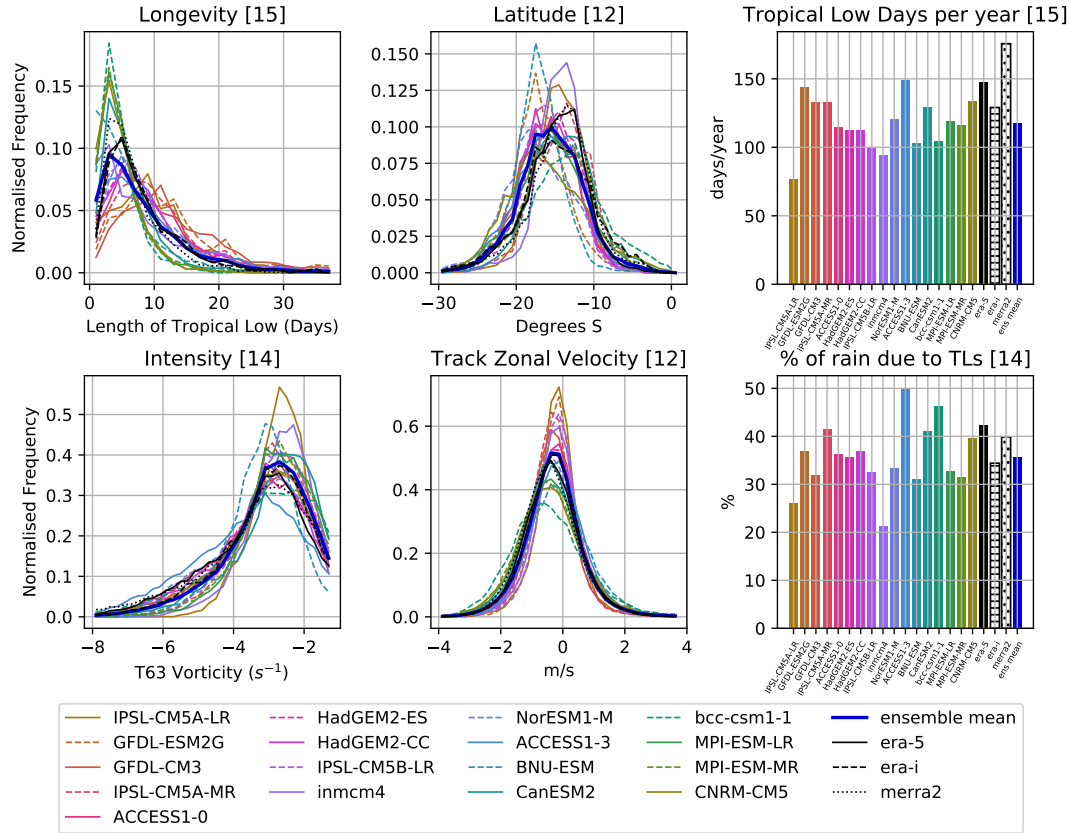


Figure 6.10. Normalised distributions (left and centre) and overall quantities (right) of tropical low properties. Top left: Longevity of TL events in days. Top centre: track latitude. Top right: total number of tropical low days per year for each model/reanalysis product. Bottom left: T63 filtered vorticity. Bottom centre: track zonal velocity, calculated as the tendency of the track longitude. Bottom right: proportion of rainfall attributable to tropical lows, based on the methodology described in section 4b averaged over 15° - 30° E and 10° - 25° S. Thin coloured lines indicate models, and are ordered by the maximum bin frequency of the upper left panel. Black lines show reanalysis products, thick blue lines show the ensemble mean. Numbers in brackets indicate the number of models which show a significant change relative to natural variability at a $p < 0.05$ level based on a Mann Whitney U-test for the left column, and a Welch's t-test for the remaining panels.

patterns relative to era-5 reanalysis, and their significance relative to natural variability over the 30 year time period, are shown in Supplementary Figure 6.18.

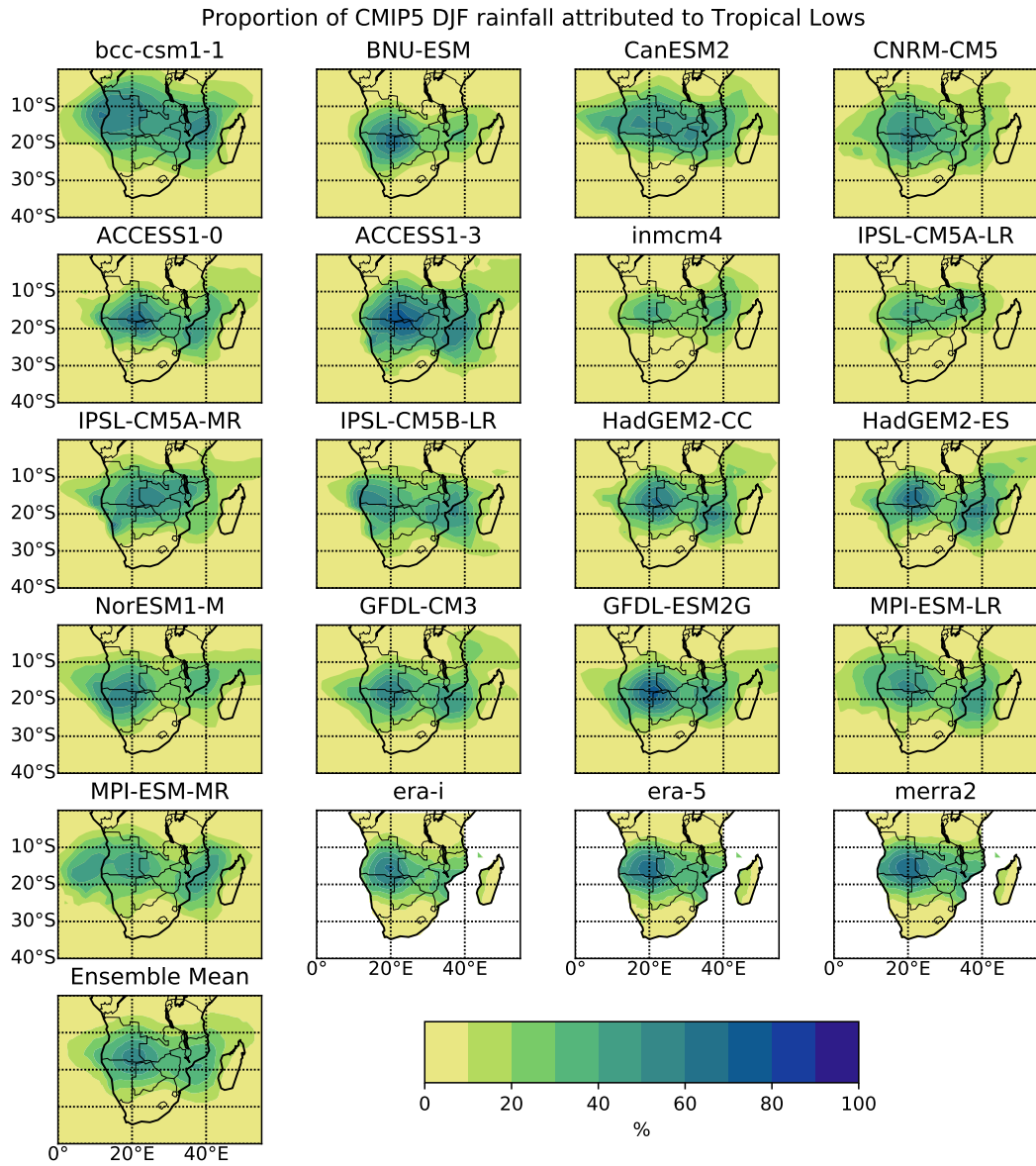


Figure 6.11. Proportion of rainfall attributed to tropical lows from December to February in historical CMIP5 sample. Rainfall is defined to be associated to a tropical low if it falls within 5° of the tropical low centroid. Panels show different CMIP5 models and the ensemble mean.

There are some differences between tropical lows in reanalysis products and in CMIP5 models, including latitude and longevity distributions, and there is a wide distribution across models in the mean number of strong tropical low days per year. However, tropical lows are consistently present in each model with key statistics varying by less than 20%. We therefore conclude that they are sufficiently resolved to examine projected tropical low changes and how those changes impact southern African precipitation projections.

6.5.2. Future change and rainfall implications

The overall trend in the spatial distribution of tropical lows between end of 21st century RCP8.5 and end of 20th century historical simulations is a decrease in tropical low frequency and in some cases a northward shift, as is shown in Figure 6.12. Some models, including ACCESS1.0, HadGEM2-CC, HadGEM2-ES and GFDL-CM3, show a sharp decline in tropical lows of about 2 per $2^\circ \times 2^\circ$ box per year that is restricted to the location of the peak of their historical tropical low locations. Northward shifts are evident in CanESM2, IPSL-CM5A-LR and MPI-ESM-LR. Only MPI-ESM-MR shows an increase in tropical low frequency. Despite many models showing similar overall patterns in tropical low decline, the spatial pattern of change and the locations of statistically significant changes varies widely across the models. The ensemble mean exhibits an overall 15% decline in the number of tropical low days that occur in each year.

Correspondingly, the spatial pattern of rainfall change between December and February, the main tropical low months, shows a high degree of variation between models, consistent with Lazenby et al. (2018). Filled green and purple contours show wetting and drying of the overall seasonal mean rainfall in each model in Figure 6.13. These are overlain by contours, in black and red, of the wetting

and drying of the component of the rainfall attributed to tropical lows. Changes which were not significant compared to natural variability at a $p < 0.05$ level were masked. It is evident that all the land-based local maxima and minima of total rainfall change located between 10° and 25°S correspond to an associated maxima or minima in tropical low rainfall change. Comparing with Figure 6.12, these changes correspond with spatial changes in the frequency of tropical lows.

In order to quantify this relationship, the Pearson's r coefficient of the spatial correlation between projected rainfall change and tropical low frequency change between $5^\circ - 25^\circ\text{S}$ for all land points is shown in the top row of Figure 6.14. The correlation for overall rainfall is low (typically 0.1 - 0.3) but consistently positive in all but two models. The correlation for the tropical low rainfall component is higher, with values averaging around 0.4. There is no consistent signal in the direction of correlation between the remaining rainfall pattern change and the tropical low distribution change. In this manner, the divergence in model rainfall projections in DJF over tropical southern Africa is linked to the uncertainty in the spatial response of tropical lows to climate change.

The ensemble mean tropical low rainfall change in this region (Figure 6.14, lower right panel) also corresponds to the ensemble mean tropical low spatial distribution change, with a decrease along 18°S centred south east Angola and an increase further north. The spatial pattern of total rainfall decline maps accurately onto the spatial pattern of tropical low rainfall decline in southern Angola and northern Namibia. However, the rainfall increase to the north of this region has its maximum further north in the Congo and is likely delivered by other synoptic systems.

Averaging over land points in the region from $5^\circ - 25^\circ\text{S}$, the projected frequency change in tropical lows is a good predictor of the inter-model spread of

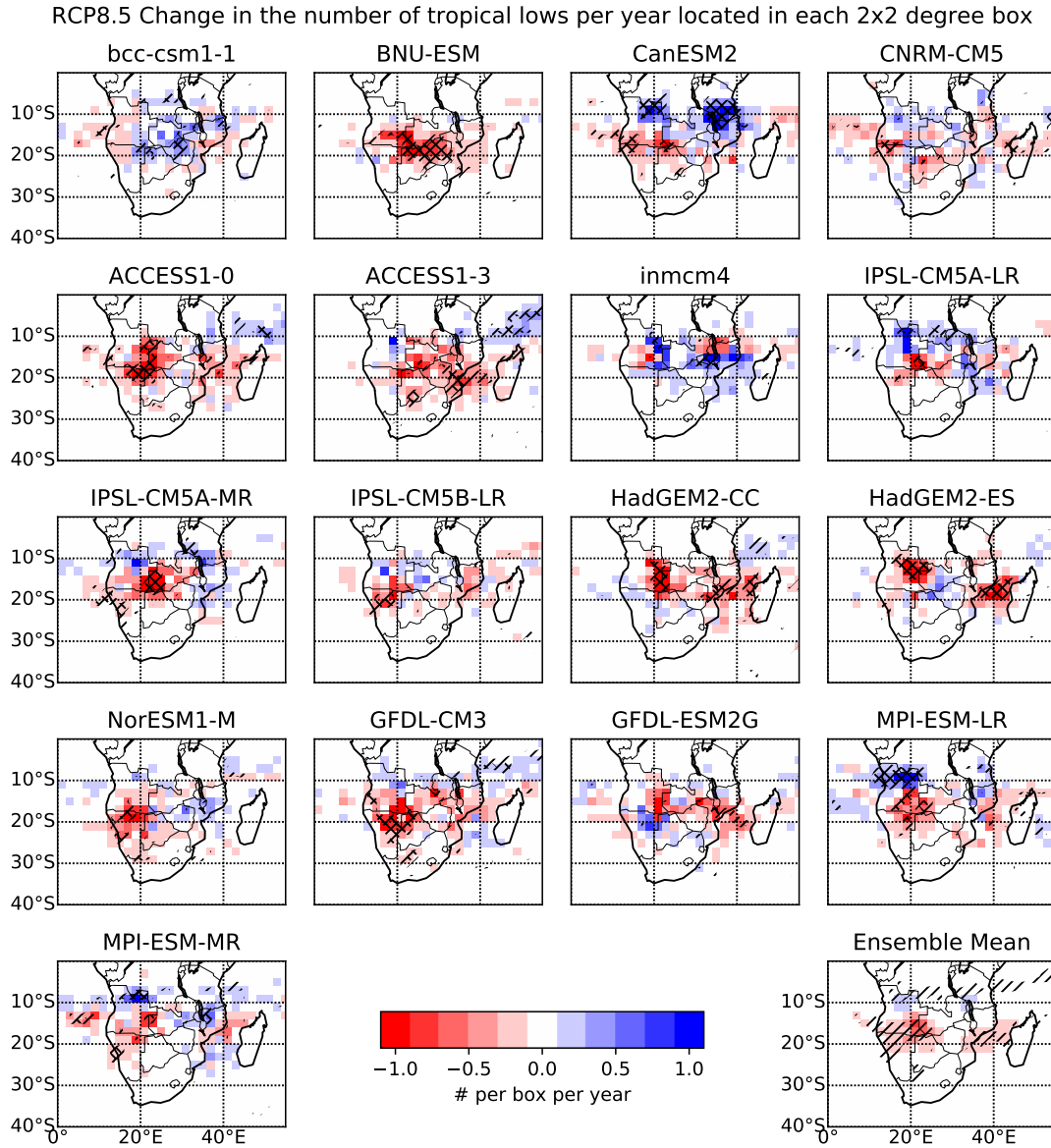


Figure 6.12. Changes in the distributions of tropical lows per grid-box per year between the RCP8.5 sample and the historical sample in each model and in the ensemble mean. Hatching on individual model panels indicates changes that are significant relative to interannual variability using a Welch's t-test. Forward (backward) hatching indicates significance at a $p < 0.1$ ($p < 0.05$) level. Ensemble mean: forward hatching indicates changes that are significant relative to the inter-model spread at a $p < 0.05$ level using a paired t-test.

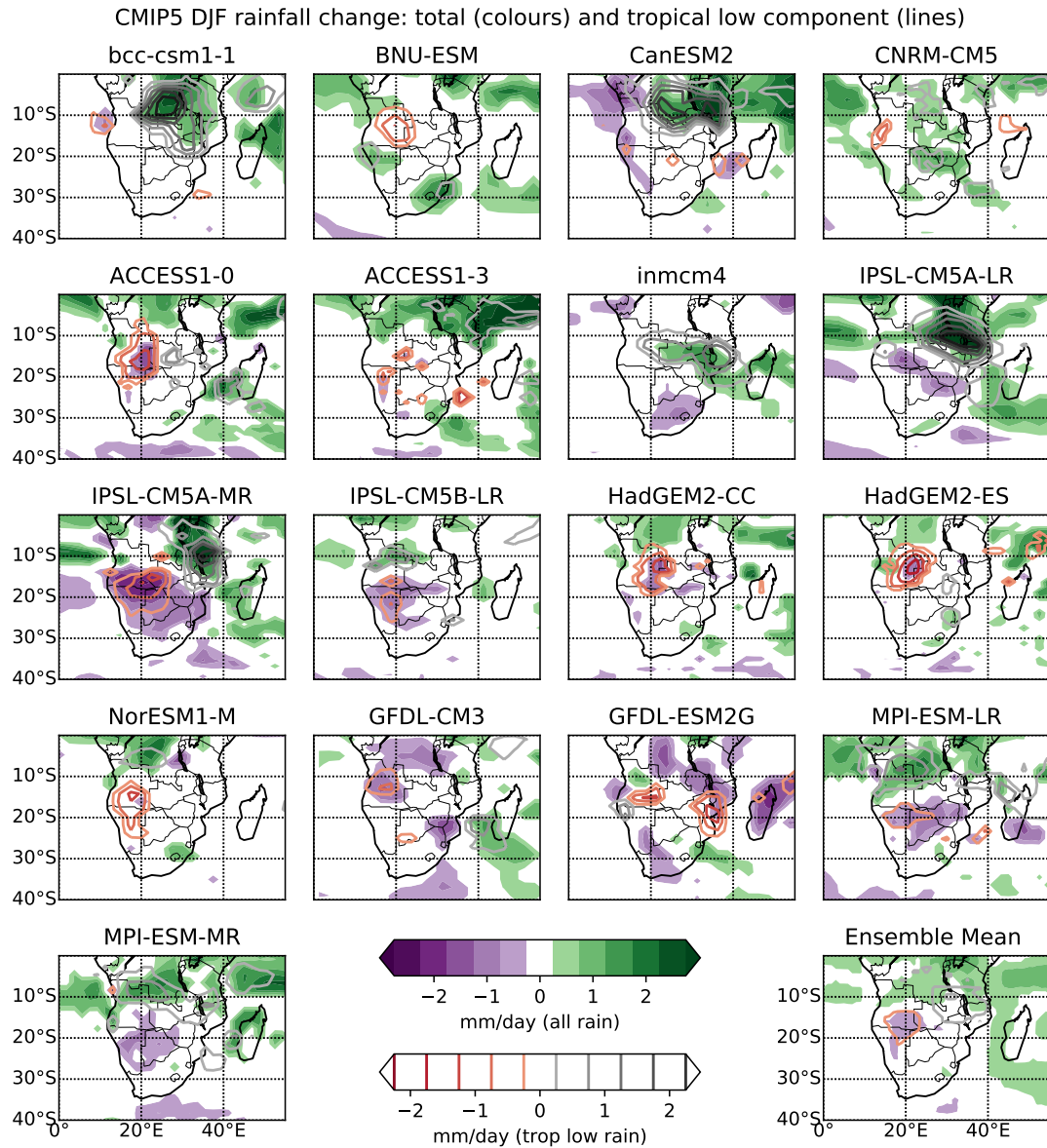


Figure 6.13. Filled contours: total rainfall change between RCP8.5 end of 21st century sample and historical end of 20th century sample for December and February over southern Africa. Purple: decrease, green: increase. Line contours: same as filled contours, but only for rainfall that has been attributed to tropical lows. Data that is insignificant at a $p < 0.05$ level relative to interannual variability (inter-model spread) based on a Welch's t-test (paired t-test) is masked for each model (for the ensemble mean).

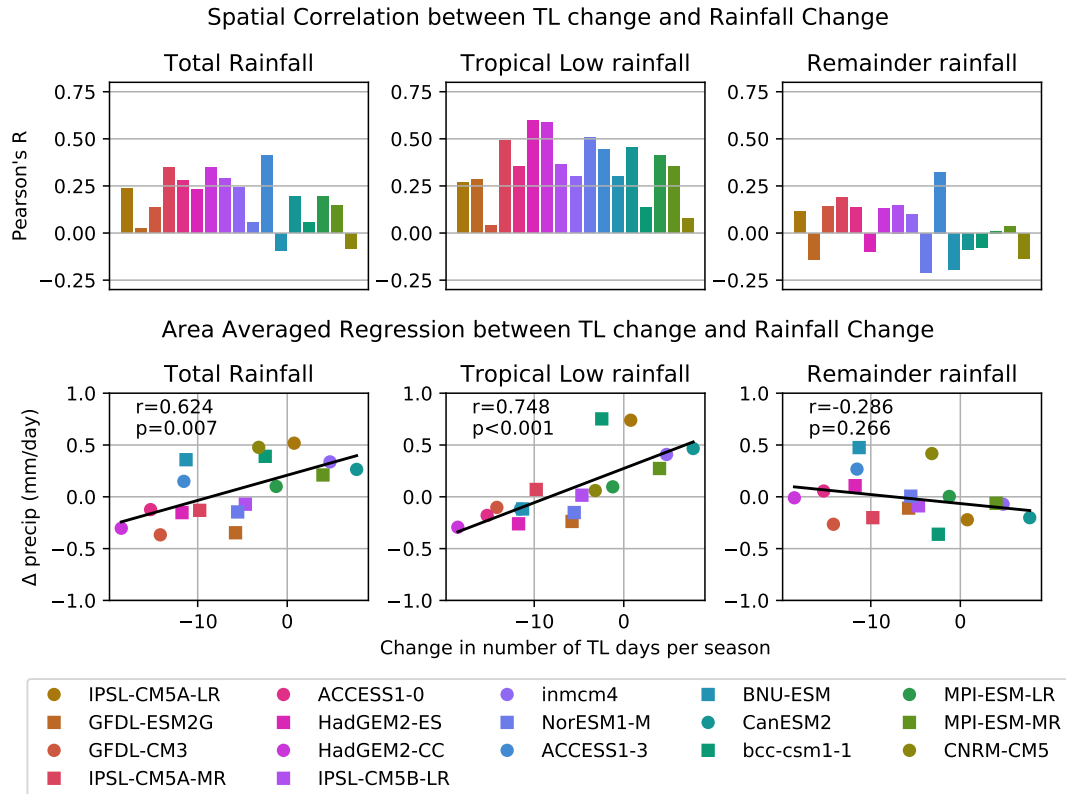


Figure 6.14. Upper row: spatial correlation between rainfall change components and tropical low spatial distribution change. Rainfall change and tropical low distribution change were calculated on a $2^\circ \times 2^\circ$ grid before correlation, and only mainland points between 5° - 25° S were considered. Lower row: inter-model regression between overall rainfall change and tropical low frequency change, averaged over the same region as listed above. Left column: change in total rainfall field. Centre column: change in rainfall located within 5 degrees of a tropical low. Right column: remainder rainfall.

rainfall change ($r=0.58$, $p=0.015$). This is indicated in the regression shown in Figure 6.14 (lower left panel). The decomposition into rainfall associated with tropical lows and a remainder component (lower centre and right panels of Figure 6.14) indicates that this influence of tropical lows is direct, as the change in tropical low rainfall is significantly correlated with the change in tropical lows, and the change in the remainder term is not. Therefore, the spread in future projections

of tropical lows is a major contributor to the spread of rainfall predictions in DJF over southern Africa.

6.6. Discussion

6.6.1. Remarks on the Congo Air Boundary

We have found that the CAB is well represented in CMIP5 models, and that its projected frequency increase is able to distinguish between models that show strong and weak declines in OND rainfall south of 10°S . In this section, we discuss the implications of the historical representation of the CAB, and compare our results to other studies of the OND rainfall decline.

The accuracy of historical representation of the CAB in CMIP5 models is reasonable, despite the perpendicular width of the CAB being of order 100 km (Chapter 5), below the resolution of most CMIP5 models. The southward displacement of the CAB between August and December is roughly 8 degrees of latitude, which in the coarsest models is equivalent to 4 grid cells, and yet the rate of the CAB's progression is very similar across models. This suggests that representation of the CAB, and its seasonal progression, is controlled by a process that climate models do not struggle to simulate, and so is not controlled by processes that act at the grid box scale. As with the tropical rainbelt globally, the seasonal progression is clearly associated with the progression of the latitude of maximum solar insolation. Meanwhile, the sharp gradients in humidity bring to mind bifurcations into dry and moist convective states observed in idealised radiative convective equilibrium models (e.g. Emanuel et al., 2014).

We have found that projections of CAB frequency have different consequences north and south of the climatological CAB. The relationship between the CAB frequency and rainfall south of the CAB's climatological position is an intuitive

one. In this region, saturation is rarely achieved when the CAB is present, unless extra-tropical processes dominate. Most rainfall comes from CAB breakdown events. When the gradual seasonal CAB breakdown is delayed, as in the future model projections, there are fewer rainy days in this region, and hence less rain. Meanwhile, there is an overall increase in rainfall to the north of the climatological CAB. However, this increase is not limited to the location of the CAB, as OND rainfall is projected to increase across much of equatorial Africa (Creese et al., 2019). Therefore, there is a consistent projected increase in rainfall in locations that are generally located the north of the CAB, but this is not necessarily driven by the corresponding projected increase in CAB frequency. The contrasting impacts that the CAB frequency has to its north and south are important for determining the precise location of projected rainfall decline. Rainfall to the south of the mean CAB position is projected to decline due to the projected increase in CAB frequency, while rainfall to the north is projected to increase. Therefore, the CAB may be expected to set the location of the boundary of rainfall decline.

The work presented here is consistent with previous studies of rainfall projections over southern Africa. Dunning et al. (2018) found that the seasonal progression of the African rain belt is tied up in the strengths of the Saharan and Angola heat lows, both of which are projected to intensify in a warmer world. Munday and Washington (2019) found a similar result, that models with a deeper future Angola heat low showed a higher intensity of drying, while Cook and Vizzy (2013) observed a strengthening of the heat low in regional climate model simulations. The Angola heat low is intrinsically linked to the CAB (Chapter 5), being consistently located approximately 1° south of the CAB and sharing circulation features. Therefore an increase in the climatological CAB frequency corresponds to an increase in the heat low intensity.

By applying the Chadwick et al. (2013) decomposition, Lazenby et al. (2018) attributed the OND rainfall change to dynamic rather than thermodynamic changes, associated with spatial shifts in the pattern of convective mass fluxes. In the context of this methodology, it is useful to consider the atmosphere to the north of the CAB and during CAB breakdown events as capable of deep convection, and to the south of the CAB as incapable of deep convection. The increased CAB frequency corresponds to a decrease in the proportion of the time that convection may occur in the region south of 10°S , and so is associated with a shift in deep convection away from southern Africa.

Figures 3(f-g) in Lazenby et al. (2018) also demonstrated that there was a decrease in near surface relative humidity between 10° and 20°S , but that this did not contribute to the overall precipitation change budget. This finding underscores the fundamental concept outlined by the CAB: that the southern African atmosphere during the Austral spring is often characterised at any instant by a dichotomy of states: one that is severely moisture limited and one that is close to saturated, with very little in between (Chapter 5). In such a scenario, small thermodynamic perturbations have less potential to change the likelihood of convection than small dynamical perturbations. This is because a dynamical perturbation may move the moist airmass into, or away from a given location, drastically changing the likelihood of convection at that spot. Meanwhile, a small change in saturation temperature is unlikely to change the likelihood of convection if the atmosphere is either severely dry or not at all moisture limited.

6.6.2. Remarks on Tropical lows

Uncertainty in future projections of tropical lows in southern Africa has been identified as a source of the uncertainty and broad model spread in rainfall projections in this region during the main rainy season. This uncertainty derives from variations in the changes of the climatological tropical low spatial distribution, however some robust changes are present, as discussed below.

The historical spatial distribution of tropical lows features a southward bias as compared to reanalysis. Models show a greater proportion of tropical lows centred near 18°S and too few located around 15°S. This bias is consistent with other studies of the historical CMIP5 rainfall bias over southern Africa. An interannual southward shift in the Angola low has been associated with an increase in rainfall over subtropical southern Africa (Pascale et al., 2019; Crétat et al., 2018), consistent with the direction of the rainfall bias (Lazenby et al., 2016). Furthermore, Munday and Washington (2017) found a correlation between the magnitude of this wet bias and the strength of the Angola low, which they found was overly intense in most models. They also found that the climatological position of the Angola low was shifted towards the Angola-Namibia border in many models.

Despite differences in the patterns, there is a consensus in all but one model of overall tropical low decline south of 15°S. This is reflected in the ensemble mean of both precipitation and tropical low location shifts. A large minority of models also show a marked increase in tropical lows north of this line, signifying a northward shift. However, this change is less robust than the decline further south. A northward shift and overall decline of tropical low location mirrors the characteristic response of tropical lows to El Niño events (Pascale et al., 2019, and Chapter 4).

Lazenby et al. (2018) proposed that the diverse range of future projections in DJF is associated with inter-model differences in the SST changes in the Indian and Atlantic Oceans. Since interannual variability of the Angola low, which in late summer is the climatological expression of tropical lows, is sensitive to SSTs in these oceans (Pascale et al., 2019), it is likely that variability in SST projections may lead to variability in tropical low distributions, which in turn lead to the spread of projected rainfall changes.

6.7. Conclusion

The CMIP5 ensemble predicts a drying trend in both spring and summer in southern Africa. Spring drying is robust across all models, and is associated with a delay in the wet season onset (Dunning et al., 2018). Such delays would cause a shortening of the growing season, impacting regional agriculture and food security (Lobell et al., 2008; Schlenker and Lobell, 2010). Model projections in the summer are more divergent and the ensemble mean drying trend in this season is less robust. However, rainfall changes in this season would be equally disruptive to the local economy. For this reason, it is imperative that the mechanisms of future rainfall change demonstrated by CMIP5 models are well understood.

The present study has placed the spring and summer projected rainfall change in the context of two classes of weather events, tropical lows and the CAB. These features are predominant in each respective season and are sufficiently well resolved in the climate models. By doing so, we have attributed model consensus on spring drying to increases in CAB frequency and latitude, and the lack of model consensus on summer rainfall change to uncertainty in the tropical low response to a warmer world. This work provides crucial context for the projected changes and grounds changes to the local climate in meteorological theory.

We have found that the spring drying trend is strongly correlated across models with a increased frequency in the CAB. This delayed breakdown causes an increase in its average frequency between October and December of approximately 13%. Model consensus on its projected change agree that the CAB will be located 0.5° - 1° further north and its gradual seasonal decline will occur half a month later at the end of 21st century than the 20th under the RCP8.5 scenario. These changes are significant as against internal variability in 12 - 15 of the 18 models. Therefore, we strongly advocate for further research into this understudied feature of the climate system, particularly in the form of observational field campaigns.

Tropical lows, which make up the synoptic expression of the Angola Low in the southern African summer, show more model spread in their future change, particularly when considering small-scale spatial shifts. This increased model uncertainty is a contributor to the divergence of DJF rainfall projections in southern Africa. However, despite spatial differences, a statistically significant decline in tropical low frequency south of 15°S is present in 11 models, with an overall ensemble mean decline of 10%. The response of tropical lows to future warming will be important for rainfall change in this region.

6.8. Supplementary material

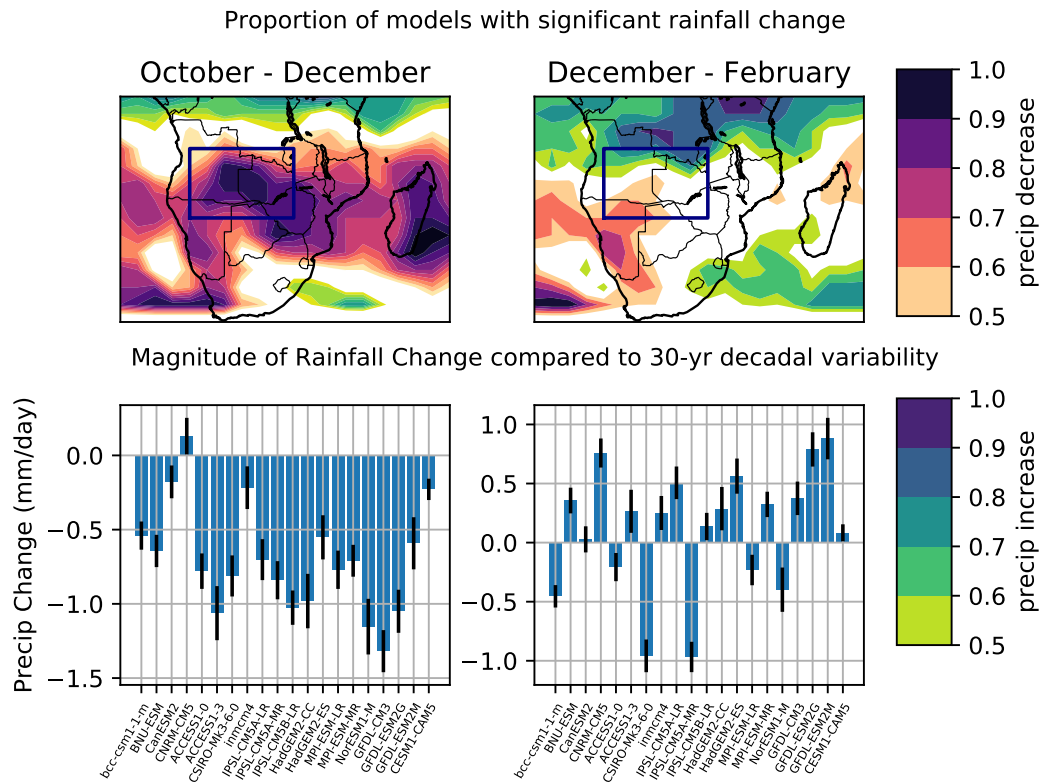


Figure 6.15. Significance of the precipitation changes in 2070-2100 compared to 1970-2100, against 30 year decadal variability of the pre-industrial control simulations. Left: October to December precipitation, Right: December to February precipitation. These have been calculated by taking the a 30 year rolling average of seasonal precipitation over all available pi-control simulation years, and then calculating the standard deviation at each grid-point, (σ_{30}). The upper panels show all locations where over 50% of the the climate models considered exhibit a precipitation increase (decline) greater than σ_{30} in green (pink) colours. The lower panels show the magnitude of each model's precipitation decline, averaged over 15° - 30° E and 10° - 30° S. Error bars indicate the model σ_{30} averaged over the same area.

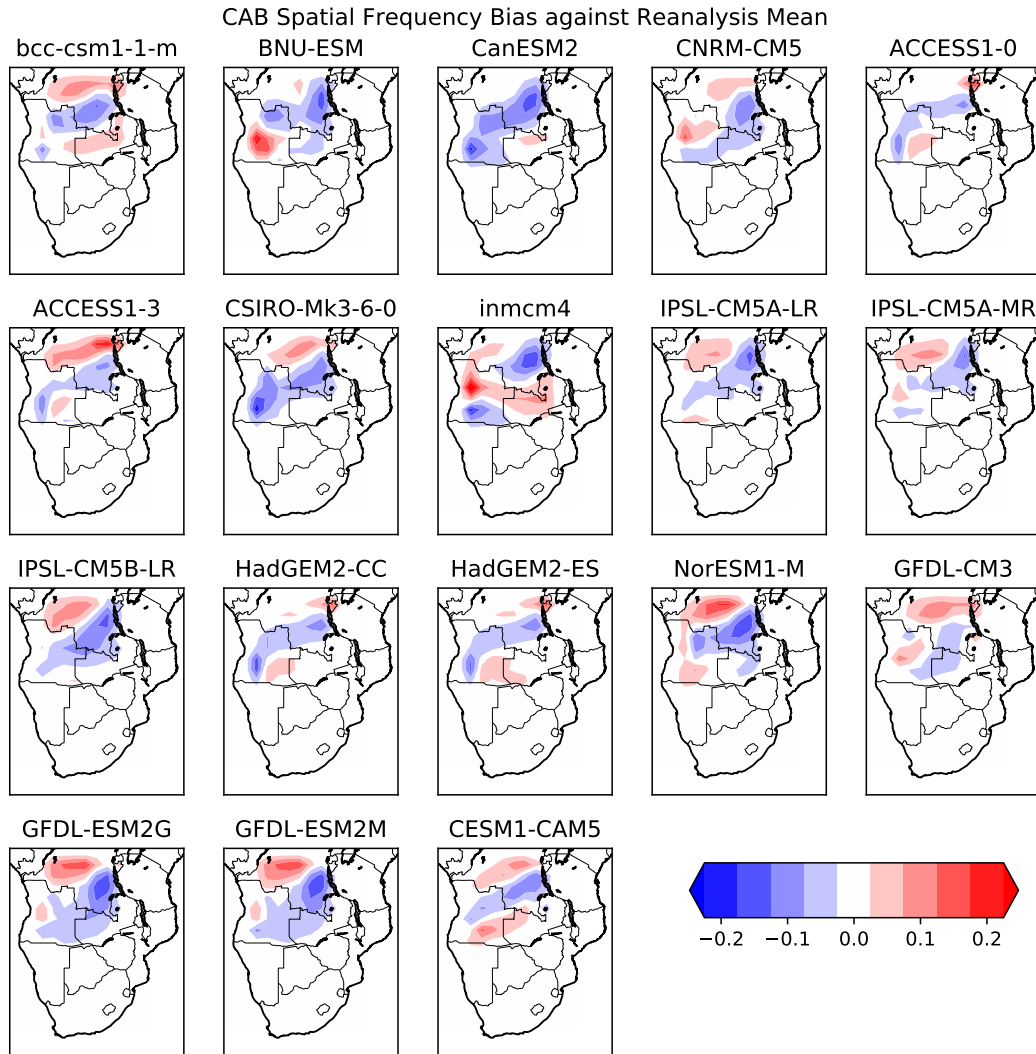


Figure 6.16. Biases of spatial distributions of CAB (number of events per interpolated grid cell per year) relative to era-5, for each model and the ensemble mean. In model panels, data which is insignificant compared to interannual variability at a $p < 0.05$ level based on a Welch's t-test is masked. Ensemble mean is masked where data is insignificant at a $p < 0.05$ level compared to inter-model spread using a one-sample t-test. Panels indicate different models.

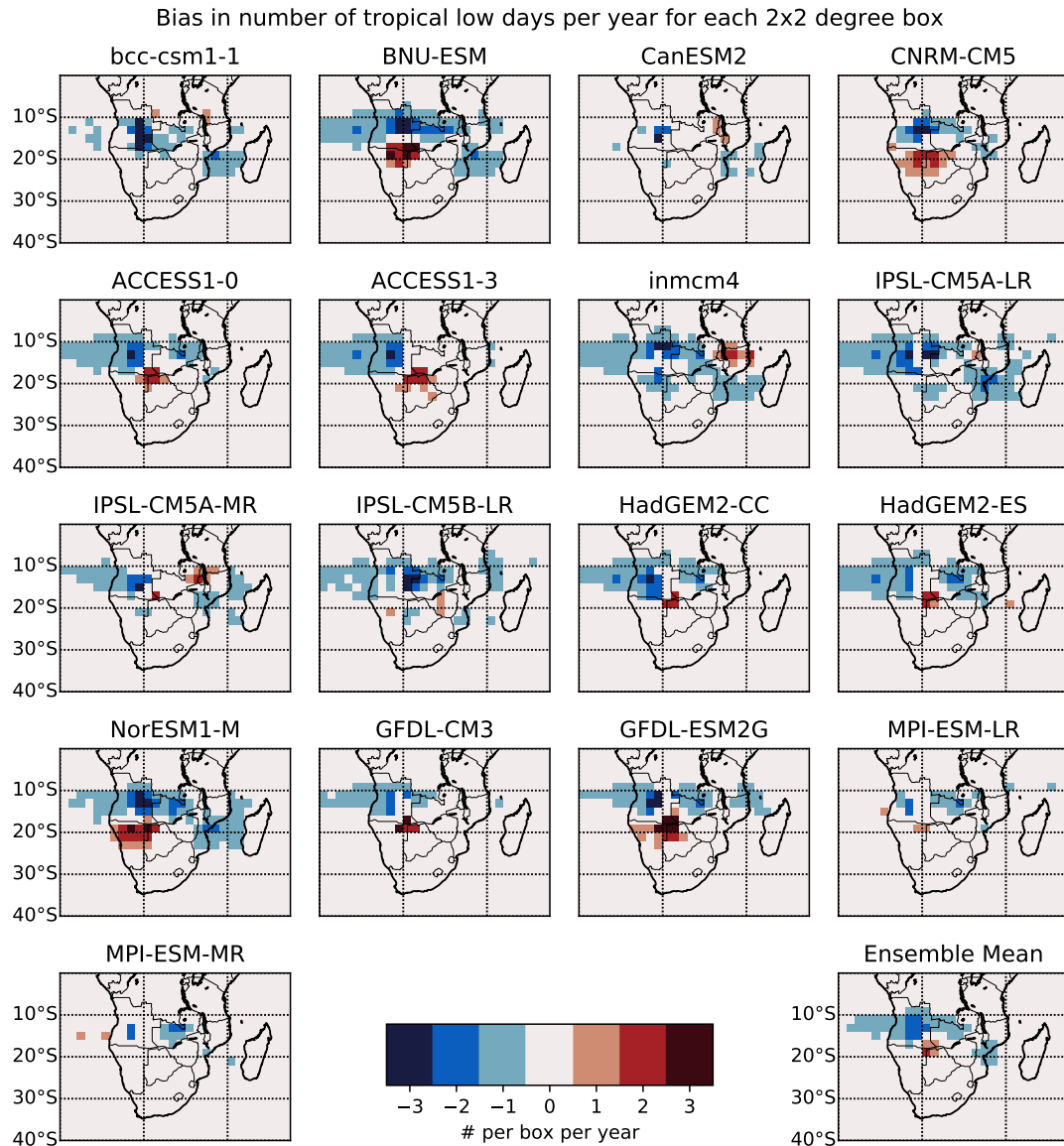


Figure 6.17. Biases of spatial distributions of tropical lows per $2 \times 2^\circ$ grid cell relative to era-5. In model panels, data which is insignificant compared to interannual variability at a $p < 0.05$ level based on a Welch's t-test is masked. Ensemble mean is masked where data is insignificant at a $p < 0.05$ level compared to inter-model spread using a one-sample t-test. Panels indicate different models, ensemble mean and reanalysis products.

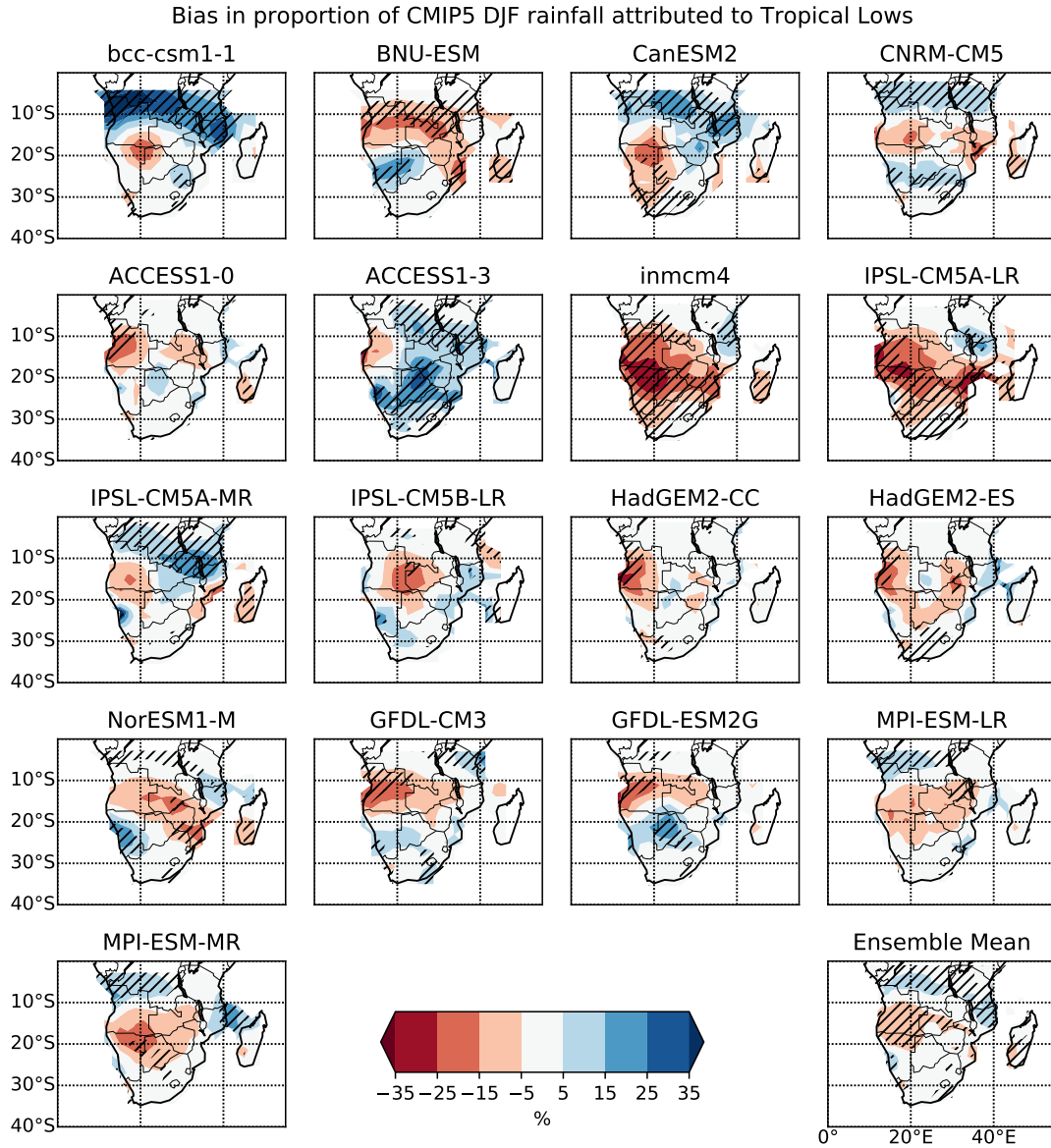


Figure 6.18. Biases in the proportion of rainfall attributed to tropical lows from December to February in historical CMIP5 sample relative to era-5. In model panels, hatching indicates data which is significant compared to interannual variability at a $p < 0.05$ level based on a Welch's t-test. In the ensemble mean panel, hatching indicates where data is significant at a $p < 0.05$ level compared to inter-model spread using a one-sample t-test. Panels indicate different models, ensemble mean and reanalysis products.

6.9. Acknowledgements

This work has been generously supported by the Origin Foundation John Monash Scholarship and by the Future Climate for Africa UMFULA project, with financial support from the U.K. Natural Environment Research Council (NERC), NE/M020207/1, and the U.K. Government's Department for International Development (DfID). The GCM data used in this study were made available through the Earth System Grid Federation (ESGF) Peer-to-Peer system (<https://pcmdi.llnl.gov/search/cmip5/>). Reanalysis data used in this analysis were provided by the Copernicus Climate Change Service (<https://cds.climate.copernicus.eu/cdsapp#!/home>), the European Centre for Medium Range Weather Forecasts (<https://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=pl/>) and NASA (<https://disc.sci.gsfc.nasa.gov/daac-bin/FTPSubset.pl>). Colour maps were sourced from cmocean (<https://matplotlib.org/cmocean/>). We thank Kevin Hodges for access to TRACK and Paul-Arthur Monerie for assistance regarding decadal variability.

6.10. Afterword to Chapter 6

Chapter 6 has demonstrated that tropical lows and the CAB are on the whole well represented in the historical CMIP5 archive. It found that tropical lows are generally shifted further south in CMIP5 models compared to reanalysis products, consistent with the prevalent over-prediction of historical rainfall. This chapter has expressed projected rainfall change in southern Africa in terms of process based projections of tropical lows and the CAB. By doing so, it has fully addressed the requirements of Aim 5.

Chapter 6 has used the algorithms and methodologies developed in Chapters 4 and 5 to investigate the representation and future change of the CAB and tropical

lows in CMIP5 models. The algorithm from Chapter 4 is used with minimal modification, though the pressure levels used for vorticity input were changed based on data availability. This was made possible by the spatial filtering step, which removes resolution dependence from the algorithm output. The algorithm from Chapter 5 was modified such that it would operate at coarser resolution.

CHAPTER 7

Conclusions

7.1. Introduction

Southern Africa's reliance on rainfall for domestic, industrial and agricultural use exposes the region to a large degree of climate related socio-economic risk. Future climate projections indicate that the wet season onset is likely to be delayed, causing a reduction in the length of the growing season. Under business as usual anthropogenic forcing, this amounts to half a month delay in the wet season onset date by the end of the 21st century (Dunning et al., 2018). Delayed rainfall onset is likely to negatively impact agricultural yields and impede economic growth. However, these rainfall trends exist in the context of poorly understood regional climate dynamics, which hampers certainty regarding the realism of these projections. In order to better constrain climate projections, a more comprehensive understanding of climate processes - particularly at the synoptic time-scales at which weather systems evolve - is necessary.

This thesis set out to improve understanding of climate processes expressing the climatology, dynamics and future change of the southern African tropical margin from the viewpoint of synoptic weather systems. It has achieved this aim by considering tropical lows and the CAB, two key synoptic weather systems that contribute to rainfall in the tropical edge. Algorithms were optimised or developed to identify the CAB and tropical lows in daily and subdaily gridded

meteorological datasets. These algorithms were then used to study the characteristics of tropical lows and the CAB, including frequency distributions, seasonal cycles, explanatory power over interannual rainfall variability, and dynamics. The penultimate chapter applied these algorithms to CMIP5 models in order to undertake a process-based evaluation of the representation of southern African synoptic in historical and future projected coupled model simulations.

Future change of southern African rainfall was then expressed in terms of changes of the CAB and tropical lows. It was found that the spring drying signal could be explained by a northward shift and delayed breakdown of the CAB. Meanwhile, future change in the portion of rainfall attributable to tropical lows contributed to model disagreement over the spatial pattern of rainfall change in the summer season. However, models agree on an overall decline and northward shift in tropical low tracks in the future, consistent with the ensemble mean pattern of rainfall change. This finding suggests that the ensemble mean projected summer rainfall decline may be more robust than rainfall projections suggest.

This final chapter reviews the results presented in this thesis. First, the results of the four papers are individually summarised. Second, the individual aims and goals of the thesis are revisited, and the means by which each aim was achieved is presented in the context of tropical lows, and then in the context of the CAB. Following this, the themes linking the chapters will be discussed. This discussion considers the role of surface wind convergence in governing tropical low and CAB behaviour, the means by which the CAB and tropical lows influence the spatial distribution of rainfall, implications for predictability, and parallels between the CAB and convective organisation. Finally, the broader implications of the results of this thesis for climate science are reviewed and potential angles for future research are discussed.

7.2. Summary of main results

7.2.1. Chapter 3: Synoptic expression of the Angola Low

Chapter 3 focussed on identifying which synoptic weather systems aggregate in the climatological mean to form the Angola Low. It was found that two distinct forms of weather systems, heat lows and tropical lows, contributed towards the cyclonic vorticity and low pressure anomaly of the Angola Low. The former systems occur during spring, and a transition to tropical lows occurs in late November. Tropical lows then persist into late March.

Heat lows and tropical lows were identified as maxima in cyclonic daily mean vorticity, and distinguished on the basis of their static stability. Upon separation, spatial distributions, composite cross sections and vertical profiles of the two classes of vortices were examined. It was found that the tropical lows featured a tall column of cyclonic potential vorticity reaching up to 300 hPa, cyclonic circulations with radius of about 1000 km, cold surface temperatures and wind convergence limited to the lowest 100 hPa. Meanwhile, heat lows are shallow systems with cyclonic vorticity extending only up to 800 hPa, with a capping anti-cyclone centred at 600 hPa. Heat lows are associated with a maximum in surface temperatures and feature a strong diurnal cycle, with accelerations of radial and tangential winds balancing pressure gradient and Coriolis terms that are out of phase by half a day. The transition from heat low to tropical low occurred when a heat low disappeared and a tropical low formed in its place a few days later. Synoptic heat lows were never observed transitioning into tropical lows. In the afterword to Chapter 3, the heat low was shown to manifest as the CAB, which appears as a sharp gradient of specific humidity in the north-south cross section. This is further explored in Chapter 5.

The Angola Low has previously been linked to interannual rainfall variability. Reason and Jagadheesha (2005) proposed that the Angola Low modulates the role of ENSO in southern Africa. In Chapter 3, this relationship was examined in the context of the heat low/tropical low split. It was found that the tropical low component of the Angola Low, and not the heat low component, was responsible for the rainfall interannual variability signal. 27% of the rainfall residual variance was explained by residual of a tropical low index after regression onto the Niño 3.4 SST index. The prevalence of tropical lows also found to be weakly correlated with ENSO.

While heat lows across the subtropics are typically stationary features, tropical lows in other regions of the world are transient systems. In order to explain the dynamics of the unusual semi-stationary behaviour of tropical lows in Angola, composite vorticity budgets were analysed. It was found that the low level wind convergence between the west coast sea breeze and the easterly Indian Ocean trade winds, which form a fundamental component of the Angola Heat Low circulation, remains present throughout the summer. This additional wind convergence acts to boost the vorticity of tropical lows when they are located in Angola through vortex stretching, anchoring them to the region.

Chapter 3 demonstrated the means by which the Angola Low, a key feature of the climatology, is expressed on synoptic time-scales. Dividing and compositing the tropical low and heat low phases was crucial for understanding the dynamics of the Angola Low. The transition from heat low to tropical low explained the late-spring change in the vertical extent of the Angola Low, from shallow to deep. The anchoring mechanism of tropical lows explains the location of the Angola Low in western Angola. Meanwhile, the decomposition into tropical lows and heat lows added clarity on the relationship between the Angola Low, ENSO and

southern African rainfall, with only the tropical low component being relevant for this relationship and with tropical low intensity being weakly correlated with the Niño 3.4 SST index.

7.2.2. Chapter 4: Tropical lows

Chapter 4 implemented an algorithm designed to identify and track tropical lows in southern Africa, and analysed the tracks, rainfall and variability of the resultant weather systems. This algorithm was based on the atmospheric feature tracker presented by Hodges (1994) and Hodges (1999), modified to match the characteristics of the tropical lows described in Chapter 3.

Tropical lows are transient low level cyclonic vortices that occur during monsoons and wet seasons in India, northern Australia, and the off-equatorial low latitude oceans. In these regions, tropical lows have been shown to contribute a significant proportion of local precipitation. In addition, oceanic tropical lows occasionally intensify into tropical cyclones, which can be associated with extreme impacts when they approach land. Chapter 4 indicated that the behaviour of southern African tropical lows is qualitatively comparable to those in other regions of the world. However, they differ with respect to their semi-stationary behaviour, which is caused by the anchoring mechanism described in Chapter 3.

Chapter 4 found that land-based tropical lows occur predominantly in Angola and western Zambia, but also exist throughout a latitude band between 10° and 20°S. The median longevity of tropical lows was 4.5 days, with some lasting up to a month. They predominantly track eastwards, but slow down near 20°S and rarely reach the Atlantic Ocean. A weak relationship between tropical lows and ENSO was demonstrated, with tropical lows occurring at higher (lower) frequencies and

further south (north) during the La Niña (El Niño) phase, but also exhibiting a significant degree of variability independent of ENSO.

Tropical lows embody the synoptic expression of the Angola Low during summer, just as TTTs embody the synoptic expression of the SIOCZ. Since previous studies have proposed a relationship between the SIOCZ and the Angola Low on climatological timescales (Cook, 2001), Chapter 4 investigated whether TTTs and tropical lows were linked on a synoptic timescale. However, timing of these events were shown to be independent and rainfall anomalies from coincident events were indistinguishable from the super-positions of TTT and tropical low rainfall anomalies, and so no such synoptic relationship was found.

Following this analysis, rainfall was attributed to tropical lows, so that the impact of tropical lows on interannual rainfall variability could be considered. It was found that 70% of rainfall in south-west Angola and 31% of rainfall across the tropical edge region (16° - 22° S) was attributable to tropical lows. Furthermore, the enhanced interannual variability in Botswana and northern Namibia was found to derive from the tropical low component of rainfall. The delivery of the ENSO rainfall signal in across southern Africa to the west of 28° E was traced to tropical lows. Tropical lows were found to provide a key component of rainfall variability in the main rainfall season in southern Africa.

In the afterword to Chapter 4, it was demonstrated that the relationship between rainfall and the Angola Low is delivered directly by tropical lows. Only the proportion of rainfall that falls within 5° of a tropical low was found to be correlated to a tropical low index. Thus, the relationship between tropical lows and rainfall is prognostic rather than simply diagnostic. However, this relationship derives from rainfall directly associated with nearby tropical lows, rather than remote moisture flux around the climatological Angola Low.

In summary, Chapter 4 has shown that tropical lows are a key contributor to interannual rainfall variability in southern Africa. It was found that tropical lows are a major source of precipitation in the tropical edge region. Variability in rainfall that was attributed to tropical lows dominated the total interannual rainfall variability in Botswana and northern Namibia. This region exhibits a local maximum in rainfall variability as a fraction of annual mean rainfall, and was used to define the focus region of this thesis. This research has shown that the variable component of rainfall in this region is associated with tropical lows.

7.2.3. Chapter 5: The Congo Air Boundary and the Kalahari Discontinuity

The second synoptic weather system studied in this thesis was the combined surface drylines and convergence lines which form the southern boundary of the African rain belt. An algorithm designed to identify these systems in ERA-5 reanalysis was presented in Chapter 5. This was then followed by a detailed and systematic consideration of the frequency, climatology and dynamics of these drylines in the Austral spring.

Chapter 5 identified two combined dryline/convergence line systems: the CAB and the Kalahari Discontinuity (KD). Their spatial distributions, seasonal cycles and diurnal cross sections were examined, and the relationship between the CAB and rainfall was assessed.

The CAB is a laterally extensive weather system, extending 1000 - 1500 km across southern Africa on average. The CAB was found to steadily shift south from the Congo Basin into Angola and Zambia between August and October. The CAB is associated with the confluence between north-westerly winds that originate from the Congo, and the Indian Ocean trade winds. From August to

the end of September, it is present as a dryline on over 90% of days. At this time, the abutting air-masses have very different moisture contents, with the easterlies being much drier and the south westerlies close to saturation. Later in the season, the easterlies steadily become wetter and the CAB begins to break down. These breakdown events allow Congo air to penetrate into the subtropical interior of southern Africa, and are sometimes associated with TTTs. By December, the CAB is rarely present.

The KD was detected throughout the spring season as a convergence line along the west coast of southern Africa. The KD marks the confluence between the Indian Ocean trade winds and the southerly Namibian nocturnal low level jet. In the early season, both air-masses are dry, and so the convergence line is not associated with a dryline. However, as the spring progresses, the Indian Ocean trade winds become moister, especially when the CAB breaks down. When this occurs, the KD becomes a dryline. When a CAB breakdown is associated with a TTT, this dryline forms the western trailing edge of the TTT system.

Composite vertical cross sections of the CAB indicated that the sharp humidity gradient extends 200 hPa above the surface into the lower troposphere, and that it is co-located with a low-level cyclonic vortex and a surface temperature maximum slightly south. These surface temperature and wind composites were indicative of heat lows, and it was demonstrated that the dynamics of the diurnal cycle of the CAB are linked with those of the Angola Heat Low. The thermally direct winds associated with the heat low, which form overnight, advects the moist air to the north of the CAB southwards and strengthen the gradient of the CAB. Meanwhile during the day, dry turbulent convection transports the surface moisture aloft as the boundary layer grows, and the humidity gradient

weakens. Similar results were found for the KD and the Kalahari Heat Low, but were omitted for brevity.

The latitude and frequency of the CAB were statistically linked with interannual rainfall variability, and this relationship was found to have an intuitive interpretation. For each month between August and December, the climatological locus of the CAB was identified, and interannual time series of rainfall in this vicinity were considered. Between 50 and 70% of the variance of each time series was explained by either the CAB latitude in the early months, or the CAB frequency in the later months. Evidently, most rainfall in the early months occurs to the north of the CAB, since the region to the south of the CAB is far too dry for saturation to be reached and since the CAB rarely breaks down at this time of year. Therefore, the further south the CAB is, the further into southern Africa this rain may penetrate. In the later months, however, the CAB is less frequent, and a large component of the rainfall comes from CAB breakdown events. This ensures that seasons with more CAB breakdowns are associated with more rainfall from October through to December.

Chapter 5 demonstrated that the southern edge of the African rainband is a well-defined dryline in the Austral spring. The rediscovery of the CAB provides a previously missing dynamical link between the Angola heat low and the location of the rainband edge. Interannual rainfall variability along the southward edge of the rain-band is dominated in early spring by the CAB latitude and in late spring and early summer by the frequency and timing of CAB breakdown events.

7.2.4. Chapter 6: CMIP5 models and projected future change

In Chapter 6, the future change of southern African rainfall in CMIP5 models is traced to changes in tropical lows and the CAB. Both features were shown to

be represented in the historical simulation of majority of CMIP5 models with a moderate degree of skill. The spatial distribution of tropical lows exhibited a southward shifted bias, which may be linked with the historical rainfall bias in this region (Munday and Washington, 2017). Rainfall decline in the Austral summer season was then linked to projected increases in CAB latitude and frequency. Meanwhile, disagreement in model projections of rainfall in the Austral summer was attributed partially to disagreements in projections of tropical low distribution changes.

CMIP5 projections exhibit a robust future drying trend from October to December, associated with a delay in wet season onset (Dunning et al., 2018). The magnitude of the rainfall decline varies considerably between models ranging from 0 - 34% of the historical simulated annual totals (Munday and Washington, 2019). Chapter 6 attributes this rainfall decline to an increase in CAB latitude and frequency. This result suggests that the CAB is projected to remain further north for longer and breakdown later in the season in the future, delaying the wet season by on average half a month.

The persistence of the CAB and its faithful representation in CMIP5 models throughout the Austral spring have important implications for the mechanisms of future change in this region. In daily reanalysis data and climate model output, the CAB is a discontinuity featuring a step change between regions of high and low humidity. This is in stark contrast with the monthly mean climatology of surface humidity, which smears the CAB into a smooth, weaker gradient. The dynamics of a sudden discontinuity are very different to those of a weak gradient. For example, the upped-ante mechanism of future drying assumes a gradual humidity gradient and does not apply when a dryline is present. Therefore, conceptual models of future change must change so that the CAB is taken into account. The

findings of Chapter 6 therefore emphasise the importance of studying the climate and its representation in CMIP models on daily timescales.

From December to February, in the core of the wet season, the ensemble mean rainfall continues to decline to the south of 15°S. However, this result is less robust, with different models showing different spatial patterns of change and some models exhibiting rainfall increases in localised areas. Chapter 6 demonstrated that this is in part due to a divergence in the projected distributions of tropical lows in future simulations. While many models show an overall decline and northward shift in tropical lows, the locations of the maximum decline vary between models, as do the locations that feature an increase in tropical lows. The variability of the spatial distribution of rainfall change in the tropical edge region was weakly correlated with the spatial tropical low distribution change. Variations in the future change of tropical lows therefore contribute to variations in rainfall projections.

Chapter 6 has placed future drying in southern Africa in the context of tropical lows and the CAB. Model agreement on Austral spring drying stems from agreement that the CAB is predicted to move south more slowly and persist on average half a month longer by the end of the 21st century. Conversely, model disagreement on Austral summer are associated with diverging projections of future changes in tropical low distributions. This chapter has demonstrated how meteorological context may be derived from considering the synoptic expression of future rainfall change.

7.3. Fulfilment of goals and aims

The overall goal of this thesis was to express the climatology, dynamics and future change of the southern African tropical margin from the viewpoint of

synoptic weather systems. In order to achieve this, five separate aims were set for the two synoptic weather systems under consideration, tropical lows and the CAB. These were:

- Aim 1: Develop flexible algorithms that are capable of identifying the CAB and tropical lows in daily and sub-daily reanalysis data and model output across a range of spatial resolutions;
- Aim 2: Characterise the behaviour, spatial distributions and frequencies of the CAB and tropical lows in climatological-length datasets;
- Aim 3: Study the dynamics that drive the behaviour of the CAB and tropical lows;
- Aim 4: Determine the extent to which the CAB and tropical lows deliver interannual rainfall variability signals; and
- Aim 5: Characterise future rainfall change in the context of the CAB and tropical lows.

The following sections will work through these aims and describe how they were achieved. The first section considers the goals and aims in relation to tropical lows, and the second considers them with regard to the CAB.

7.3.1. Aims: Tropical lows

In the context of tropical lows, Aims 1 - 4 were achieved across Chapters 3 and 4, while Aim 5 was the focus of Chapter 6. Chapter 3 identified the importance of tropical lows to the Angola Low and presented analysis on their characteristics, behaviour and dynamics. Chapter 4 presents a more comprehensive algorithm for the detection of tropical lows and analyses statistics relating to their tracks, rainfall contribution and interannual variability. Chapter 6 addressed the representation of tropical lows in historical and future CMIP5 model output.

In Chapter 3, a simple methodology was used to locate moist convecting deep cyclonic vortices. This involved identifying vortices with $\zeta_{800} < -3 \times 10^{-5} \text{ s}^{-1}$, and with dry static stability too high for heat lows to be present. Meanwhile in Chapter 4, a more comprehensive feature tracking algorithm was applied to identify tropical lows, which more fully addressed the requirements of Aim 1. This algorithm was based on TRACK (Hodges, 1994, 1999), a feature tracker designed for use in meteorological and oceanographic applications. TRACK has been used to study cyclonic vortices across the world, including monsoon depressions, tropical cyclones, extra-tropical storm tracks and African Easterly Waves (Hurley and Boos, 2015; Hodges et al., 2017, 2003). The algorithm was optimised for performance in southern African conditions, which involved setting the expected horizontal length scale and vertical extent of tropical lows, informed by the results of Chapter 3.

Characteristics of tropical lows, satisfying Aim 2, were described in Chapters 3 and 4. Characteristics of tropical lows that do not require systems to be traced through time, such as vertical profiles and cross sections were studied in Chapter 3, while track statistics were compiled in Chapter 4. It was found that tropical lows typically extend to about 400 hPa, feature a core of cyclonic vorticity approximately 1000 km wide, and are associated with cool temperatures near the surface and warmer temperatures aloft. They cluster preferentially in Angola, though they exist all throughout southern Africa between the latitudes of 10° - 25° S.

In order to address Aim 3, the dynamics of tropical lows were primarily studied in Chapter 3, where a unique anchoring mechanism was found to explain the preferential semi-stationary behaviour of tropical lows in Angola. This mechanism was related to the interaction of the west coast sea breeze with the tropical low

systems. The sea breeze advects inland overnight, pushing ahead of it a zone of enhanced wind convergence, which acts to intensify the cyclonic vorticity of the tropical lows through vortex stretching when they are located in Angola.

The extent to which tropical lows influence interannual rainfall variability in southern Africa - Aim 4 - was touched on in Chapter 3 and expanded in Chapter 4. Chapter 3 began with the established interannual correlation between rainfall and the strength of the Angola Low, and decomposed this correlation into heat low and tropical low components. The correlation with precipitation was found to apply only to the tropical low component, which suggested that tropical lows were in some way linked to precipitation across southern Africa. Based on this result, daily rainfall was attributed to nearby tropical low events in Chapter 4. The interannual rainfall variability of the component of rainfall that was associated with tropical lows accounted for most of the variability in northern Namibia and Botswana. Meanwhile, the southern African ENSO rainfall signal was shown to be delivered predominantly via the tropical low component of rainfall in Angola, Namibia, Botswana and Western Zambia. This suggests that tropical lows are of vital importance for interannual rainfall variability on the southern margin of the African rain band. Further analysis in the afterword of Chapter 4 indicated that this rainfall relationship is primarily due to rainfall falling in the direct vicinity of tropical lows, and not due to remote effects.

Chapter 6 evaluated the representation of tropical lows in the CMIP5 archive, in line with Aim 5. It found that tropical lows were generally shifted too far south in historical CMIP5 experiments as compared to reanalysis products, but that overall their representation was adequate. Future change of tropical lows indicated an overall decrease in frequency, and in some cases a northward shift. However, outside of these general trends, a large degree of spatial variability was

present in the shifts in the spatial distribution of tropical lows. This variability was reflected in the spatial pattern of projected rainfall changes. The spatial correlation between rainfall change and tropical low frequency change was low but positive across models. This result suggested that divergent projections of tropical low behaviour are partly responsible for divergent rainfall projections in southern Africa.

7.3.2. Aims: Congo Air Boundary

Aims 1 - 4 were all achieved for the CAB in Chapter 5. This chapter developed an algorithm to identify the CAB, studied the CAB's characteristics and dynamics, and developed a link between interannual CAB and rainfall variability. The dynamics of the CAB linked in closely with heat low analysis performed in Chapter 3. Aim 5, concerning future projections, was achieved in Chapter 6.

Addressing Aim 1, an algorithm was developed to detect the CAB in Chapter 5. This algorithm identified drylines and convergence lines associated with the CAB, along with the KD, based on near-surface specific humidity and boundary layer wind speeds using techniques from image processing. It flagged lines of grid-cells that formed sharp edges of humid air masses and ridges of wind convergence where air masses meet, which were then filtered based on location, length and angle criteria. It was shown in Chapter 6 that this algorithm was sufficiently flexible to be modified and applied across different resolutions and to climate model output.

Chapter 5 applied this the algorithm to 39 years of ERA-5 reanalysis data to study the CAB's characteristics, behaviour, spatial distribution and frequency in accordance with Aim 2. The CAB was proven to exist on most days in August and September, before dropping in frequency across October and November to

near zero in December. The CAB was found to track steadily south during this period, beginning near Angola's northern border and ending near its southern border.

A moisture budget and vertical cross-sections of key atmospheric variables such as temperature, winds and humidity revealed that the dynamics of the CAB (Aim 3) are intrinsically linked to the Angola Heat Low. The diurnal circulation associated with the heat low acts to sharpen and maintain the humidity gradient. In this manner, the dynamics of the heat low, discussed in Chapter 3, indirectly fed in to Aim 3. The seasonal southward shift of the southern edge of the tropical rain-belt therefore likely to be associated with the seasonal southward shift of the heat low, shown by Figure 3.4 in Chapter 3.

The control influenced by the CAB on precipitation (Aim 4) was statistically demonstrated in Section 5.8 by linking interannual variability of CAB properties to that of local precipitation. In regions between 8° and 16°S , 55 - 70% of interannual rainfall variability was associated with CAB latitude in early spring, and CAB frequency in late spring. It was therefore determined that the CAB variability delivers a large proportion of the local interannual rainfall variability signal to Angola and western Zambia.

Finally, future projections of spring rainfall decline in southern Africa were characterised by changes in the CAB in Chapter 6, addressing Aim 5. In this chapter, it was shown that the CAB was simulated to a sufficiently realistic standard in all but one of the CMIP5 models considered, with the exception being CanESM2. At the end of the 21st century under the RCP8.5 scenario, the CAB frequency increases significantly across models from late November through to January, indicative of a delay in the CAB breakdown. The magnitude of this frequency increase was correlated with the magnitude of rainfall decline across

models. Furthermore, the CAB marks the boundary between regions of rainfall increase and decline. Lazenby et al. (2018) proved that a projected circulation change associated with the projected rainfall decline must exist. The results of this thesis indicated that this circulation change is the delayed breakdown of the CAB.

By addressing the aims of this thesis, the CAB has been used to provide meteorological context to the future rainfall decline in southern Africa. This work has established that the CAB is an important feature of the southern African climatology, described its characteristics and clarified its dynamics. This has enabled interannual rainfall variability and future rainfall decline in Angola and western Zambia to be understood from the viewpoint of variability and change in the location and frequency of the CAB.

7.3.3. Summary

Through Aims 1 - 5, this thesis has presented a recipe for unravelling a future change signal with respect to regional meteorology. The application of this recipe to both the spring and summer seasons in southern Africa has been largely successful.

The bespoke algorithms developed and optimised through Aim 1 allowed for the spatial distributions and frequencies (Aim 2) to be easily computed; for the dynamical processes (Aim 3) to be identified; and for relationships with interannual rainfall variability (Aim 4) to be quantified. The algorithms could then be applied to climate models, where the spatial distributions, frequencies and dynamical understandings were used to evaluate model processes and derive process based projections. Relationships with interannual variability were used to precisely discern the link between process change and rainfall change. In this

manner, Aims 1 - 4 all contributed to the achievement of Aim 5. Along the way, a wealth of enhanced understanding of the basic state of the southern African climate was delivered.

The recipe presented and followed in this thesis is widely transferable and may be used to provide synoptic contexts to basic climatological states and future change signals across the world.

7.4. Discussion

The results of this thesis were presented as four chapters, each a self-contained publication. Together, these chapters have achieved the aims of this thesis. They also reached the overarching goal of expressing the climatology, variability and future change of the southern African tropical margin from the viewpoint of synoptic weather systems. However, the results of these separate chapters also include several overlapping themes that warrant individual consideration. These themes include: the underlying dynamical processes of the southern African region; the means by which the behaviour of synoptic lows and drylines shape the spatial pattern of the rainfall climatology; the implications of unpredictable behaviour of tropical lows for seasonal and future climate prediction; and the way that the rediscovery of the CAB fits in with wider literature regarding the organisation of convection in the tropics. The following discussion will draw further links between the constituent chapters of this thesis, by describing how the chapters addressed these themes.

7.4.1. Dynamics of tropical lows, heat lows, the CAB and the KD

A common theme across this thesis has been the role of low-level wind convergence in shaping the behaviour of synoptic weather systems in southern Africa. These winds are fundamental to the dynamics of heat lows, the maintenance of

the CAB and KD, and the anchoring mechanism of tropical lows. The convergent winds flow down pressure gradients that are set up by differential surface heating. This is made possible by the low latitude of the study area, which ensures that the Coriolis force is weak, and so geostrophic balance takes approximately half a day to take effect. There are two distinct temperature gradients driving divergent winds. The first is the north-south gradient between the humid and cool Congo Basin and the hot and dry interior of southern Africa, which is strongest in spring and curves the LLW wind southwards. The second is the land-sea contrast between the South Atlantic Ocean and the southern African interior, which sets up the west coast sea breeze. A third temperature gradient exists between the Indian Ocean and southern Africa. However, the winds that flow across this temperature gradient winds are largely in geostrophic balance with the Mascarene High. Therefore, these winds cannot be said to be driven by the land sea temperature gradient, though this gradient does accelerate these winds overnight. This acceleration was apparent in the diurnal vertical cross-sections shown in Figures 3.5, 3.18 and 5.6.

The converging winds, heat lows, drylines and convergence lines are shown in Figure 7.1. The location of the heat lows in this schematic are derived from Figure 3.4. The combined dryline / convergence line between the LLW and the easterly trade winds in spring is the CAB. Meanwhile, the convergence line between the west coast sea breeze and the easterly trade winds (which is also a dryline in summer) is the KD.

Chapter 3 demonstrated that the convergence between the easterly trade winds and the west coast sea breeze is fundamental to the dynamics of the Angola Heat Low. However, Chapter 5 indicated that the Angola Heat Low was associated with the CAB, and hence the convergence between the LLW and the

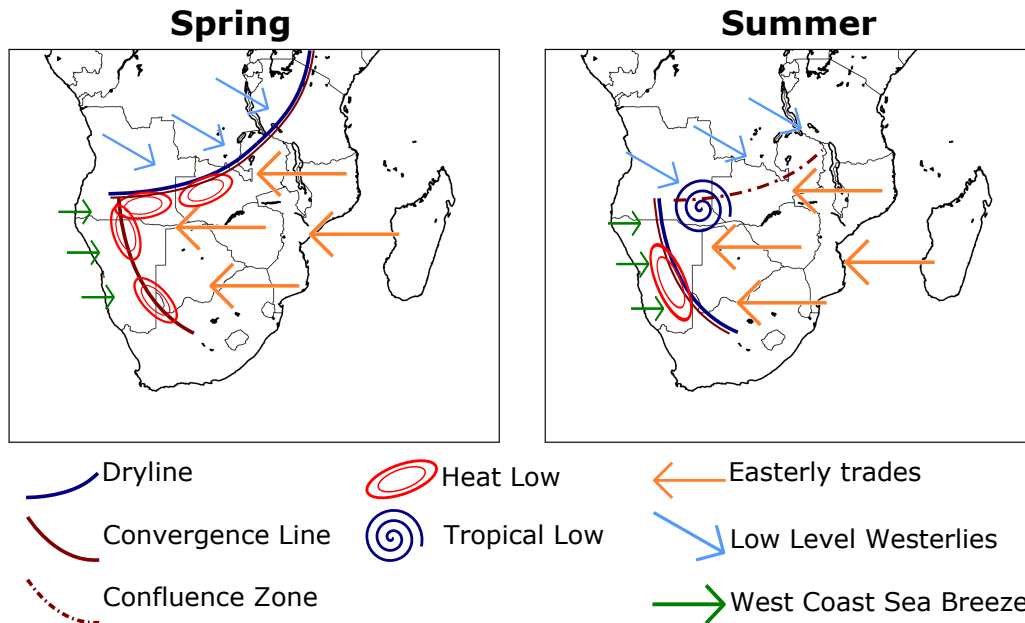


Figure 7.1. Schematic of converging winds, convergence lines and zones, dry-lines, heat lows, and tropical lows.

easterly trade winds. As indicated in Figure 7.1, the reality is that both are true. The Angola Heat Low is associated with the triple-point of the convergence of all three winds. This is apparent from the inflow patterns from the north-south composites in Figure 3.18 included in the afterword of Chapter 3. The north-south cross sections show low level meridional wind convergence into the centre of the heat low, matching the zonal wind convergence shown in the east-west cross sections (Figure 3.5). This is represented by the schematic in Figure 7.1 by two overlapping heat lows in spring, one parallel to the CAB and the other parallel to the KD.

The KD is the result of the convergence of the west coast sea breeze with the easterly Indian Ocean trade winds. In early spring, the KD is only a convergence line, but it becomes a dryline in late spring and early summer. The Kalahari Heat

Low is present in association with the KD as early as October (Figure 3.4) but strengthens when the CAB breaks down and the Angola Heat Low disappears. The upper edge of the Kalahari Heat Low and the KD remains near the Angola-Namibia border (15°E , 18°S). However, the centre of the heat low shifts to the Kalahari. The KD, the Kalahari Heat Low and the west coast sea breeze are all governed by the same dynamics, and may be regarded as aspects of the same atmospheric feature.

When the CAB has broken down, moist unstable air spreads though Angola and Zambia, originating from both the Congo and the Indian Ocean. The meeting point of the Indian Ocean trade winds and the LLW winds is no longer a well defined convergence line, but more of a confluence zone in which tropical lows form. The KD forms a western boundary to the confluence zone. In this region, the west coast sea breeze interacts through vortex stretching with tropical lows in Angola through the anchoring mechanism described by Chapter 3. Since we have established that the sea breeze and the KD are different aspects of the same feature, this anchoring mechanism may be viewed as an interaction between tropical lows and the KD. The anchoring can be seen in action in the sample tropical low studied in Chapter 4 (Figure 4.1), where an elongated region of low-level cyclonic vorticity along the west coast of southern Africa begins to merge with a tropical low in Angola during the final two days of the sample period.

The above exposition implies that the dynamics of the synoptic weather systems studied in this thesis are all either driven or affected by surface wind convergence associated with surface heating. In a way, this is not surprising. This thesis has focussed on the dynamics of the poleward excursion of the tropical rainband during the Austral spring and summer over the southern African subcontinent. Over the tropical oceans, the rainband is confined to latitudes between 15°N and

15°S, and it is only over land that the rainband is pulled poleward, through monsoon circulations and the dynamics of other summer wet seasons. The driver of these monsoons is widely known to be differential land-sea heating. The southern African wet season is not a monsoon, as it lacks the requisite reversal of winds. Nevertheless, differential heating is the main driver of the wet season, and the generation of converging low level jets is the key means by which it is expressed in the momentum budget.

7.4.2. The significance of synoptic timescales in southern Africa

This thesis has considered the southern African climatology as an aggregation of weather systems that operate on daily and sub-daily timescales. In this section, the value and implications of this synoptic approach are discussed. First, the motivations for this approach are revisited. Second, the climatological consequences of the synoptic dynamics explored in this thesis are presented. This includes the roles of the tropical low anchoring mechanism and the gradual southward progression of the CAB in shaping the rainfall climatology, and implications of tropical low predictability for seasonal forecasts.

Mesoscale and synoptic weather systems evolve on daily and sub-daily time frames. Since weather systems form the building blocks of the climatology, these are the most natural timescales at which to study atmospheric dynamics. Synoptic and mesoscale events are key to the processes that drive the climatological distributions of rainfall. In order to study the atmosphere at longer seasonal-scales, averages must be taken that will inevitably smear different weather systems together. This suggests that seasonal means are not capable of exposing rainfall mechanisms unless averages are performed using carefully constructed composites

that take the spatial features of the weather systems into account. For these reasons, this thesis has taken the approach of identifying and tracking key weather systems, the CAB and tropical lows, on daily timescales. This approach has allowed for the seasonal cycles of these weather systems to be considered. This was achieved through analysis of spatial distributions, the construction of centred composites and the attribution daily rainfall to nearly storm tracks. The following sections consider the implications that this analysis has revealed for the rainfall climatology and for seasonal forecasting

7.4.2.1. *Southern African climatological rainfall distribution*

The sub-daily approach taken by this thesis has revealed important dynamical interactions at play in southern Africa that determine key features of the climatology. It was found that the interaction of the KD/west coast sea breeze with tropical lows caused these weather systems to linger preferentially in eastern Angola. This anchoring explains the climatological local maximum of January precipitation located in central Angola. The aversion of tropical lows to the coastal region also explains the sharp rainfall drop-off in southern Angola between 20°E and 15°E, where coastal deserts dominate the landscape. Indeed, the incursion of cool, low specific humidity air associated with the KD/west coast sea breeze into Namibia each night may well be important for explaining the hyper-aridity of the Namib and Kalahari Deserts.

In a similar vein, the inter-linkage between the Angola Heat Low and the CAB drives the slow southward progression of the tropical rain-band through the Austral spring. In the interior of southern Africa, monthly rainfall climatologies from August through to November show evenly spaced parallel isohyets of constant rainfall along the southern tropical margin. Comparison with Chapter 5 reveals

that these isohyets match the location of the CAB in each month. Dunning et al. (2018) found that the southward march of the tropical rain-band over the southern DRC, Angola and Zambia was the slowest lateral progression of onset in the continent. Again, the iso-lines of constant climatological onset date are parallel to the orientation of the CAB.

It is generally well understood that the progression of the tropical rain-band follows the latitude of maximum solar insolation. However, this qualitative view is insufficient when attempting to understand the dynamics of future change. The precise dynamical mechanism by which the rain-band moves poleward in spring needed to be unravelled before the dynamics of the projected delayed onset could be resolved. However, this mechanism had not been fully addressed in southern Africa prior to the publication of Chapter 5, which described the relationship between the CAB and the Angola Heat Low. The location of the heat low is linked to the maximum surface temperature, and so moves south through the season, bringing the CAB with it. Identifying the CAB in daily reanalysis data and analysing its diurnal cycle in sub-daily data was crucial for establishing the means for its dynamical evolution and hence the onset of the rainy season.

7.4.2.2. *Tropical lows and predictability*

This thesis has illustrated the influence of tropical lows on climate uncertainty in southern Africa. This has been examined both on interannual and climatological timescales, and has implications for both seasonal forecasting and future climate projections.

On interannual timescales, ENSO is the leading order mode of rainfall variability. However, the response of southern African rainfall to ENSO can be hard

to predict. Some seasons with high magnitude ENSO events have not been associated with commensurate precipitation anomalies, and some seasons exhibiting severe droughts have not been associated with ENSO. The canonical example of this is the 1997 - 98 El Niño, which was one of the three strongest El Niños of the last 40 years but did not result in drought in southern Africa. Studies focussing on this season led to the consensus that the Angola Low modulates the ENSO signal (Reason and Jagadheesha, 2005; Lyon and Mason, 2007), but no mechanism for this modulation has been proposed until Chapter 4.

By studying tropical lows on a synoptic timescale, Chapter 4 showed that the Angola Low influences rainfall via its aggregate tropical low systems, which generate large-scale environments favourable for convection. Rainfall deriving from these systems makes up a 31% of rainfall in the tropical edge (16° - 22° S) and dominates the high rainfall variability in Botswana and northern Namibia. In particular, the majority of the ENSO signal to the west of 28° E derives from tropical lows. However, tropical lows and ENSO are only weakly correlated. Chapter 4 found that only 17% of the variance of tropical low frequency and 23% of the variance of tropical low mean latitude is explained by the Niño 3.4 SST index. This leaves a large proportion of interannual tropical low variability unexplained by ENSO.

A similar result was found regarding tropical lows in future climate models studied in Chapter 6. While an overall reduction and northward shift was evident across models, there was a large degree of variability in predictions of small scale changes in the spatial distribution of tropical lows. Much of this variability derived from differences in the locations of the tropical low frequency peaks in historical simulations. Spatial patterns of rainfall changes then mapped onto these small-scale tropical low distribution changes. This was evident from the

low magnitude positive spatial correlations between future change in tropical low distributions and future rainfall change across all models but one. Therefore, slight differences in the spatial distributions of tropical lows across historical simulations influences the location of tropical low decline in future models, which in turn controls the location of future rainfall decline. In this way, the difficulty that climate models present in simulating tropical lows translates into increased model divergence regarding the location of projected rainfall decline.

Tropical lows are weather systems with non-trivial predictability that exacerbate the uncertainty in seasonal prediction of southern African rainfall. Their importance for wet season rainfall in southern Africa poses a challenge for climate models, as small deviations in tropical low simulation in different models lead to larger divergences in the spatial pattern of rainfall change. In this manner, this thesis has traced a component of precipitation uncertainty in the southern African climatology to this specific class of weather systems. A better understanding of the dynamics driving these systems is necessary to reduce uncertainties surrounding interannual variability and future change of southern African rainfall from December through to February.

7.4.3. The dichotomy of convective and subsidence regions

The persistent presence of the CAB on a daily timescale during spring reveals a subtle but important point regarding the mean state of the atmosphere in southern Africa at this time of year. In the seasonal mean, the tropical margin is a smooth surface humidity gradient varying from moist to very dry over distances of order 1000 km, passing through a spectrum of intermediate humidities along the way. At a daily timescale, however, the tropical margin is characterised by a sharp discontinuity between two airmasses. The wide gradient in the synoptic

mean derives from the fact that the CAB is in a slightly different position for each day aggregated by the average, owing to its southward progression. Surface humidity in the margin of the tropical rain-belt exhibits dichotomy of two distinct states, as opposed to the smooth spectrum implied by the seasonal mean. These states are the hot, dry air associated with the easterly trade winds to the south of the CAB, and the cooler, near saturated Congo airmass to the north.

This result has several implications. Firstly, the Reynolds decomposition is not appropriate in this setting. Reynolds decompositions are a useful tool used by climate scientists to conceptualise atmospheric states. However when a moving CAB is consistently present, subtracting the mean state from an instantaneous will not produce an isentropic eddy field. Instead, this would retain a sawtooth pattern due to a mismatch between the mean and instantaneous fields. The results regarding the CAB reveal that different conceptual models that are capable of accounting for the spatial discontinuity must be used in the context of southern Africa, particularly during spring.

As discussed in Chapter 6, the existence of two distinct atmospheric states with a sharp boundary between them has implications for the mechanisms of future change. In such a scenario, small thermodynamic perturbations have less potential to change for the likelihood of convection than small dynamical perturbations. This is because a dynamical perturbation may move the moist airmass into, or away from a given location, drastically changing the likelihood of convection at that spot. Meanwhile, a small change in saturation temperature is unlikely to change the likelihood of convection if the atmosphere is either severely dry or not at all moisture limited. This means that the upped-ante mechanism of future change, which is associated in thermodynamic reductions of relative humidity in a warming atmosphere with increasing moisture capacity, will be overwhelmed

by circulation changes. This reflects the findings of Chadwick et al. (2013) and Lazenby et al. (2018), who demonstrated that seasonal mean rainfall projections across the tropics are dominated by shifts in atmospheric circulation patterns, rather than by uniform rich-get-richer or upped-ante mechanisms. The projected northward shift of the CAB is a clear example of such a circulation shift, whose existence is predicted by Chadwick et al. (2013) and Lazenby et al. (2018).

Secondly, the dichotomy of states is indicative of a broader truth regarding the fundamental underpinnings of the tropical atmosphere. Similar patterns of spatial organisation into distinct regions of high and low humidity have been found in numerous regions and parameter spaces, including over the tropical oceans and in cloud resolving models (CRM). Mapes et al. (2018) demonstrated that the distribution of column water vapour (CWV) over the ocean is bimodal in structure, with a moist mode maintained by convective and dynamical processes. They identified the meandering margin of this moist airmass using the 48 mm contour of CWV, and observed that the margin was approximately 100 km wide, comparable with the CAB.

CRMs are a class of idealised simulations that focus on simulating cloud-scale motion and often simplify atmospheric dynamics. Most CRMs do not feature rotation and assume a fixed (oceanic) surface temperature, along with periodic lateral boundary conditions. Tompkins (2001) first observed that under certain conditions, convection in CRMs self-organises into distinct convective regions which feature uplift, high humidity and low OLR, and subsidence regions which feature low humidity, clear skies and high levels of radiative cooling. Since then, a large body of literature has evolved, examining the conditions necessary and sufficient to achieve self-organisation, the associated radiative-convective feedback processes and the spatial structure of the organised cells (e.g. Bretherton et al.

(2005); Muller and Bony (2015); Wing et al. (2017)). The edges of these convective regions resemble the CAB and the tropical margins in their characteristics. Theoretical studies underpinning the concept of convective organisation suggest that under the framework of radiative convective equilibrium (RCE), which considers steady state achieved by an atmosphere devoid of lateral energy transport, a bifurcated solution with multiple equilibria may exist in some circumstances (Emanuel et al., 2014). These equilibria resemble the two atmospheric states separated by the CAB - one moist and convective, and the other dry and dominated by subsidence.

The rediscovery of the CAB therefore adds to a growing body of literature regarding the contrast between the tropics and the subtropics, and the existence of a sharp, well defined boundary between them. The existence of this boundary implies the existence of constraints on the atmospheric state which control spatial distribution of water vapour. These constraints are fundamental in governing the behaviour of the mean state of the tropics, and so understanding them is crucial when predicting the response of the tropics to climate change.

7.5. Concluding remarks

7.5.1. Avenues for future research

This thesis has characterised the representation of two synoptic weather systems, tropical lows and the CAB, in reanalysis datasets and climate model output. It has found that the dynamics of these systems are fundamental for the climatological distribution and projected future change of precipitation in southern Africa. However, nearly all of the characterisation of the basic atmospheric state in this thesis has been founded on reanalysis datasets. In southern Africa, a dearth

of in-situ upper air observations ensures that reanalysis products are heavily reliant on their constituent forecast models, rather than assimilated observations. This means that very little of this thesis is based on direct observations of the southern African atmosphere. The exception is Section 5.6, which verified the existence of the CAB in automatic weather station data.

Given the influence that the CAB and tropical lows have been shown to hold over southern African rainfall, direct measurements of these weather systems would be invaluable. The seasonal cycles and spatial distributions of the CAB and tropical lows presented in this thesis may be used to inform observational campaigns, with aims towards measuring vertical profiles of either system. The reliable southward shift of the CAB through Western Zambia in the Austral spring seems particularly well-suited to a targeted field campaign.

This thesis has also demonstrated that tropical lows act as a wild-card in southern Africa, responding to modes of variability independent of ENSO. This poorly understood variability has flow-on effects that reduce the certainty of seasonal forecasts and future projections of southern African rainfall. Further research into these modes of interannual variability from a viewpoint of synoptic timescale dynamic, is strongly advised. For example, a focus on the interannual variability of the proportion of tropical lows which are anchored against the proportion that are more transient may reveal teleconnections that increase the predictability of tropical lows on interannual timescales.

7.5.2. Contributions of this thesis

This thesis has advanced the scientific understanding of the basic state of southern African regional climate system. It has characterised the climatology, behaviour and variability of two prevalent classes of weather systems: tropical

lows and CAB. This work has clarified the roles of the CAB and tropical lows in delivering rainfall to southern Africa and demystified their dynamics through interactions with surface wind convergence. This allows the simulation of these processes and dynamics to be evaluated in climate models, reducing preventable uncertainty associated with seasonal forecasts and climate projections.

This thesis has provided the meteorological context to projections of rainfall change in southern Africa. It has applied process based model evaluation techniques in Chapter 6 to determine that the CAB and tropical lows are well represented in the CMIP5 dataset, though tropical lows possess a southward shifted bias. The future spring drying projections have been linked to a projected increased frequency of the CAB, and it has been shown that simulation of tropical lows is instrumental in spatial disagreements regarding the direction of summer rainfall change. More broadly, this thesis has demonstrated the value of using feature tracking techniques to study ‘process based projections’ of the climatologies of synoptic weather systems under warming scenarios. It has shown that this methodology can be used to identify mechanisms of rainfall change associated with circulation changes. The identification of these mechanisms has led to a reduction of the uncertainty of drying trends over southern Africa. In turn, this allows for increased confidence in the application of these projections to guide adaptation.

An additional finding of this work was the persistence of sharp drylines, and not gradual humidity gradients, along the southern edge of the African rain-belt. This result challenges preconceptions regarding the state of the southern African atmosphere. Conceptual models, upon which scientific hypotheses are developed, must take the existence of the CAB into account, instead of being centred around

a smoother mean state. This result echoes growing awareness of the importance of convective organisation across other regions of the tropics.

In conclusion, this thesis has studied the regional climate dynamics of southern Africa at a synoptic timescale, and developed a deeper understanding of the regional synoptic processes at play along the southern tropical margin. It has characterised the climatology, dynamics, variability and future change of the tropical margin of southern Africa from the viewpoint of tropical lows and the CAB. Surface wind convergence has been shown to be fundamental to the dynamics of both weather systems. The interannual variability of tropical lows and the CAB have explanatory power over interannual variability of southern African precipitation. A bespoke approach to future projections has expressed future rainfall changes in terms of process based projections of tropical lows and the CAB. Based on this analysis, a circulation based mechanism of spring rainfall drying and an explanation for inter-model diversity of summer rainfall projections have been provided. This work has been based on the application of objective feature tracking algorithms optimised to tropical lows and the CAB.

Synoptic feature tracking has linked the meteorology of the southern African tropical margin to its climate. This has led to an enhanced understanding of a chronically understudied rainfall climate. Tropical lows and the CAB modulate rainfall that underpins the lives of hundreds of millions in southern Africa. Crucially, process-based projections of these features are necessary to reduce the climate uncertainty hanging over the future development and growth of the region. This thesis has provided the meteorological context of the future drying trend and has constituted a vital step towards constraining climate uncertainty in southern Africa.

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